DRAINAGE PRINCIPLES AND APPLICATIONS

I INTRODUCTORY SUBJECTS

II THEORIES OF FIELD DRAINAGE AND WATERSHED RUNOFF

III SURVEYS AND INVESTIGATIONS

IV DESIGN AND MANAGEMENT OF DRAINAGE SYSTEMS

Edited from:
Lecture notes of the
International Course on Land Drainage
Wageningen

deel 1

INTERNATIONAL INSTITUTE FOR LAND RECLAMATION AND IMPROVEMENT
P.O. BOX 45 WAGENINGEN THE NETHERLANDS

PREFACE

The International Institute for Land Reclamation and Improvement was founded in Wageningen in 1956 and charged with the task of collecting and disseminating knowledge in the fields mentioned in its name.
During its first few years of existence, the Institute received a steadily increasing stream of visitors from abroad, who asked a wide variety of questions about matters of agro-hydrology in which the Dutch are known to have considerable experience: problems of waterlogging and the process of draining off excess water. It soon became clear that the guidance and training of these individual visitors took up a disproportionate amount of the Institute's time, and this fact forced us to consider how – aside from our publication programme – we could best satisfy the vigorous interest that was being shown.
And so the idea was born to organize a course that could systematically handle the subject of 'land drainage', and the basic knowledge relevant to it. Initial plans were drawn up in 1960. A Board, consisting of representatives from allied Dutch institutions, was appointed to supervise the scientific and practical programming. Prof. Dr. F. Hellinga served as the first Chairman of this Board.
To handle the administrative, financial, and social matters connected with the course, coöperation was sought – and obtained – from the International Agricultural Centre at Wageningen. In 1962 the first 'International Course on Land Drainage' was launched. Its language was English; it lasted three months; its participants numbered twenty-five.
What was originally regarded as an incidental event – one that might be repeated at some future date if need be – proved a 'hit' that demanded repetition, and the course became an annual event. The tenth course in 1971 brought the total number of partici-

pants to 281, who came from a total of 62 different countries.
The Institute is grateful for the vast measure of coöperation it has always received from other Dutch institutions, which made their research and field experts available to lecture in the course, along with the Institute's own team of lecturers.

From the outset, participants were provided with lecture notes to lend support to the spoken word. Many non-participants, however, were also interested in obtaining these notes, but we were unable to comply with requests for their supply because we felt that, in general, the text was not sufficiently 'balanced', not adequately 'crystallized', to be read independent of the lectures. Editing was often crude, although most texts have improved over the years. With the gradual refinement in the balance of the subject-matter – tested against the needs of our students – and the ever-increasing pressure to make the notes available to a wider public, the Board of the Course decided in 1969 to have the entire lecture notes re-edited, and then to have them issued by the Institute in a simple four-volume publication.
An Editorial Committee consisting of members of the Institute's staff was set up to undertake the work. The Committee comprised:

Mr. P. J. Dieleman, Chairman (1969–71)
Mr. J. G. van Alphen (1969)
Mr. G. P. Kruseman (1969–70)
Mr. R. J. Oosterbaan (1970–71)
Mr. S. J. de Raad (1970–71)

By the middle of 1971, after two years of hard work, the Committee unfortunately broke up as, one by one, its members left for assignments in other parts of the world. During the last half of 1971, only one staff member, Mr. J. H. M. Aalders, continued the work of preparing the manuscript for publication. After his temporary appointment came to an end, a Working Group of other staff was formed, whose aim was to finish the job within the framework designed by the original Editorial Committee. The members of this group were:

Mr. J. Kessler, Chairman,
Mr. T. Beekman,
Mr. M. G. Bos,
Mr. R. H. Messemaeckers van de Graaff,
Mr. N. A. de Ridder,
Mr. J. Stransky
Mr. Ch. A. P. Takes,
Mrs. M. F. L. Wiersma-Roche.

Having served as Director of the Institute during the period when the International Course on Land Drainage came into being and when the decision to publish the lecture notes was made, I would like to express the satisfaction I feel with the issue of the first volume in the series. Over the last three years, a large proportion of the people employed at the Institute have given much of their time and energy, even their leisure hours, to completing this work. I want to thank everyone involved, and I include not only the authors, the lecturers and the staff members already mentioned, but others too who worked so splendidly on the drawings, layout, and the production. It is my fervent hope that their communal effort will truly help in the proper implementation of land drainage throughout the world.

Agadir (Morocco)
May, 1972

J. M. van Staveren
Director (1956–1971)
International Institute for
Land Reclamation and
Improvement.

INTRODUCTION

Land drainage is the removal, by artificial means, of excess water from the soil or from the land surface, its objective being to make the land more suitable for use by man. In agriculture, its aims are to increase production, to sustain yields, or to reduce production costs – all helping the farming enterprise to maximize its net profit. As such, land drainage is an age-old practice.

In The Netherlands, with much of its flat land lying below the water level of the sea or that of the rivers, drainage has always been a vital necessity. It developed from the building of simple sluices in natural channels through which the excess water could be discharged by gravity when the sea or river levels were low, into the present-day sophisticated system of parallel pipe drains, collector drains, main drains, and pumping stations. This development was paralleled by an increased understanding of the principles of drainage, upgrading it from a practice based on experience and skill into a science based on the complex interrelations between the hydrological, pedological, and agronomical conditions.

In the nineteenth century the French hydrologists, Darcy and Dupuit, were the first to formulate the basic equations for groundwater flow through porous media and to apply them to flow to wells. At the beginning of the twentieth century, Rothe applied these equations to groundwater flow to drains, and he was to derive the first drainage formula. But it was Hooghoudt who, in the thirties, gave the real stimulus to a rational analysis of the drainage problem, by studying it in the context of the plant-soil-water system. Since then, great contributions towards a further refinement of this rational analysis have been made by scientists all over the world: Childs in England, Donnan, Luthin, and Kirkham in the United States, and Ernst and Wesseling in The Netherlands.

But when drainage theories are applied in practice, we still face a number of limitations. These limitations are a consequence of the wide variability we encounter in nature when dealing with soils and plants. We are faced with such questions as: how to characterize a soil profile consisting of a large number of different layers changing in position and magnitude from one place to another; how to measure the physical soil 'constants'; how to formulate the agronomical requirements in respect to 'excess' water?
All these factors contribute to an inevitable inaccuracy which we have to accept when working in land drainage. Therefore, the statement made by Clyde Houston in 1961 in still valid:
'Although excellent progress has been made in recent years in developing drainage criteria and investigational tools, it still takes good judgement, local experience, and trial and error – along with a thorough understanding of the basic principles – to design a successful drainage system.'

In the International Course on Land Drainage an effort is being made to cover, as completely as possible and within a period of three months, the underlying principles and the application of the rational approach to land drainage.
About 30 lecturers of various disciplines each year contribute their specialized knowledge and experience to the course. Even so, not all aspects that may have a bearing on successful drainage can be fully discussed or even mentioned within the time limit set by a three-month course. A choice has to be made and explicit emphasis is therefore given to the agrohydrological aspects, while deliberately less attention is given to the hydraulics of open water flow and to engineering aspects which are more extensively treated in handbooks than are the agrohydrological aspects.
The material presented in the four volumes of this publication is based on the lecture notes prepared by the lecturers of the Drainage Course. In many instances a subject has been presented by more than one lecturer during the ten years that the course has been held. As each lecturer contributed his knowledge of the subject, each chapter must be considered the result of their combined input. For this reason a list of their names is given with each chapter, apart from that of the actual author(s).
For practical reasons, it was decided not to publish all the material in one large volume, but to make a logical subdivision into four volumes. The subjects have been grouped in such a way that each volume can be consulted independent of the others.
Volume I describes the basic elements, physical laws, and concepts of the plant-soil-water system in which the processes of land drainage take place.
Volume II presents the drainage theories and mathematical models for groundwater flow and watershed runoff, and formulates the objectives of drainage for salinity control and the prevention of waterlogging.
Volume III discusses the various surveys and investigation techniques to determine the parameters of the plant-soil-water system which are to be introduced in the drainage design computations.

Volume IV deals with the design and dimensioning of drainage systems, some of the main engineering features, and aspects of operation and maintenance.
The reader will note that the basic principles of the subject have received the main emphasis in this publication. Although due attention has also been given to the application of these principles, no ready-made solutions could be presented that would fit all the different conditions under which drainage is applied. A thorough understanding of the principles, however, should enable the reader to introduce the modifications and special techniques adapted to the special conditions he is dealing with.
We hope that the edited lecture notes of the International Course on Land Drainage, as presented now in these four Volumes of 'Drainage Principles and Applications', will find their way all over the world. Not only to our former participants and to those who will join the course in the future, but also to all the others who are dealing actively with practical or theoretical aspects of land drainage. Although a number of deficiencies, inherent to the fact that the publication consists of edited lecture notes written by many authors, may become apparent, we trust that the book will prove its usefulness. Any criticism and suggestions which might lead to improved future editions of this book will be welcomed.

The editors

LIST OF SUBJECTS AND AUTHORS OF VOLUMES I–IV

Volume I	INTRODUCTORY SUBJECTS	
Chapters		
1	Hydrogeology of different types of plains	N. A. DE RIDDER
2	Soils and soil properties	W. F. J. VAN BEERS
3	Salty soils	B. VERHOEVEN
4	Plant growth in relation to drainage	G. A. W. VAN DE GOOR
5	Physics of soil moisture	P. H. GROENEVELT J. W. KIJNE
6	Elementary groundwater hydraulics	P. J. DIELEMAN N. A. DE RIDDER
7	Electrical models: conductive sheet analogues	S. A. DE BOER W. H. VAN DER MOLEN
Volume II	THEORIES OF FIELD DRAINAGE AND WATERSHED RUNOFF	
Chapters		
8	Subsurface flow into drains	J. WESSELING
9	Salt balance and leaching requirement	W. H. VAN DER MOLEN
10	Effects of irrigation on drainage	J. NUGTEREN
11	Field drainage criteria	J. KESSLER
12	Flow to wells	J. WESSELING
13	Seepage	J. WESSELING
14	Drainage by means of pumping from wells	N. A. DE RIDDER
15	Rainfall-runoff relations and computational models	D. A. KRAIJENHOFF VAN DE LEUR
16	Hydrograph analysis for areas with mainly groundwater runoff	J. W. DE ZEEUW

Volume III SURVEYS AND INVESTIGATIONS

Chapters		
17	Surveys and their sequence	J. M. VAN STAVEREN
18	Analysing rainfall data	J. KESSLER S. J. DE RAAD
19	Determining evapotranspiration	J. W. KIJNE
20	Hydropedological survey	K. VAN DER MEER
21	Groundwater survey	N. A. DE RIDDER
22	Assessing groundwater balances	J. KESSLER N. A. DE RIDDER
23	Measuring soil moisture	W. P. STAKMAN
24	Determining hydraulic conductivity of soils	J. KESSLER R. J. OOSTERBAAN
25	Deriving aquifer characteristics from pumping tests	J. WESSELING G. P. KRUSEMAN
26	Deriving soil hydrological constants from field drainage tests	P. J. DIELEMAN

Volume IV DESIGN AND MANAGEMENT OF DRAINAGE SYSTEMS

Chapters		
27	Subsurface field drainage systems	J. C. CAVELAARS
28	Surface field drainage systems	S. RAADSMA F. E. SCHULZE
29	Main drainage systems	J. A. VAN DORT G. HEKKET M. G. BOS
30	Mechanical maintenance of ditches	H. M. ELEMA
31	Maintenance of ditches with chemicals	H. G. VAN DER WEIJ
32	Drainage of certain specific soils	W. A. SEGEREN H. SMITS
33	Drainage of rice fields	G. A .W. VAN DE GOOR
34	Management of drainage projects	F. HELLINGA J. M. VAN STAVEREN
35	Economic evaluation of water management projects	F. P. JANSEN

MAIN TABLE OF CONTENTS

v Preface
VIII Introduction
XI List of subjects and authors of volumes I –IV

1 1 HYDROGEOLOGY OF DIFFERENT TYPES OF PLAINS
3 1.1 Hydrogeology and drainage
3 1.2 Classification of water-bearing layers
6 1.3 Stream-formed plains
17 1.4 Coastal plains
21 1.5 Lacustrine plains
23 1.6 Glacial plains
26 1.7 Loess plains
27 1.8 Groundwater conditions in areas affected by dissolution
28 1.9 Influence of faults

33 2 SOILS AND SOIL PROPERTIES
36 2.1 General
44 2.2 Basic soil characteristics
54 2.3 Physical properties of mineral soils
63 2.4 Soil moisture
65 2.5 Soil air
68 2.6 Soil temperature
69 2.7 Soil fertility and productivity

75 3 SALTY SOILS
77 3.1 Origin and occurrence
78 3.2 Salt types and their distribution
79 3.3 Effects of salinity on crops and soils
80 3.4 Classification
84 3.5 Reclamation

91 4 PLANT GROWTH IN RELATION TO DRAINAGE
93 4.1 Drainage and agriculture
94 4.2 Plant and soil-water relations
100 4.3 Drainage and physical soil conditions
103 4.4 Drainage and cultivation practices
105 4.5 Drainage and nutrient supply
109 4.6 Drainage and salinity
111 4.7 Drainage and diseases or pests
112 4.8 Groundwater level and crop production

123 5 PHYSICS OF SOIL MOISTURE
125 5.1 The presence of water in the soil
125 5.2 The retention of water by the soil
131 5.3 Forces and potentials
140 5.4 Transport of water in the soil

153 6 ELEMENTARY GROUNDWATER HYDRAULICS
156 6.1 Groundwater and water table defined
156 6.2 Physical properties, basic laws
164 6.3 Darcy's law
169 6.4 Some applications of Darcy's law
174 6.5 Basic equations of groundwater flow
187 6.6 Some aspects of two-dimensional flow
195 6.7 Boundary conditions

201 7 ELECTRICAL MODELS: CONDUCTIVE SHEET ANALOGUES
203 7.1 Analogues
204 7.2 Analogy between the flow of groundwater and electricity
208 7.3 Two-dimensional models using Teledeltospaper
211 7.4 Streamlines, equipotential lines; boundary conditions
213 7.5 Simulation of boundary conditions in electrical models
217 7.6 Measurements in the model
218 7.7 Example: determination of the free water surface in a dam

223 List of principal symbols

227 Subject index

INTRODUCTORY SUBJECTS

1. HYDROGEOLOGY OF DIFFERENT TYPES OF PLAINS

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Lecturers in the Course on Land Drainage

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PURPOSE AND SCOPE

A brief account of the principal geo-morphological and geogenetical characteristics of various types of plains and their related groundwater conditions.

CONTENTS

		page
1.1	Hydrogeology and drainage	3
1.2	Classification of water-bearing layers	3
1.3	Stream-formed plains	6
1.3.1	Valleys and flood plains	7
1.3.2	Alluvial fans	11
1.3.3	Deltas	15
1.4	Coastal plains	16
1.5	Lacustrine plains	21
1.6	Glacial plains	23
1.7	Loess plains	26
1.8	Groundwater conditions in areas affected by dissolution	27
1.9	Influence of faults	28
Literature		31

1.1 HYDROGEOLOGY AND DRAINAGE

An area's drainage problems are closely related to its geomorphological and geogenetical conditions. The presence or absence of layers with good water-transmitting properties, of barriers to groundwater flow, of springs, as well as the relation between groundwater and surface water (either fresh or salty), will directly or indirectly affect the groundwater conditions in or near the rootzone. The groundwater conditions of geomorphologically (and climatologically) similar regions are often comparable. Somewhat oversimplified, it can be stated that once the type of landscape is known the principal groundwater conditions of that landscape are known too.

The present discussion will be restricted to flat areas, because it is in such areas that agriculture is preferably practised. Flat areas, if large enough, are called plains. They may have been formed by such different landforming agents as waves, running water, ice, and wind. Each agent leaves its mark by typical geomorphological features and typical internal sedimentary structures, causing more or less typical groundwater conditions.
Such typical features, structures, and groundwater conditions will be dealt with in more detail below, but first the water-bearing layers will be classified according to their water-transmitting characteristics.

1.2 CLASSIFICATION OF WATER-BEARING LAYERS

All plains referred to in this chapter are made up of unconsolidated or weakly consolidated sediments, laid down in horizontal or simply structured, well or poorly defined layers. A common feature of these layers is that they are thin with respect to their horizontal extension.
For hydrogeological purposes the layers are classified as:
- pervious
- semi-pervious
- impervious

A layer is said to be pervious if its water-transmitting properties are favourable or, at least, favourable in comparison with those of overlying or underlying strata. The resistance to vertical flow within such a layer is small and may generally be neglected, so that only those energy losses caused by

horizontal flow need be taken into account.
A layer is considered semi-pervious if its water-transmitting properties are relatively unfavourable. The horizontal flow rate over a significant distance is negligible, but vertical flow cannot be neglected because the hydraulic resistance to such a flow is small due to the relatively small thickness of the layers. The flow of water in semi-pervious layers will therefore be considered essentially vertical.
A layer is considered impervious if its water-transmitting properties are so unfavourable that only negligible amounts of water flow through it - whether vertically or horizontally. Completely impervious layers seldom occur near the surface but are common at greater depths, where compaction, cementation, and other consolidating processes have taken effect. The above classification is one of comparison, but the scale of the flow pattern must also be taken into account. A certain layer may be considered impervious in a problem of shallow, horizontal flow over short distances, whereas it constitutes part of a complex semi-pervious layer in a problem of deep horizontal flow over great distances in an underlying pervious layer.
The layers containing ground water combine into aquifer systems. For a mathematical treatment of groundwater flow problems, an aquifer system should be relatively simple and belong to one of the following types (Fig.1):
- unconfined
- confined
- semi-confined.

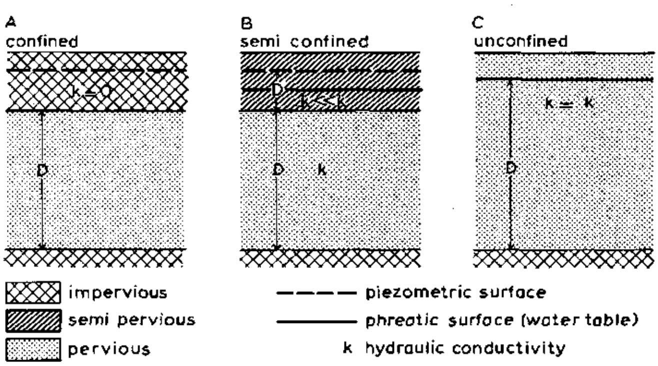


Fig.1. Aquifer types.

An unconfined aquifer, also called a phreatic or watertable aquifer, consists of the saturated part of a pervious layer which is underlain by an impervious layer (Fig.1C). The upper boundary is formed by a free water table (phreatic surface). The water in an unconfined aquifer is called unconfined or phreatic water.

A confined aquifer consists of a completely saturated pervious layer whose upper and lower boundaries are impervious layers (Fig.1A). Since completely impervious layers seldom occur near the surface, confined aquifers are rare in drainage problems. The water in wells tapping such aquifers stands above the top of the pervious layers. The water in a confined aquifer is called confined water.

A semi-confined (or leaky) aquifer consists of a completely saturated pervious layer (Fig.1B). In the covering layer a water table is present, often differing in height from the piezometric head (Chap.6, Vol.I) of the water confined within the pervious layer. Because of this difference in hydraulic head, there will be a vertical flow component tending to raise or to lower the water table. The latter, for example, occurs when the aquifer is pumped. The water in a semi-confined aquifer is called semi-confined water.

The term artesian water is ill-defined. Originally it was used for water in aquifers whose piezometric level was above ground surface. Thus, a well tapping such an aquifer is free-flowing (Fig.2). In literature one may find the term used for water in any confined or semi-confined aquifer, regardless of the elevation of the piezometric head above the phreatic level.

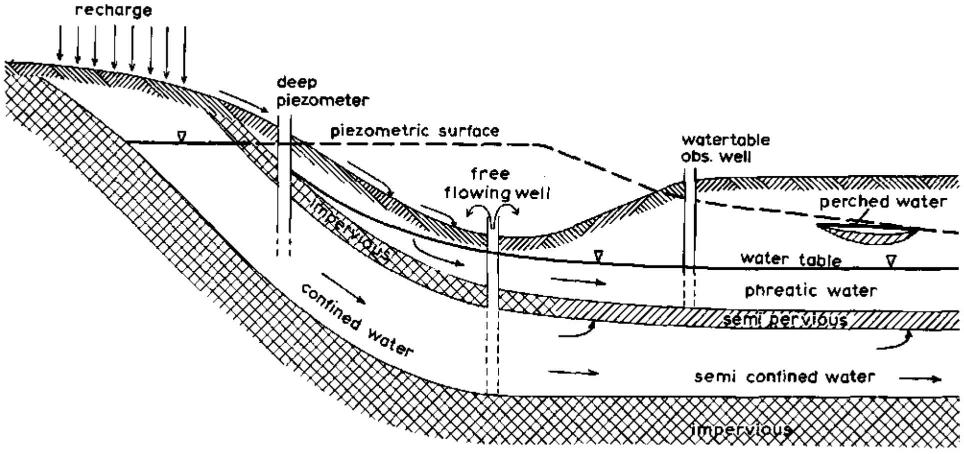


Fig.2. Cross-section of aquifer system.

Since, by definition, the covering layer of a confined aquifer does not transmit water, such aquifers are of little importance in drainage problems. In this and the following chapters, therefore, the aquifers considered are unconfined or semi-confined unless otherwise stated.

1.3 STREAM-FORMED PLAINS

Streams are one of the chief agents by which sediment is transported and deposited. When the stream's energy increases with increasing discharge, the water erodes and enlarges its channel and carries away the increased load until the load is in balance with the stream's transport capacity. When the stream's energy decreases, some of the load is dropped and the channel becomes shallower. The stream decreases its load by dropping those particles that require the most transport energy, and increases its load by picking up those particles that require the least energy. Thus due to the varying transport capacity of a stream, the available particles are sorted according to weight and size. Consequently stream deposits show a stratification of generally well-sorted sediments.
The stream's energy is at its lowest during base flow, i.e. when the river is fed by groundwater discharge only, and at its highest when the river is swollen due to large amounts of surface runoff (peak discharge). The energy, however, does not depend only on the volume of water but also on the gradient of the stream. A stream has a concave longitudinal section, i.e. the gradient decreases from headwaters to mouth (Fig.3).

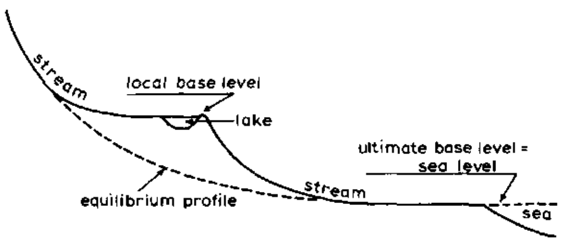


Fig.3. Longitudinal section of a river.

Obviously upon discharging into a large body of standing water (sea, lake) the stream's energy quickly reduces to zero. Hence a stream cannot cause any significant erosion below sea level. Consequently, near its mouth, the river's profile is tangential to sea level. Sea level is therefore called the ultimate base level of erosion, or simply the base level. The levels of lakes and other upstream bodies of standing water form local base levels. They disappear when the lake has been silted up completely or has otherwise vanished.

In equilibrium state the longitudinal section of a stream forms a smooth curve (Fig.3). The gradient of the curve decreases towards the sea and a condition of low energy is reached in which the elevation of the land is low, the slopes are gentle, and the stream load is reduced. However, the sea level, taken over long time spans, does not remain stable owing to such natural causes as climatic changes (e.g. glaciations) or tectonic movements of the ocean floors. Such events greatly influence the processes of erosion and sedimentation of a stream (Fig.4).

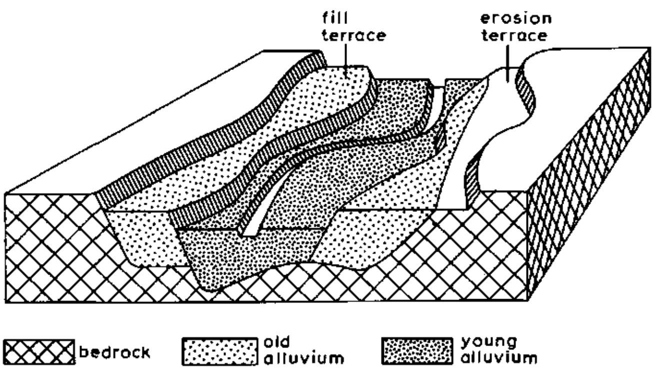


Fig.4. Aggraded valley plain.

Along a stream, from its source high in the mountains down to its mouth where it enters the sea, the following land forms are found:
- valleys and flood plains,
- alluvial fans,
- deltas,

which will be discussed separately in the following.

1.3.1 VALLEYS AND FLOOD PLAINS

In mountainous regions the valleys of streams are narrow and V-shaped in cross section. The stream occupies the entire valley floor and there is no space for large scale agricultural activity. The stream is still in its phase of down-cutting. In the middle and lower parts, where the longitudinal section of the river has already acquired a near equilibrium form, the erosion pattern changes from vertical to horizontal and broad valleys may develop. Depending on the hydrological regime, the stream may be classed as either a meandering river or a braided river (Fig.5).

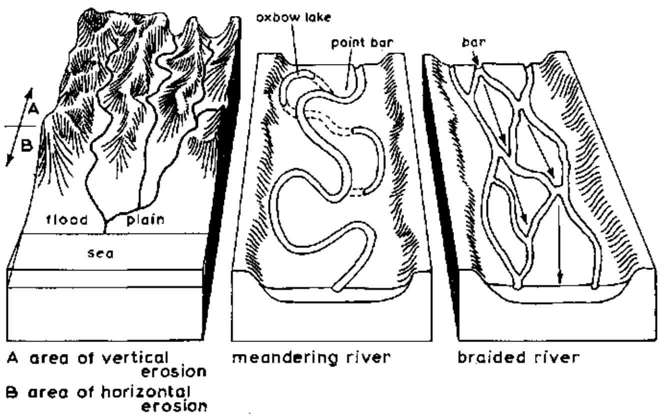
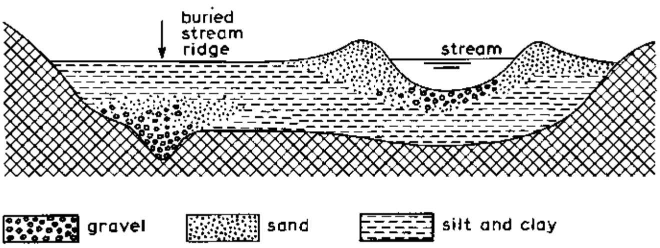


Fig.5. River types.

meandering river

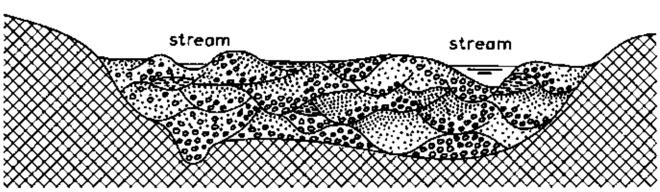
When the difference between base flow and peak discharge of the stream is not too great and when the bed is approaching its equilibrium profile, the stream will develop a sinusoidal form, made up of a large number of bends which are called meanders. The outer sides of the bends are eroded and the eroded material is deposited on the inner sides, forming point bars. As a result, the meanders move slowly outward and downstream, developing a flat valley floor.

During periods of peak discharge the water will overflow its banks and inundate all of the valley floor, which is therefore called a flood plain. When this happens the velocity and turbulence of the water decreases rapidly. The coarsest part of the suspended load (gravel and sand) settles down close to the stream channel, forming a natural levee. The finer particles come to rest farther away from the stream and the clay particles are deposited in shallow depressions known as backswamps. During the history of a valley, new stream channels develop regularly. The abandoned river beds (ox-bow lakes) fill up and, together with the levees, form a river ridge. Since these ridges are elevated and usually contain sandy material, they are well drained. The lower-lying basins are usually made up of poorly permeable clays. As a result, swamps are formed in which conditions are favourable for peat formation. Consequently flood plain deposits are characterized by extended, relatively thick and rather heterogeneous, predominantly fine-grained deposits, with intercalations of peat and buried stream-ridge deposits (Fig.6A).



A. meandering river.

Fig.6. Cross-sections of river valleys.



B. Braided river.

braided rivers

If there is a great difference between base flow and peak flow and the stream is loaded during peak discharge with coarse material, no meanders will be formed. Such conditions prevailed, for example, at the end of the Pleistocene glaciations when huge amount of debris were transported by the meltwater of the receding glaciers and ice sheets. Similar conditions occur in regions with a semi-arid climate where rivers with a highly variable discharge are found. During floods the river will erode the valley walls along more or less parallel lines and when the flood subsides the coarse-grained bed load will be left behind as bars and islands, obliging the stream to divide into a number of minor channels. Such a stream is said to be braided. The channels shift frequently, with the result that the deposits show characteristic scour and fill structures. Due to the varying transport capacity of each flood, the sediment as a whole is very heterogeneous but is predominantly coarse-grained

(Fig.6B). Hence, braided river sediments generally represent excellent aquifers.

groundwater conditions

Due to the climatic changes that took place at the end of the Pleistocene, many young flood plains are underlain by sediments of the braided river type. Consequently the river deposits of such a plain often show an upward grading from coarse to fine material. The upper finer-grained sediments, deposited by a meandering river, frequently form a poorly pervious layer, confining the water in the underlying pervious braided-river deposits (semi-confined aquifer). The latter are generally in hydraulic contact with the river, whose minimum level is often above the top of the coarse strata. Hence the water in these strata is under pressure. In humid areas the water table will usually be found at shallow depth and corresponds to the mean river level.

During the high stage of the river the piezometric surface (Chap.6, Vol.I) of the water in the underlying aquifer will rise above the water table and there will be upward groundwater flow from the sand and gravel layers into the overlying clayey deposits (Fig.7A). This upward flow contributes to the high water table, resulting in waterlogging in the backswamps and other local depressions of the flood plain. Near large meandering rivers like the Rhine, Po, Danube, Hwang Ho, and many others, these seepage phenomena can clearly be seen throughout most of the year.

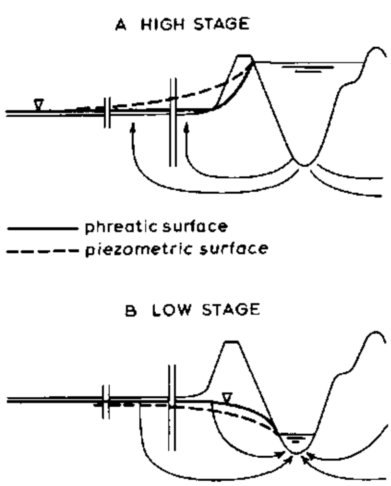


Fig.7. The influence of a river on the groundwater regime of a flood plain.

During low stages of the river the piezometric surface will drop below the water table and a natural drainage flow will occur from the semi-pervious layer through the underlying coarse layers towards the river (Fig.7B). This natural drainage, however, is often insufficient to cope with the excess water from seepage and precipitation.

In arid regions the water table is sometimes found at greater depth. However, the stream losses (influent seepage, Chap.10, Vol.II) may build up a groundwater mound (Chap.21, Vol.III). When the groundwater level rises to close to the soil surface, salinization may occur, which renders leaching and drainage necessary.

1.3.2 ALLUVIAL FANS

Sometimes the transition between the mountainous area and the area of much smoother topography is gradual; sometimes it is abrupt: for example, when caused by faults. At such a sharp transition the transport capacity of the river decreases suddenly because it diverges in numerous channels over the plain at the foot of the mountains. The resulting deposition of alluvium is chiefly concentrated at the foot of the mountains in the form of a fan (Fig.8).

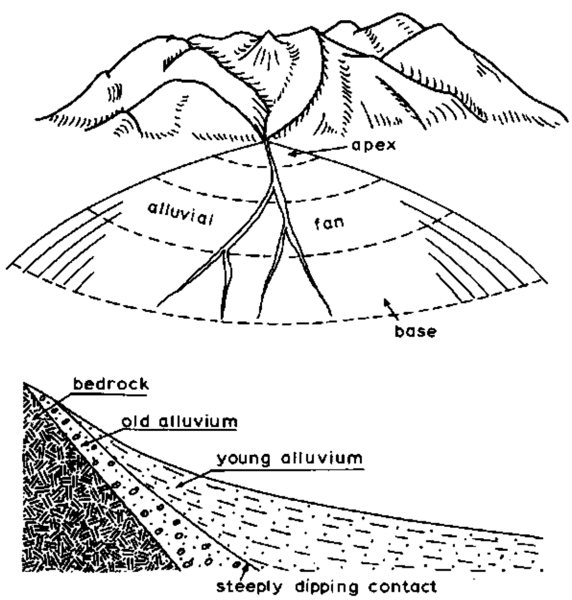


Fig.8. Structure of an alluvial fan.

The sorting of the deposited material is relatively poor, but there is a tendency for the coarsest material (often up to boulder size) to settle near the top (or apex) of the fan, while silt and clay are deposited at its base. In major discharge channels, however, coarse-grained material may be carried far downstream. Alluvial fans vary considerably in size. Their radius may be up to 50 kilometers. The angle of dip of the fan surface rarely exceeds 10^{o} and there are many alluvial fans with an angle of less than 5^{o} or 6^{o}.

Alluvial fans are found near areas of bold relief, and their development is most conspicuous under moderately arid to semi-arid conditions. Characteristic of such climatic conditions are brief and infrequent periods of heavy rainfall. Alluvial fans also develop under humid conditions: for example, along the Alps and the Himalayas. They are flatter than the fans of arid regions owing to the abundance of running water which favours the development of gentler gradients. When a large number of rivers discharge along a steep mountain front, the fans of the individual streams often coalesce into a piedmont plain (or bahada). After heavy rainfall a river emerging from deep mountain valleys is loaded with detrital material. This material fills up the existing channels and causes the formation of new channels in another lower-lying section of the fan. This process is repeated again and again until the mountain stream and the alluvial fan have reached a stage of equilibrium. Three depositing agents can be distinguished on alluvial fans:
- sheet floods
- stream floods
- streams.

Sheet floods occur when large amounts of water and detritus emerge from the mountain valley. This viscous material tends to spread out in the form of a sheet covering all or parts of the fan.

Stream floods are confined to definite channels and refer to floods caused by a lesser amount of water. Their spasmodic and impetuous character is such that the term "stream flood", rather than "stream", is applied. The deposits of violent stream floods tend to be identical with those of sheet floods except that they lack the lateral extension.

Streams require a steady, rather than an abundant, supply of water from the mountains. Since a steady supply is largely lacking in the arid and semi-arid regions, the action of streams in such regions is insignificant. In more humid regions, however, stream deposits are of considerable importance.

The grain-size distribution of fan deposits varies widely and is a function of:

- The range in particle size of the original detritus.
- The type of the transporting and depositing agent. Sheet floods form deposits with a very low degree of sorting; stream deposits show fair to good sorting, whilst stream flood deposits occupy an intermediate position.
- The distance the material is transported. Material transported over short distances is poorly sorted. Therefore, the deposits near the apex commonly have a lower hydraulic conductivity than would be expected from their grain size. The deposits in the central parts of the fan, though less coarse than in the apex, are better sorted and may have a fairly high hydraulic conductivity. The best sorted sediments occur near the base of the fan; because of their fine texture, the hydraulic conductivity is relatively low.

Alluvial fan deposits are laid down in beds approximately parallel to the surface of the fan. One might therefore expect a fair stratification, but, in fact, the fairly complicated development of most alluvial fans causes a complex internal structure. Layers of a particular grain size often vary widely in thickness as well as in areal extent. Sandy material often represents lenticular stream deposits, while mud-flow deposits are laid down in more continuous sheets.
Interfingering of fine and coarse-grained layers is also a common feature, layers of coarse sand often wedging out in downstream direction.

groundwater conditions

In hydrological terms an alluvial fan can be divided into three zones (Fig.9):

- the recharge zone
- the transmission zone
- the discharge zone.

The recharge zone comprises the pervious gravel fields at the head of the alluvial fan. Because of the generally coarse-grained deposits, the ground water in this zone is unconfined. The water table is usually relatively deep and rather flat due to the high permeability of the gravels and coarse sands. The aquifer is recharged by infiltrating precipitation, runoff from the mountain front, losses from the river channels, percolating flood waters, and by subsurface inflow through the gravel fill of the valley mouth.

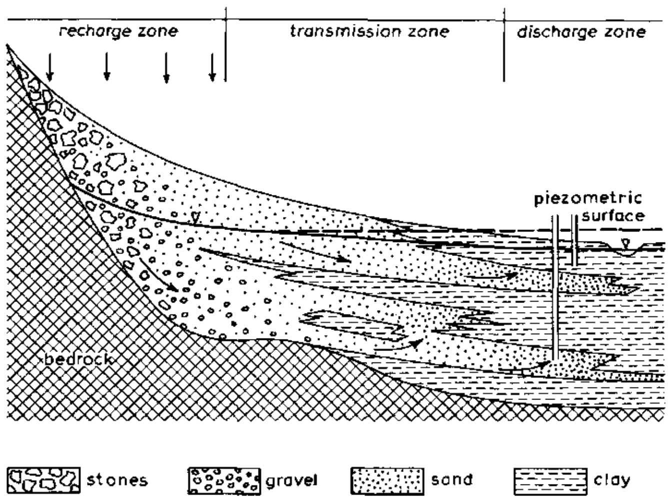


Fig.9. Cross-section of an alluvial fan.

The transmission zone starts where clay layers are found in the subsoil; as a consequence, the deeper pervious layers are semi-confined aquifers. The surface layers present a phreatic aquifer that is recharged by flood water, rainfall channel losses, etc. Under the influence of the differences in hydraulic head between the various pervious layers, water will flow upward and recharge the phreatic reservoirs. The topographic slope is generally steeper than the slopes of the water table and of the piezometric surface. This means that in downstream direction the water table is increasingly closer to the ground surface although it is seldom found at dangerously high levels in either the recharge or the transmission zone. The piezometric levels become higher and may even rise above ground surface, so that deep wells may yield free-flowing water.

The discharge zone is found in the lower part of the fan where the topographic slope is slight and the water table shallow. Here too, the water in the deeper layers is under pressure and a vertical upward flow exists. Springs are often found at the foot of the fan, yielding water of better quality than that of the shallow ground water, which, especially in dry regions, may be quite salty due to the high evaporation rates. Drainage problems are generally limited to this part of the fan.

1.3.3 DELTAS

A phenomenon comparable to the formation of alluvial fans is that of a stream flowing into a body of standing water (a lake or a sea) and forming a delta. The flow of the stream is checked by friction as the stream water diffuses into the standing water. The stream loses energy and deposits its load. This process of sedimentation is enhanced when the standing water is a salt water body. The salts dissolved in sea water tend to coagulate or flocculate the suspended fine particles into aggregates so large that they promptly settle to the bottom.

In a typical delta three types of deposits may be recognized (Fig.10):
- top-set beds
- fore-set beds
- bottom-set beds.

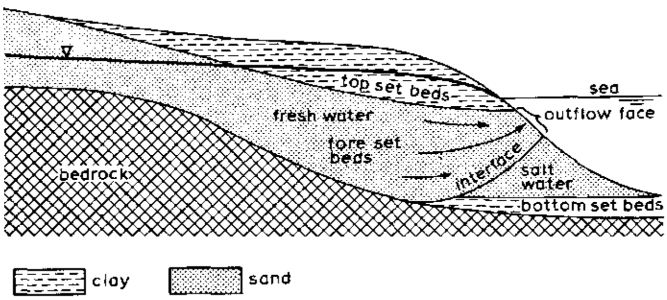


Fig.10. Cross-section of a delta, showing interface of fresh and salt water and outflow face at the coast.

The bulk of material supplied, which is mostly sandy, is deposited under water in regularly stratified, inclined layers (fore-set beds). The mud is carried farther forward and settles in more or less horizontal layers (bottom-set beds). As a result of the lengthening of its course, the stream channel has to raise itself to retain its equilibrium profile. During this process horizontal layers (top-set beds) accumulate in the upstream part of the delta and the original valley mouth. Consequently, the delta slowly rises above sea level.

Many of the characteristics of valley deposits can be recognized in the top-set beds: natural levees of relatively coarse material along the stream channels and relatively fine material on the flats between the channels. However, not all rivers build up deltas. At the mouth of many rivers the material brought into the sea is swept away by marine currents and comes to rest somewhere on the sea bottom.

groundwater conditions

The top layers of a delta are finer grained than the underlying fore-set beds; hence semi-confined aquifer conditions often occur. Owing to the influence of the tide on the deep ground water, the piezometric levels near the sea will also reflect the tidal movement (Chap.13, Vol.II).

Since the bulk of the deltaic sediments has been deposited in a marine environment, the ground water in the deeper layers is initially entrapped sea water. Under the influence of the flow of fresh ground water from the valley towards the sea, the salty ground water will be slowly replaced by fresh water (Fig.10). This replacement will be effective throughout the delta except in the coastal area, where sea water intrudes, and here a fresh-water layer will be floating on salt water. The fresh-water body will be moving seawards because its phreatic level is above sea level; it flows out in a narrow zone at the coast. The initially sharp interface between the salt and fresh water bodies will, owing to diffusion and dispersion, gradually pass into a brackish transition layer. The rate at which this transition layer develops depends on various factors, one of which is the permeability of the aquifer material.

If the deltaic sediments consist of two coarse-grained layers separated by an impervious clay bed at a depth slightly below sea level, the upper aquifer may contain fresh water only, whereas the lower aquifer will contain fresh water floating on salt water near the coast. The interface of the two water bodies is as indicated in Fig.11.

1.4 COASTAL PLAINS

A coastal plain is an emerged part of the continental shelf. It may be a very narrow or even fragmentary strip of the former sea floor exposed along the

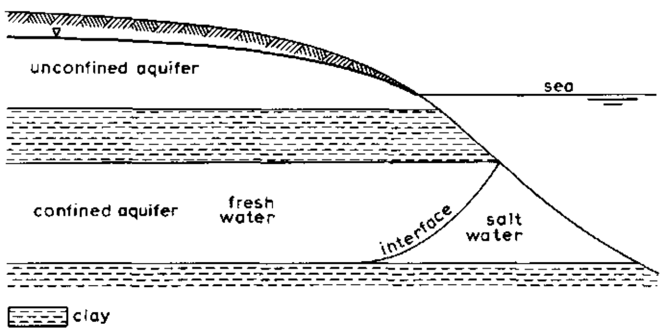
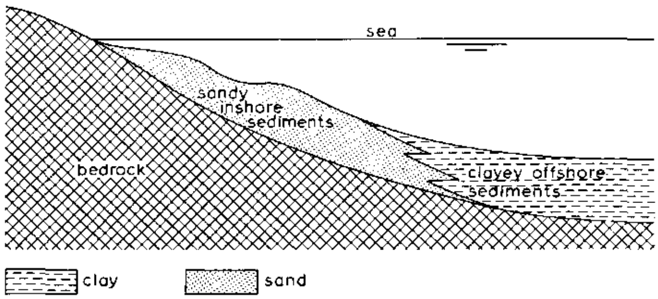
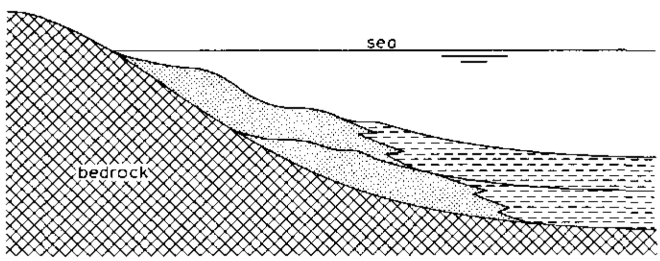


Fig.11. Cross-section of a delta, showing an upper clay layer intersecting the interface of fresh and salt water. The salt water body in the confined aquifer has the form of a wedge.

margin of an old land area, or it may be a vast, almost featureless plain, fringing hundreds of kilometers of coastline.
The sea floor emerges either because it is uplifted by regional crustal movements, or because of a universal drop in sea level, for instance due to glaciation. There are young coastal plains which have recently emerged and there are others formed in some earlier geological time and now lying far inland from the sea. In general, uplift of the sea floor is not a single or sudden event, nor is it continuous. Normally it is intermittent and often it is interrupted by resubmergence. This phenomenon has a great bearing on the internal structure of a coastal plain, i.e. on the horizontal and vertical distribution of coarse- and fine-grained materials. The sandy nearshore sediments usually pass laterally into more clayey offshore sediments (Fig.12). When the sea level is rising, the various environments of deposition migrate landwards. Sediments of deeper-water zones cover areas of the sea floor where formerly only shallow-water sediments accumulated. Thus the alternating beds of clay and sand in the subsurface of coastal plains reflect migrating environments of deposition due to a changing sea level.



A. Sea level is constant. Sandy sediments are inshore and clayey sediments offshore in belts parallel to the coast.



B. Rise in sea level to new constant level. Sediments are deposited as in A. New sediments overlap the old.

Fig.12. Deposition of sediments along the coast.

The sediment material is debris, either transported from the mainland by streams or formed by marine erosion of shores near and remote. The sediments are distributed in extensive beds and sorted by marine currents. While sedimentation is in progress, the sorting action of the water is remarkably delicate. The layers of sediment commonly show a great perfection of stratification. As deposited, the layers of a coastal plain have a gentle slope seaward, corresponding to the slope of the sea floor. The deeper layers often show a steeper dip due to downbending of the sea floor as the layers were deposited. If the offshore slopes are gentle, waves will not be able to attack a shoreline vigorously because the larger ones will break farther away from the coast. There a submarine sand bar will develop, which in time will emerge as a barrier island or barrier chain with a lagoon behind it. Numerous breaches in the barrier chain will be maintained as tidal inlets, particularly opposite the mouths of streams. Thus the lower parts of such streams are at times essentially tidal rivers. The lagoons are the sites where fine material is deposited. During low tide extensive tidal flats may be exposed in the lagoonal areas behind the barrier chains.

The surface of a newly emerged coastal plain is rather flat and slopes gently seaward. It may also be slightly undulating with large depressions containing swamps and lakes. The short and small streams which originate on the newly uplifted coastal plain are known as consequent streams. Their origin and position are determined by the initial slope of the newly formed land. In general they take a course at about normal angles to the coastline, running parallel to each other. If the streams rise in the old land and extend their courses across the plain, they are called extended consequents.

As time passes, consequent streams may develop natural levees. If, furthermore, a barrier bar or a dune ridge is present along the coast, the area between two such rivers is a closed basin in which water is standing for a certain period of time. In these quiet environments very fine-grained sediments are deposited, leading to the formation of swampy areas composed of heavy clay soils with poor internal drainage (Chap.2, Vol.I). Such conditions exist, for instance, on some of the coastal plains of southern Turkey.

While the bulk of the coastal plain sediments are deposited in marine environments, the upper layers may be of fluvial origin, at least in the areas bordering the old land. Very often one can distinguish a transition from river sediments to marine sediments from the old land seaward. Changing sea levels often give rise to the formation of terraces. When coastal plains are uplifted well above sea level, they become increasingly dissected by stream erosion. A typical network of natural drainage courses may develop. This erosion finally results in characteristic landscapes known as cuesta landscapes, an example of which is the Paris Basin. The development of such landscapes will not be discussed here.

Beach formations, sand bars, and dune ridges often comprise the younger portions of coastal plains. Splendid examples of coastal plains are found along the Atlantic and the Gulf Coasts of the United States and in south-eastern England.

groundwater conditions

Groundwater conditions of coastal plains are complex, as will be obvious from the above description. In the outer lowlands of a newly emerged coastal plain, the groundwater table is shallow and natural drainage is often poor. Dune ridges along the coast and the higher river levees of the consequent streams may enclose vast depressions without a visible outlet, thus giving rise to vast swamps. The inner lowlands, bordering the old land, have deeper water tables and are better drained because the soils are more pervious owing to their coarser texture. Here are found the outcrops and recharge areas of the aquifers encountered at greater depths in the outer lowlands of the plain (Fig.13).

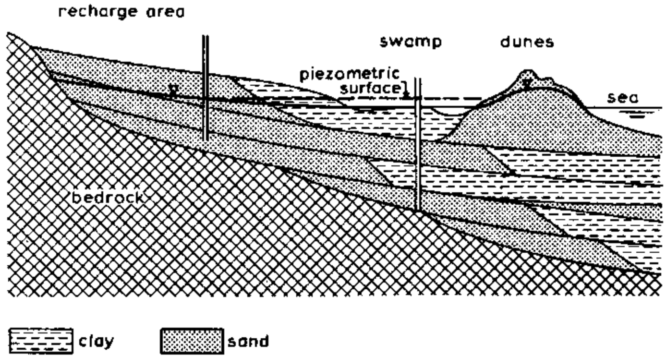


Fig.13. Groundwater conditions in a coastal plain. In the recharge area there is unconfined groundwater. Towards the sea, the groundwater in the deeper sandy sediments is confined. Behind the dune ridge are swamps: shallow water tables and upward seepage flow from the deeper confined sandy layers.

The groundwater in the deeper layers of lower portions of coastal plains is usually semi-confined and consequently an upward seepage flow may exist. The principal reasons for this are:

- The aquifers dip seaward and are often bounded below and above by impervious or poorly pervious layers.

- A feature common to most coastal plains is that the nearer to the sea the finer grained and less pervious the aquifer material becomes (offshore sediments).
- Many water-bearing layers, which may initially be very thick, wedge out seaward into impervious or poorly pervious layers.

Since coastal plain deposits are laid down in different environments, different groundwater qualities can be expected. Water of low salt content is generally found in the outcrop of the aquifers, both near the surface and at greater depth. Near the coast the groundwater becomes brackish, and in the deeper layers entrapped sea water may be present.

1.5 LACUSTRINE PLAINS

As far as their origin is concerned, there are many different types of lakes: glacial, river-made, volcanic, fault-basin, and land-slide lakes. They must all be regarded as young landscape features. Most of them disappear in the course of time: they either fill up with sediments deposited by inflowing rivers or they are drained when an outlet of sufficient depth has developed. When this happens the lake floor emerges and can be used for agricultural and other activities.

Lake floors are characterized by:
- Flatness. The fine-grained sediments, whether of glacial or fluvial origin, have smoothed the floor in such a way that it has become entirely flat.
- "Fossil" coastal features. In huge bodies of standing water the action of waves produces such morphological features as beaches, cliffs, and wave-cut platforms, while the combined action of waves and currents may produce sand bars and spits. Sand bars are sand ridges formed under water and running parallel to the shore. Spits resemble sand bars, but are connected at one end to a headland. The coarse material required for their formation is provided by erosion caused by the waves hammering the rocks surrounding the lake. These fossil beaches, cliffs, and terraces, outlining the former extent of the lake, are the most significant evidences of a former lake.
- Delta structures, occurring where streams used to flow into the former lake. The sediments are coarse-grained, although the covering sediments may be more

finely textured. They are sometimes entirely of fluvial origin and, therefore, totally different from the actual lake floor sediments.

- Sediments generally made up of finely laminated clays. The inflowing streams carried large quantities of material into the lake, of which the coarse material settled near the river mouth, the finer particles, such as silt and clay, being transported further into the lake. Lake-inwards, therefore, sediments grade from coarse-grained to progressively finer-grained deposits.
- Poor drainage. If the newly emerged lake floor has no visible outlet, inflowing streams may flood large areas, especially during the rainy season. The water collects in local depressions, where swamps may form and where - in warmer climates - the water will evaporate, leaving behind the suspended sediments, mixed with fine salt crystals. Deposits from such ephemeral water bodies commonly build up clay-surface plains of extraordinary flatness, called playas.

groundwater conditions

Since there is such a wide variety of lakes, there is also a variety of the groundwater conditions in lake plains. In general, lake plain sediments are fine-grained and therefore do not transmit large quantities of groundwater. However, there are exceptions, for example at those sites where coarse sediments, deposited along the rim of the lake have later been covered by fine sediments. Here deep wells may yield large amounts of water,sometimes even free-flowing.

In humid regions, lake plains are often the sites of rich agricultural land, although artificial drainage is usually required. Water quality problems do not exist as the groundwater is fresh and is not subject to mineralization. In arid and semi-arid regions the situation may be quite different. The shallow water tables, which are found along the inflowing streams, often cause a strong capillary movement, rendering the soils salty. Leaching of these salts by rain or river water may cause an increase of the salt content of the groundwater. The leaching of buried playa deposits, if present, may add to the mineralization of the deeper groundwater.

In the flat lake plains one might expect the water table to be rather flat as well. Locally, however, there may be large water-table gradients. In the recharge areas the water table may be very high while in other parts of the plain, where rivers are absent, a very deep water table may be found. Despite these large water-table gradients, the low transmissivity of the lake sediments prevents a rapid flow of the groundwater to the areas of low groundwater heads.

1.6 GLACIAL PLAINS

During the last one to two million years a major part of the northern hemisphere has repeatedly been under the influence of advancing and receding glaciers and continental ice-sheets. The erosion caused by these ice masses on the underlying hardrocks was very intensive and smoothed the preglacial land forms. Because of the low level of the sea during glacial periods, the melt waters locally eroded deep valleys and channels in those regions where thick sedimentary deposits occurred.
The transport capacity of glaciers and land-ice sheets is extremely great, but their sorting capacity is almost zero. Hence, moraine deposits which result directly from the ice (e.g. glacial till or terminal moraines formed at the ice front) are unsorted.

When the ice is advancing, it first penetrates existing valleys, as a result of which the discharge of the rivers is hampered and the water is forced to take another course to the sea. The slowly moving ice exerts a strong force on the valley walls, which are pushed up. Under the influence of the low level of the sea, the meltwater from the ice sheets erodes deep gullies and channels (During the Saale glaciation, channels as deep as 100 m below the present sea level were formed). The course of these channels differs strongly from that of the preglacial water courses, as far as their direction is concerned. When the ice-sheets recede, these glacial meltwater channels are filled with coarse and very coarse fluvioglacial deposits. These channels are even sometimes overfilled and there are extensive areas in front of the ice-sheets and terminal moraines where thick layers of fluvioglacial material are laid down, giving rise to broad outwash plains. The finest particles settle in glacial lakes where they form the well-known varves (alternating laminae of silt and clay).

Glacial plains left behind after the recession of ice-sheets are usually gently undulating, with numerous local depressions in which water is standing (glacial lakes). In the following interglacial period, with its warmer climate, peat growth even occurs in such depressions. The drainage pattern on an emerged glacial plain is at first undeveloped, but gradually a new system of streams and water courses comes into being, which is often different in capacity and direction from the underlying, buried glacial channels.

During glacial periods, periglacial climatic conditions (cold and dry) prevail in the regions in front of the ice-sheets. In such regions with permanently frozen soils (permafrost) the principal agent of deposition is the wind. It transports sand from the barren land and deposits it elsewhere, giving rise to more or less thick layers of wind-blown sand, which may cover extensive areas. Further away and at the lee-side of hills, the finer silt particles may come to rest as thick layers of loess.

groundwater conditions

Young glacial plains are often characterized by poor drainage conditions. This is mainly due to the low permeability of the unsorted glacial deposits, such as boulder clay, till, etc. Broad glacial till plains that are undissected by erosion generally have a flat and shallow water table. Precipitation may cause flooding and a further rise of the groundwater level (Fig.14A).
Where such plains are dissected by streams(Fig.14B) or where a more undulating landscape prevails (Fig.14C), the drainage conditions on the higher ground may be better. The excess water (mostly surface runoff) will collect in the depressions and low-lying valley floors, often leading to flooding and high groundwater tables.

Eroding ice sheets and melt water sometimes cause the formation of deep fluvioglacial channels between ridges of bedrock or till (Fig.14D). The channels may be filled with coarse out-wash (or fluvioglacial) deposits or till. When filled with out-wash deposits, the buried glacial channels may serve as good subsurface drainage channels. In such a landscape of undulating topography there is a groundwater flow from the higher ridges towards the lowlands in between. At the foot of these ridges one may find discharge areas, wet soils, and high water tables. Often the high ridges are composed of coarse materials in which a rapid groundwater flow prevails. In general, no drainage problems are encountered on these high areas. Where coarse-grained out-wash deposits wedge out into the less permeable till (Fig.15), the groundwater may be confined or semi-confined.

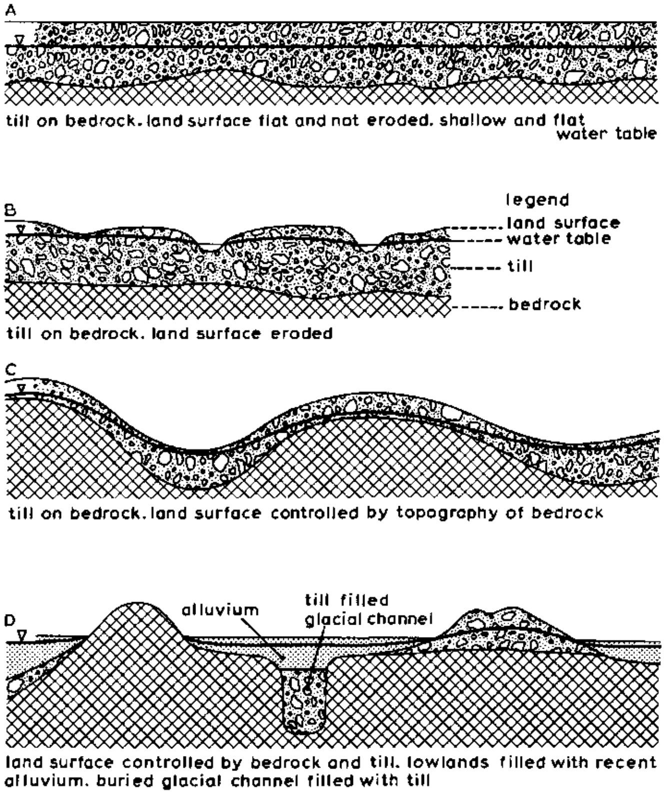


Fig.14. Groundwater conditions in glacial landforms.

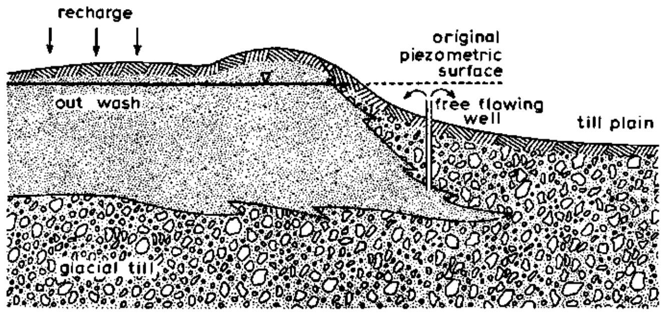


Fig.15. Outwash deposits wedging out into glacial till. A free water table in the outwash on the left side. Confined groundwater in the outwash wedge under the till plain on the right side.

1.7 LOESS PLAINS

In desert regions and on bare agricultural lands in semi-arid regions, deflation, i.e. the picking up and removal of loose particles by the wind, may cause serious erosion. In a few extremely dry years, losses of more than one meter of soil were reported in the USA.

The blown-away particles are carried in two separate layers:

- The lower layer consists of sand grains, and rises less than a meter above the ground surface. The sand is not transported very far and settles as dunes or ripples.
- The upper layer consists of clouds of silt (and some clay) particles, which may rise to heights exceeding 3000 meters and may be carried over great distances. When deposited it is called loess, which is recognizable - in the absence of more conspicuous sedimentary forms - by a uniform grain size (10-15 micron). Loess layers of over 30 m thick have been reported; they thin out away from the source area.

The Quaternary loess deposits are of two types:

- Glacial loess, mostly found in Northern Europe and the northern part of the U.S.A. These deposits consist of fine material picked up from barren areas in front of the ice-sheets, with a periglacial climate (dry and cold).
- Desert loess, found, for example, in western China. These deposits, which are sometimes more than 60 m thick, consist of material blown from the desert basins of central Asia.

A further type, the so-called "loess-like deposits" are found in certain areas, for example in the Great Plain of Hungary. Such sediments are redeposited loess that was eroded by rivers and deposited under water elsewhere. These sediments differ from true loess by their slightly higher clay content.

groundwater conditions

Loess deposits usually have a good permeability. They represent deep unconfined aquifers, which, when dissected by rivers, drain easily towards the valleys. Local depressions may have a thin veneer of clayey material, which hampers infiltration, thus often causing surface drainage problems. Moreover, thin layers of rather impervious concretionary limestone may develop, giving rise to perched water tables (Fig.16).

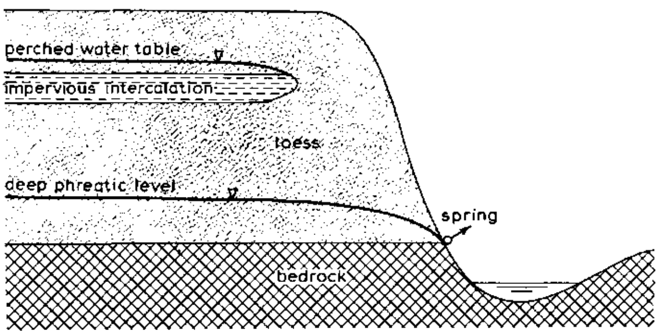


Fig.16. Groundwater conditions in loess deposits.

1.8 GROUNDWATER CONDITIONS IN AREAS AFFECTED BY DISSOLUTION

A process that takes place mostly below the surface is the dissolution of limestone by percolating water, resulting in large holes, caverns, and channels, through which water can flow easily. A typical topography (karst topography) often forms at the surface, consisting of dry valleys, sinks, disappearing streams, and large springs. Drainage problems will not arise when the soluble limestone is near the surface and if the surface has a relatively high topographical position. The dry valleys (Fig.17) carry water only after heavy precipitation, and losses to the underground channels in the limestone are then extremely high.

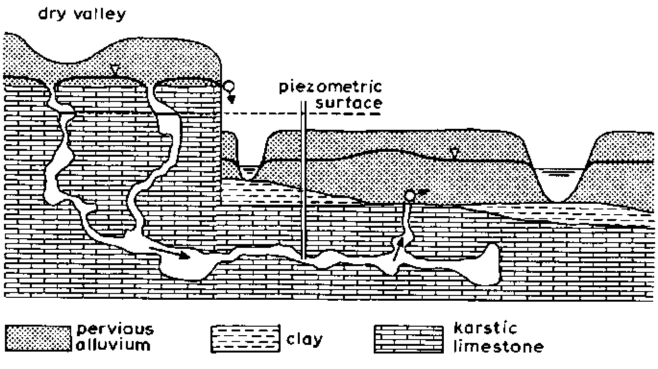


Fig.17. Groundwater conditions in a karst region.

When karstic limestone dips below deeply incised valleys or alluvial plains, buried springs may occur locally, causing groundwater mounds in the alluvial cover. Water from karst sources contains calcium bicarbonate and its biological quality is often poor because of the fast transport from recharge to discharge area. Examples of plains where the groundwater regime is influenced by karst phenomena are:

- The Danube flood plain (Roumania) where the drainage problems are partly caused by an intensive feeding of the groundwater in the river alluvium from underlying karstic limestones.
- The Konya-Plain (Turkey), which is underlain by thick cavernous limestone, cropping out along the rim of the plain. This karstic limestone is an excellent aquifer and it is recharged by precipitation and runoff from the surrounding hills. It dips under the poorly pervious lake sediments, which comprise the bulk of the basin fill. From the rim of the plain inward the water in this limestone aquifer yields free-flowing water.

1.9 INFLUENCE OF FAULTS

Owing to uplifting or downwarping forces in the interior of the earth, the layers of the crust are subject to great strains. As a result, fractures (faults) may develop along which blocks of the crust may move. Faults may be due either to shearing under compression or to tearing apart under tension. Faults caused by compression tend to be tightly closed and generally act as groundwater barriers. Faults caused by tension are more irregular, rough, and open, and groundwater may move upward along them, sometimes from great depth. Both types of faults may bring aquifers into contact with impervious rock and cause the normal groundwater flow to stop abruptly. As a result springs may develop. If coarse sandy layers are offset by a fault, the layers in the fault zone are downbent and may assume an almost vertical position. Clay layers or lenses within the coarse sediments will cause a high resistance to horizontally-moving groundwater. As a result, great differences in water table height are found on either side of the outcropping fault. These differences may be of the order of from several meters to 20 or 30 meters, as is known from the tectonic blocks in the Ville in W.-Germany. The more highly elevated land may be waterlogged (Fig.18) and springs and marshes may occur. The lower-lying soils downstream of the fault have a deep water table and are dry.

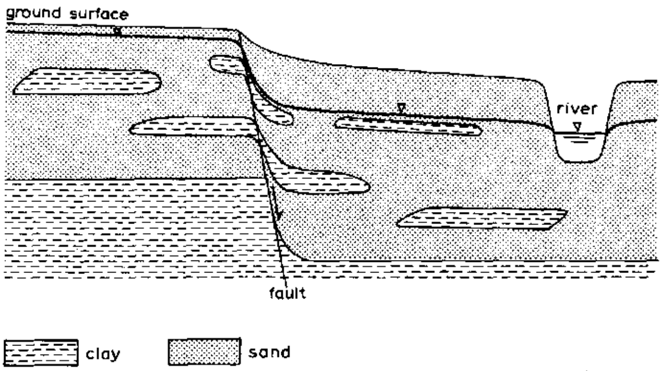


Fig.18. The effect of faulting on the water table elevation. The fault is partly sealed by downbent thin clay layers and thus acts as a barrier to groundwater flow which is from left to right. The high land has a shallow water table. The sunken land has a deeper water table. If the fault is completely sealed, groundwater is discharged through springs occurring along the fault.

When a block of the earth's crust sinks away between two parallel faults, a graben or tectonic valley is formed (Fig.19). The valley walls consist of mountains (horsts), and may be quite steep. Springs often occur on the faults, discharging, from great depth, water that is sometimes warm and mineralized.

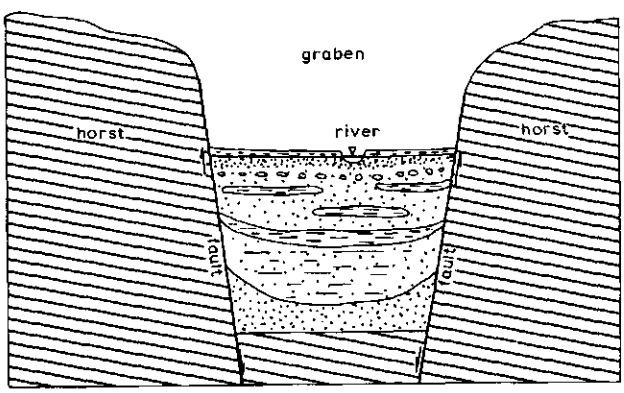


Fig.19. Valley of tectonic origin.

Examples of tectonic valleys are the valleys of the Jordan River in the Near East, the Owens River in California, the Menderes and Gediz Rivers in Turkey. At present they are being filled up with river deposits. Beneath these deposits, depending on the geologic history of the region, one may find marine, aeolian, lacustrine, volcanic, or even glacial material, the thickness of which may exceed 1000 m. The Central Valley in California is filled with more than 7000 m of sediments, mostly of marine origin and containing saline water. These are overlain by non-marine sediments, 200 to 1000 m thick, containing fresh water.

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INTRODUCTORY SUBJECTS

2. SOILS AND SOIL PROPERTIES

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PURPOSE AND SCOPE

Elucidation of a number of physical and chemical characteristics of soils - and related aspects - which have a direct relation to agricultural water management and plant growth.

CONTENTS

			page
2.1	General		36
	2.1.1	Soil and land	36
	2.1.2	Major components and functions of soil in relation to plant growth	37
	2.1.3	Factors of soil formation	37
	2.1.4	The soil profile	39
	2.1.5	Outline of normal soil investigations	41
	2.1.6	Relationships between basic soil characteristics, physical soil properties, and agricultural qualities	42
2.2	Basic soil characteristics		44
	2.2.1	Soil texture	44
	2.2.2	Mineralogical composition	48
	2.2.3	Physico-chemical characteristics of clays	50
	2.2.4	Organic matter and soil fauna	52
2.3	Physical properties of mineral soils		54
	2.3.1	Soil porosity	54
	2.3.2	Soil structure	56
	2.3.3	Soil consistence	60
	2.3.4	Soil colour	61
2.4	Soil moisture		63
	2.4.1	Water-transmitting properties	63
	2.4.2	Total readily available moisture	64

		page
2.5	Soil air	65
2.5.1	Composition of soil air	66
2.5.2	Volume of soil air	66
2.5.3	Rate of oxygen supply	66
2.5.4	Plant requirements	67
2.5.5	Aeration condition and soil processes	67
2.5.6	Soil aeration and drainage	68
2.6	Soil temperature	68
2.6.1	Soil temperature and plant growth	68
2.6.2	Soil temperature and drainage	69
2.7	Soil fertility and productivity	69
2.7.1	Definitions	69
2.7.2	Supply of nutrients	70
Literature		72

2.1 GENERAL

2.1.1 SOIL AND LAND

The term "soil" is often used loosely and conveys different things to different people, including soil scientists.
For civil engineers it is a general term for unconsolidated earth as distinguished from solid rock.
The soil physicist considers soil a porous medium, suitable to be treated mathematically.
The soil chemist sees the soil as a powder, mostly coloured, fine or coarse grained, with an upper limit of 2 mm (fine earth), having complicated chemical and physical properties.
The pedologist regards the soil as a natural body, and is mainly interested in the result of biochemical weathering of the soil parent material, the soil profile with its various pedogenetic layers or horizons. The pedologist examines and classifies soils as they occur in their natural environment. He lays minor emphasis on their practical utilization. Nevertheless his findings may be as useful to highway and construction engineers as to the farmer or agronomist.
For the soil surveyor soil is a collection of natural bodies occupying portions of the earth's surface. They support plants and have properties due to the integrated effect of climate and living matter acting upon parent material, as conditioned by relief, over periods of time. In this sense soils are landscapes as well as profiles (layered vertical sections). For the agronomist the soil is a medium for plant growth and he is especially interested in the conditions of the topsoil.

Finally the drainage specialist, who may be regarded as representing a combination of civil-engineer, hydrologist, soil physicist, and agronomist, is primarily concerned with soil properties which affect the movement of water into and through soils.

The term "land" comprises more than the term soil. There are many definitions of land. In agricultural land classifications, land is considered a complex of all the factors above, on, and below the earth's surface, which affect man's agricultural, pastoral, and forestry activities.

In the physical environment that determines land use or natural vegetation, the soil is only one factor. Others are the climate and the land features associated with the soil: topography and hydrology.

2.1.2 MAJOR COMPONENTS AND FUNCTIONS OF SOIL IN RELATION TO PLANT GROWTH

The soil can be regarded as a porous medium, i.e. a material system in which solid, liquid and/or gaseous materials are present.
In a mineral soil, the mineral materials occupy a substantial volumetric fraction, up to 50-60% by volume. The organic matter content of a mineral soil is usually very low, generally less than 3%.
Organic soils have an organic matter content of more than 20% by weight. These soils are also referred to as peat or peaty soils (organic matter largely undecomposed) or muck soils (organic matter largely decomposed).

Soils provide crops with essential plant nutrients, in addition to water and oxygen for root respiration. Unless the supply of water and oxygen can be maintained, the rate of uptake of nutrients is reduced.
Other aspects of soil which have a bearing on plant growth are (Chap.4, Vol.I):
- its temperature should be favourable to plant growth,
- its mechanical resistance to the movements of roots and shoots should not be too high,
- it should provide an environment free of chemical or biological conditions detrimental to plant growth, such as extreme acidity, excess soluble salts, toxic substances, disease organisms.

2.1.3 FACTORS OF SOIL FORMATION

The soil is a product of the action of climate on the parent rock at the earth's surface, as modified by landscape (topography and hydrology) and vegetation, the final result depending greatly on the factor time. The wide variety in each of these factors throughout the world is responsible for the many different soils that occur (see for example BUNTING, 1965 or PAPADAKIS, 1969).

parent material

Locally, the parent material may be the predominant factor that determines the soil. Such material may be basic or acid, calcareous or non-calcareous, etc. (Compare: granite, basalt, shale, sandstone, limestone, loess, alluvial deposits, etc.). The mineralogical and granular composition of the parent material (rock or unconsolidated sediments) will greatly determine the nature of the soil that results from its weathering. For example, clay soils are formed from basalt, a fine-grained rock low in quartz and consisting almost entirely of easily weatherable silicates, which can yield clay upon weathering.

topography or land form

Relief, slope, and physiographic position affect the soil because they control both the amount of percolating rainwater or floodwater and the quantity and rate of surface runoff. These factors therefore influence erosion, deposition, and groundwater level. The land form is in part responsible for the dryness or wetness of certain areas within a region that has essentially the same overall climate. If the topographic factor predominates, the so-called a-zonal soils result, e.g. hydromorphic soils of swamps or seepage areas, and halomorphic (saline) soils of poorly drained regions.

biologic agents

Both plants and animals have a profound influence on soil development. The decomposition of dead leaves and roots of trees, shrubs, and herbs provides organic colloids (humus) and humic acids, which exert their influence on the process of soil leaching, especially in the cool temperate climate (podzolization). The action of burrowing animals (prairie dogs, marmots, worms, and termites) can mix the soil thoroughly and can prevent the formation of differentiated soil horizons by the vertical translocation of soil components.

time

In only slightly sloping areas of humid tropical regions, where high rainfall and high temperatures produce intensive weathering and leaching, the time factor is of predominant importance in the formation of soils. Soils under tropical forests are therefore usually chemically poor soils. In recent alluvial deposits the influence of climate and time is not yet noticeable and consequently there is no soil profile development. Time, however, is a passive factor; it is only important if there is a parent material that can be changed by weathering and if water is available to keep weathering processes active.

2.1.4 THE SOIL PROFILE

The vertical section through the soil is called the soil profile. The pedologist, the agronomist, and the hydrologist all look upon it from a different point of view. As a result, the investigations that each will conduct into the profile will differ as well.

the pedological soil profile

The pedological soil profile can be subdivided into layers approximately parallel to the soil surface. The pedogenetic soil layers are called horizons. Not all distinct soil layers are horizons, because soil layers can also be geogenetic, e.g. water or wind deposits. The morphology of the soil expressed in the pedogenetic profile reflects the combined effect of the relative intensities of the soil-forming factors responsible for its development.
The upper and most weathered part of the pedological soil profile, comprising both the A- and B-horizons, is called "solum". Its thickness varies from half a meter to several meters (Fig.1).

the agrological soil profile

The agrological soil profile coincides with the rootzone of crops, which, for field crops, is usually limited to the upper 1.20 meter. For this reason, soil surveys are also generally limited to this depth. The agrological soil profile consists of two main layers: the topsoil (sometimes called surface soil or plough layer) and the subsoil. The topsoil usually coincides with the pedological A-horizon. The subsoil, in its agronomic sense, is the part of the root zone below the plough layer. It should be noted that the term "subsoil" as used by drainage specialists usually refers to the soil strata below the drains. Irrigation and drainage engineers are interested in the water intake rate of the topsoil, whereas the agronomist is interested in its workability (ease of cultivation), structural stability (surface sealing or crusting hazards, erosion hazards), and particularly in its fertility. In contrast with the subsoil, the agricultural qualities of the topsoil can be strongly influenced by soil management and weather conditions. For irrigation purposes, the water-holding capacity in relation to the effective soil depth (see Sect. 2.4.2) has to be known, whilst for drainage it is the water-transmitting properties of the subsoil that are important.

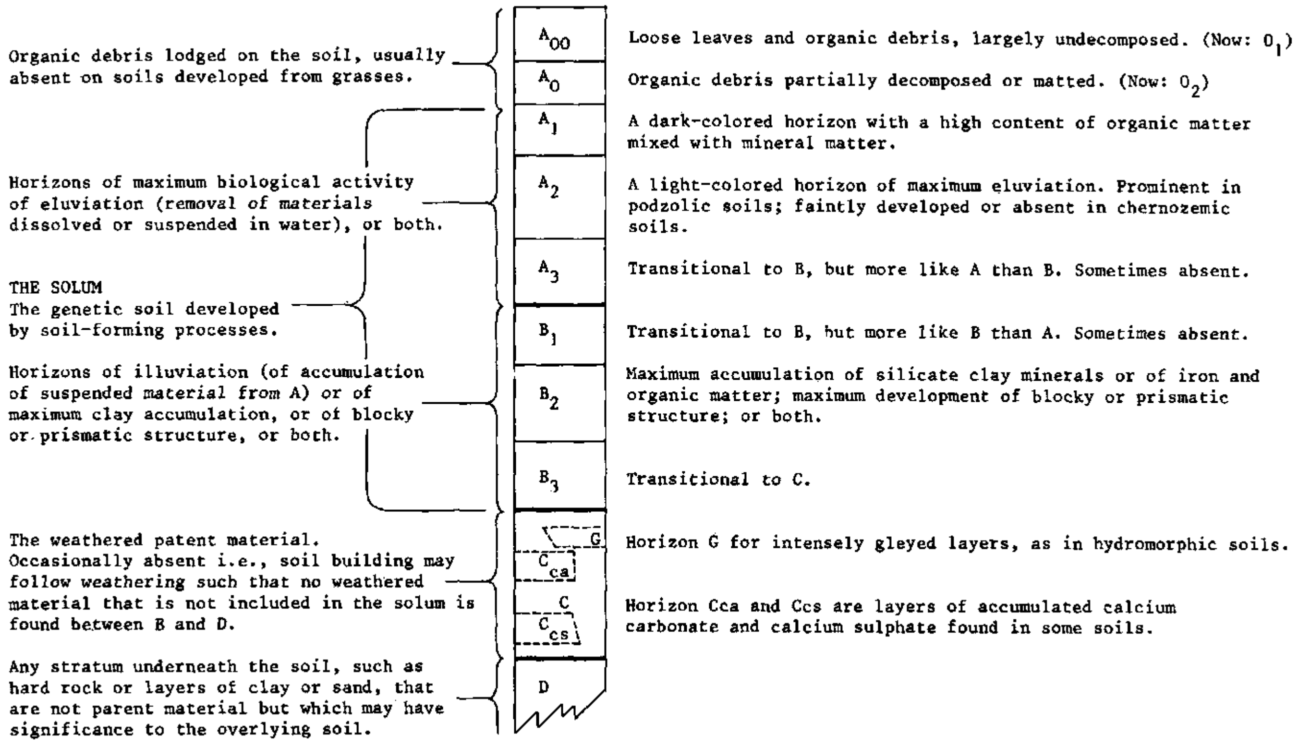

Fig.1. A hypothetical soil profile having all the principal horizons. It will be noted that horizon B may or may not have an accumulation of clay. Horizons designated as C_{ca} usually appear between B_3 and C. The G may appear directly beneath the A (Soil Survey Manual, 1951).

the hydrological soil profile

The hydrological soil profile comprises the root zone of the agrological profile, plus the substrata down to an impervious layer. The water-transmitting properties of these substrata are of considerable importance to the drainage specialist.

2.1.5 OUTLINE OF NORMAL SOIL INVESTIGATIONS

For each soil layer, or for selected soil samples, investigations are carried out either in the field, in the laboratory, or both, their purpose being to obtain information on the following main soil characteristics:

a. Physical

Field or morphological investigations:
colour, including mottling; texture (estimated); structure (shape, size and stability of aggregates); compactness; cementation; actual moisture conditions; consistence (wet: plasticity and stickiness; moist: friability; dry: hardness); concretions (carbonate, iron, manganese); visible salts; other special features, such as surface crusts, cracks, coatings, slicken sides, clay skins, degree of ripening, variability of the soil profile over relatively short distances, etc.

Laboratory investigations:
mechanical analysis (texture);
bulk density and particle density;
moisture retention curve, permeability, aggregate stability.

b. Chemical

Field investigations:
free carbonates (effervescence with HCl), pH (field test with indicators).

Laboratory investigations:
plant-nutrient contents (N, P, K, Ca, Mg etc., trace-elements),
pH, total soluble salt, and gypsum.

c. Physico-chemical

Laboratory investigations:
cation exchange capacity, exchangeable cations, base saturation, potassium and phosphorus fixation.

d. Mineralogical

Laboratory investigations:

clay mineral identification, mineral nutrient reserve.

e. Biological

Field investigations:

organic matter (nature and distribution), root distribution, macro-fauna and micro-organisms (bacteria, fungi, etc.).

aerobic and anaerobic conditions.

Laboratory investigations:

organic matter content, C/N ratio.

The following sections may provide some logic and order to this outline.

2.1.6 RELATIONSHIPS BETWEEN BASIC SOIL CHARACTERISTICS, PHYSICAL SOIL PROPERTIES, AND AGRICULTURAL QUALITIES

Since the basic soil characteristics result from the interactions of the soil forming factors, an experienced soil scientist can predict these characteristics once he knows the factors sufficiently well (Fig.2). We shall call these basic soil characteristics A-factors. They comprise:

- soil texture, especially the clay content of the soil,
- mineralogical composition of the clay fraction : ratio of alumino-silicates and sesquioxydes, swelling or non-swelling types of clay minerals,
- the physico-chemical characteristics of the clay fraction : kind and quantity of adsorbed ions,
- the organic matter : kind and quantity,
- the free carbonate content of the topsoil.

These A-factors, in turn, interact and find their expression in another set of physical soil properties in accordance with the processes operating in soils and the laws of behaviour of soil materials. When we call this second set (first derivatives) of physical soil properties B-factors, it is clear that a knowledgeable soil scientist can infer them to a greater or lesser extent from a known combination of A-factors, or even directly from data on the five soil forming factors.

As the B-factors can be regarded as first deratives of the A-factors, one must rank them as primary physical soil properties. The B-factors are (see Fig.2):

- soil structure, which comprises:
  . aggregate formation (size, shape, distinctness),
  . porosity (total porosity and pore size distribution),
  . structural stability,
  . structural profile.
- soil consistence under wet, moist,and dry conditions,
- soil colour.

The given main relationships can be schematized as follows.

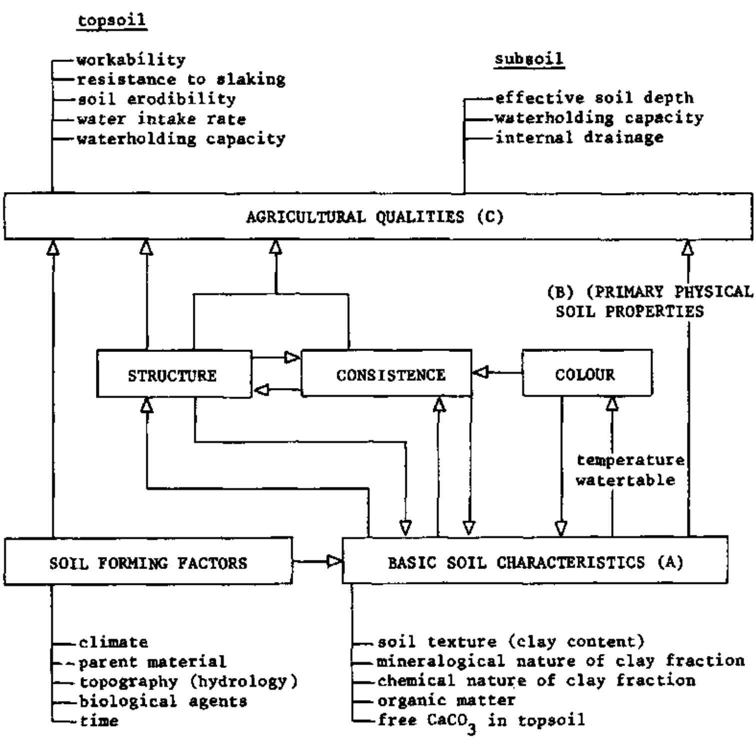

Fig.2. Relationships between basic soil characteristics, physical soil properties and agricultural qualities.

There is a close relationship between these three B-factors. For instance, a dark coloured or grayish soil with a coarse prismatic or platy structure is usually hard when dry and sticky when wet, and a red tropical soil with a granular or subangular-blocky structure is usually friable when dry and only slightly sticky when wet.

A third set of soil properties are the physical agricultural qualities of soils (C-factors), which can be regarded as properties derived from the B-factors, or as second derivatives of the A-factors. In analogy to the above, these properties can be ranked as secondary soil properties. The C-factors are (see also Sect. 2.1.5):

a. concerning the topsoil:
   . workability,
   . resistance to slaking (destruction of aggregates),
   . surface sealing or soil crusting,
   . soil erodibility,
   . water intake rate,
   . water-holding capacity.

b. concerning the subsoil:
   . effective soil depth or root zone depth,
   . water-holding capacity,
   . internal drainage.

These agricultural qualities can be measured on small plots in the field or on soil samples representing such individual sites. The broader knowledge of A- and B-factors and how they are interrelated must be used, in combination with a soil survey, to ensure that the sampling areas selected are representative of the land.

2.2 BASIC SOIL CHARACTERISTICS

2.2.1 SOIL TEXTURE

The mineral soil elements can be classified according to their size. The size distribution of the ultimate soil particles is referred to as "texture". It can be estimated in the field or determined in the laboratory.

particle size limit

There are various textural classifications, but the most commonly used for agronomic purposes is that of the U.S. Dept. of Agriculture. The main particle size limits are given in Table 1.

Table 1. Particle size limits.

Soil	Separate	Diameter	Limits (mm)
SAND		2.00 - 0.050	
	very coarse		2.00 - 1.00
	coarse		1.00 - 0.50
	medium		0.50 - 0.25
	fine		0.25 - 0.10
	very fine		0.100- 0.050
SILT		0.050- 0.002	
	coarse		0.050- 0.020
	fine		0.020- 0.002
CLAY		< 0.002	

Material larger than 7.5 cm in diameter is usually called a stone. Material between 7.5 cm and 2 mm is termed gravel, while material smaller than 2 mm is referred to as fine earth.

Coarser soil particles are separated by sieving. The U.S. Standard sieves are indicated by a number, which refers to the number of openings per inch (LAMBE, 1951) as shown in Table 2. The French textural classification is given in Table 3.

Table 2. Sizes of U.S. Standard sieves.

Sieve No.	Diameter (mm) of opening
10	2.00
18	1.00
20	0.84
35	0.50
60	0.25
70	0.20
140	0.105
200	0.074
300	0.050
400	0.037

Table 3. French particle size limits.

Soil	Separate	Diameter	Limits (micron)
SABLE		2000-50	
	sable grossier		2000-200
	sable fin		200- 50
LIMON (silt)		50-20	
	limon grossier ou sable très fin		50- 20
	limon fine		20- 2
ARGILE (clay)		< 2	

Material > 2 mm is called "elements grossiers".

textural classes

The relative proportion of sand, silt, and clay in a soil determines its textural class. The number of possible combinations is obviously infinite.

For practical purposes, however, certain arbitrary divisions are made, and a descriptive name is applied to all particle-size compositions within each arbitrary division. Fig.3 shows the textural classification currently used by the U.S.D.A. with the Dutch classification superimposed on it.

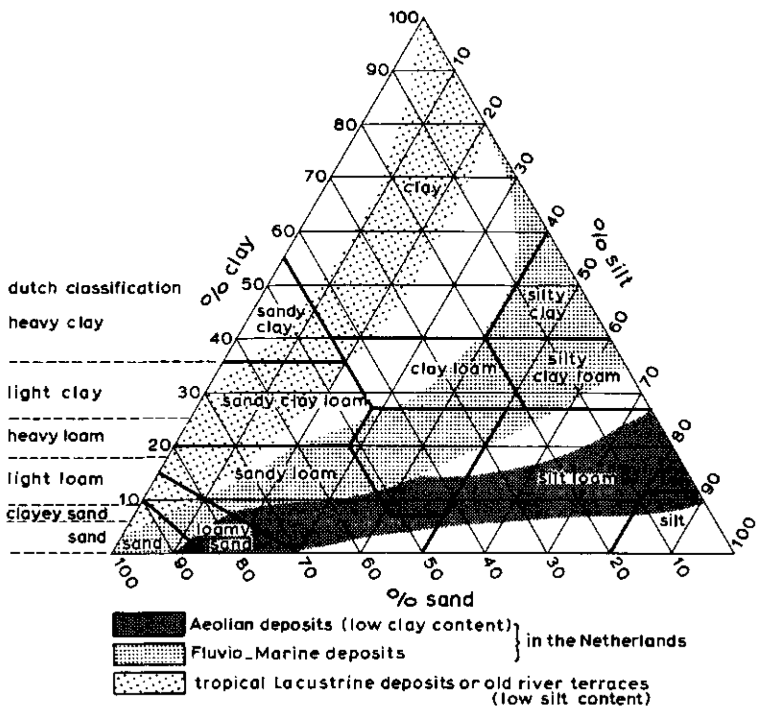

Fig.3. Textural classification.

The textural variation in young alluvial areas in The Netherlands is also indicated in the figure.

In soil survey work, the textural class is estimated in the field. The soil is rubbed between thumb and fingers and its "feel" is noted. The following soil textural classes can be distinguished:

SAND	: gritty, loose, without cohesion whether moist or dry.
SANDY LOAM	: very gritty, some cohesion because of colloidal material.
LOAM	: characteristics of grittiness predominant, but particles stick.
SILT LOAM	: smooth and floury.

CLAY LOAM : slightly gritty, plastic, some tendency to shine if rubbed when moist or cut when dry. Dry lumps can be crushed with some difficulty.

SILTY CLAY LOAM : smooth and floury with little grit, very plastic, noticeable shine if rubbed when moist or cut when dry. Dry lumps can be crushed between fingers, but with difficulty.

CLAY : stiff, plastic, without grit, even when a small sample is bitten between the teeth; tendency to shine strongly when rubbed. Dry lumps have a polished surface when cut, cannot be crushed between fingers (SOIL SURVEY MANUAL, 1951).

textural grouping

The texture of the surface soil is a characteristic closely associated with the workability of the soil (ploughing, seedbed preparation). It also has a bearing on erodibility, the water intake rate, and the formation of soil crusts and soil cracks. The 15 or more recognized textural classes may be grouped into 7 or even fewer groups for purposes of farm planning, soil conservation, etc.

The following terms and groupings are commonly used:

Textural term	Alternative term	Textural classes to be included
1. very heavy	very fine textured	heavy clay (more than 60% clay)
2. heavy	fine "	clay, silty clay, sandy clay
3. moderately heavy	mod. fine "	silty clay loam, clay loam, sandy clay loam
4. medium	medium "	silt loam, loam,very fine sandy loam
5. moderately light	mod.coarse "	fine sandy loam, sandy loam
6. light	coarse "	loamy fine sand, loamy sand
7. very light	very coarse "	sand, coarse sand

Depending on the purposes of grouping and on the prevailing local conditions, a broader grouping of the textural classes may be desirable, e.g. (1+2), 3, 4, 5, (6+7) or (1+2), (3+4+5), (6+7).

light and heavy soils

Sand, when dominant, yields a coarse textured soil which is called "sandy" or "light", since such a soil is easily worked. On the other hand a fine textured soil is made up largely of silt and clay, with corresponding plasticity and stickiness, which implies that the soil is likely to be difficult to work or "heavy". Therefore, the use of the terms "light" and "heavy" refers to ease of working and not to soil weight.

texture, soil permeability and water retention

There is a strong relationship between permeability, water retention, and texture. The heavier the soil, the more restricted the permeability and the higher the water retention (Fig.4). This relation, however, is modified by such factors as the nature of the clay fraction, the coarseness of the sand fraction, and the soil structure. In addition, and especially in alluvial soils, the mode of formation (fluviatile or lacustrine) has a certain influence on soil permeability: lacustrine heavy clays may have a higher permeability than coarse sands due to a specific process of structure development after drainage (see Chap.32, Vol.IV).

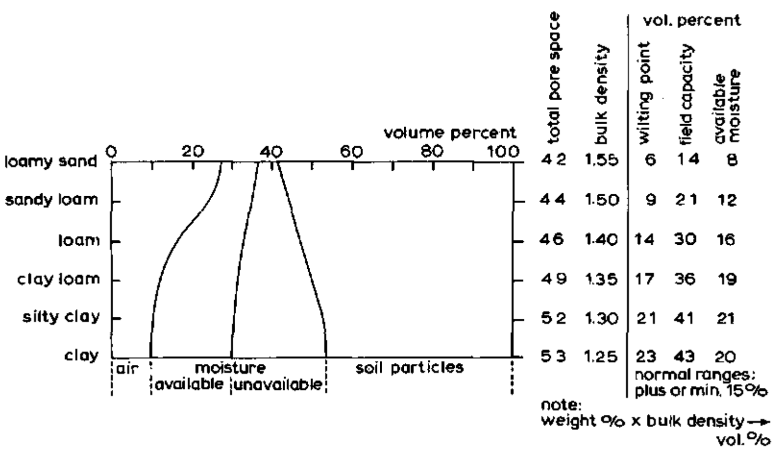

Fig.4. Average physical values and textural relationships.

2.2.2 MINERALOGICAL COMPOSITION

For the present discussion it is useful to distinguish two main groups of minerals according to their size: the minerals of the silt and sand fraction and the minerals of the clay fraction.

the silt and sand fraction

These two fractions have about the same mineralogical composition.
The identification of these mineral components enables the origin of the soil, its stage of weathering, and its mineral reserve (plant nutrient) to be determined. For this purpose the sand fraction is separated into two fractions, the so-called light fraction (specific gravity < 2.9) and the heavy fraction (specific gravity > 2.9). The light fraction consists mainly of unweatherable quartz (SiO_2) and weatherable feldspars (Na, K, Ca alumino-silicates). Strongly weathered soils contain no, or very few, feldspars.
The heavy fraction usually makes up 1 to 2 percent of the total sand fraction. The weatherable minerals are Fe, Mg, Ca-silicates (augite, hornblende muscovite, biotite), Fe-oxydes (ilmenite, magnetite), calcium carbonate (calcite), etc. In addition, various kinds of unweatherable minerals (zircon, tourmaline, etc.) occur which are useful for diagnostic purposes.
The silt and sand fractions of most soils, therefore, consist of light coloured minerals, mainly quartz and some feldspars, and very few dark-coloured Fe- and Mg-minerals.

the clay fraction

The properties of this important fraction vary from one soil to another, depending on the relative importance of kind and size of the various mainly inorganic components. In most soils the mineral components consist largely of crystalline hydrous alumino-silicates, but in strongly weathered reddish tropical soils the greater part of the clay fraction consists of crystalline and non-crystalline iron and aluminium oxydes and hydroxydes (sesquioxides).

The general structural scheme (lattice) on the layered alumino-silicates is generated by a combination of two types of structural units or sheets:

- sheets of silicon oxyde: one atom silicon is surrounded by four oxygens,
- sheets of aluminium oxyde and hydroxyde: six oxygens or OH-groups surround a larger atom like aluminium.

A combination of one silicate sheet and one aluminium sheet generates a 1:1 type clay mineral (e.g. kaolinite, halloysite, etc.). When a second Si-sheet is added, a 2:1 type clay mineral results. In this arrangement two Si-sheets sandwich a central Al-sheet.

The 2:1 type clay minerals can be divided into two main groups with corresponding differences in physical and physico-chemical properties:

- clay minerals which expand on wetting: the smectite or montmorillonite group and the vermiculite group. These silicates contain Mg, Ca, and Fe, whilst water and certain organic molecules can enter between the unit layers, causing the lattice to expand, generating large swelling pressures. Conversely, when the water is withdrawn there is considerable shrinkage.
- clay minerals of the non-expanding type. The main groups are: illite, micas, and chlorite. Here potassium bonds the sheets together.

Clays containing minerals with elongate or fibrous shapes such as halloysite (1:1) and attapulgite (2:1), are more permeable than clays composed mainly of platy minerals.
The smectites are usually associated with the most plastic and less permeable clays; illites are intermediate, while the kaolinites confer the least plastic properties and a higher permeability.
Acid conditions leading to the removal of silica and bases combined with a good internal drainage favour the formation of kaolinite (e.g. in the highly weathered tropical soils). Enrichment with Ca, Mg, and Si due to lateral inflow, poor drainage, and a pronounced dry season are generally favourable for montmorillonite formation. Conditions favouring the formation of the mica type clay minerals (illites) are less defined. Generally, the illites are intermediate weathering products derived from very similar minerals existing in the parent rock. In many parts of the world the clay minerals of "young" soils are almost identical with the clay minerals of the parent material and are not a reflection of the external environment.

2.2.3 PHYSICO-CHEMICAL CHARACTERISTICS OF CLAYS

For practical purposes, the clay fraction can be considered to coincide with the mineral colloid fraction. Properties which are conferred upon a soil by the colloid component are shrinkage and swelling, flocculation and dispersion, plasticity and cohesion.
Clays have pronounced physico-chemical properties because of the combined influence of two factors: the high specific surface and the electric charge on the basic silicate structure of clay minerals. The specific surface is the surface area per unit weight, and it ranges from 15 m^2/g for the coarsest non-expanding clays, to 800 m^2/g for the finest, expanding clays. The high specific

surface results from the small particle size and the platy or fibrous-elongate morphology of the minerals. The electric charge results from ion substitutions in the crystal structure (e.g. Al^{3+} for Si^{4+}; Mg^{2+}, Fe^{2+} for Al^{3+}), or by ionization of hydroxyl groups in water. Although some positive charges occur, especially on the edges of the crystals, the dominant charge associated with the flat surface of the crystal lattice is negative.

Cation Exchange Capacity

Clay particles are characterized by a so-called ionic double layer: the solid, negative charged clay particle is surrounded by a diffuse layer of positively charged cations (Ca^{++}, Mg^{++}, Na^{+}, K^{+}, NH_4^{+}, H^{+}, Al^{+++}) in the liquid phase. The concentration of these cations near the clay particles is much higher than the concentration in the soil solution (Gouy-Chapman theory for the diffuse double layer). Moreover, these cations cannot be leached out so easily, although they can be replaced by other cations. Therefore, the cation exchange capacity of the soil particles acts as a kind of temporary storehouse for the bases that are either released from the primary minerals by weathering or added by fertilizers. Associated with the diffuse layer of adsorbed cations are a large number of water molecules. As the clay minerals weather and become more inert, the cation adsorption capacity declines, and the anion adsorption capacity (phosphates etc.) increases.

The cation exchange capacity (CEC) of soils varies not only with the kind and percentage of clay, but also with the content of humus. Clay and humus together constitute the so-called adsorption complex of the soil. The CEC is expressed in milligram equivalent per 100 gram of soil (meq. or me/100 g). Its order of magnitude for humus and the most important clay minerals is given in Table 4.

Table 4. Cation exchange capacity of some soil materials.

Material	CEC (me/100 g)
humus	200
vermiculite	150
montmorillonite	100
illite	30
kaolinite	10
quartz (silt size)	< 0.01

It is obvious from Table 4 that the exchange capacity of the organic colloids is much higher than that of the inorganic.

proportion of the different exchangeable cations

The chemical force with which the bases are held diminishes in the sequence of Ca > Mg > K > Na and therefore these ions tend to accumulate in the same order in a soil subjected to leaching. In most soils Ca^{+2} will constitute about 80 percent of the total exchangeable bases. In coastal areas Mg^{+2} may dominate. In highly weathered tropical areas H^+ and Al^{+++} dominate. Some salt-affected soils in arid areas or soils flooded with seawater have a high amount of exchangeable Na^+. If these soils contain a low amount of soluble salts, the sodium ions cause dispersion of clays, and soils will become puddling and acquire a slow permeability. Ca-ions, on the other hand, cause flocculation of clay and therefore promote a good structure and a good permeability.

If 100 gram of soil has a CEC of 20 meq and 12 meq exchangeable bases, the soil is said to have a base saturation of 60%.

The pH indicates the relative amount of H^+ and OH^- in solution and it also reflects the percentage of base saturation. The pH of the more or less leached soils in humid regions is lower than the pH in arid regions. If the percentage base saturation is 90 or 100, the pH of the soil is about 7, or higher if carbonates, especially those of sodium, are present.

2.2.4 ORGANIC MATTER AND SOIL FAUNA

kind and quantity

When fresh organic matter is incorporated into the soil, part of it is rapidly decomposed by the action of micro-organisms. A slowly decomposable residue, called humus, remains, which consists of a mixture of brown or dark amorphous and colloidal substances. The term micro-organism, or microbes, is used in literature to include both micro flora (e.g. bacteria and fungi) and microscopic animal life (e.g. protozoa and nematodes). The rate of decomposition and the kind and quantity of the end products formed depends on temperature, aeration, chemical soil conditions, and the type of micro-organisms.
The amount of organic matter in mineral soils varies widely. Most soils have an organic matter content of 2-4 percent; a content of less than 1 percent (arid

regions) is considered low.

The main sources of supply under farming conditions are:

- organic residues of crops: stubble and especially the root residues,
- farm manure (animal excrements),
- compost,
- green manure (crops ploughed under when in an immature, succulent stage, especially legumes).

importance of organic matter

The influence of organic matter on the physical and chemical properties of soils is already great when present in small quantities. The beneficial function of organic matter may be summarized as follows.

Physically, organic matter, more than any other single factor, promotes the formation and the stability of soil aggregates. The decomposition of fresh organic matter, in particular, produces microbial germs and mycelia of organisms, which are most effective in aggregating soils. Aggregation leads to increased porosity, which means increased aeration, infiltration and percolation of water, and a reduction in runoff and erosion hazards. Furthermore, the improved soil structure and the high water-adsorption capacity of humus increase the moisture retention capacity of the soil. In a chemical sense, the decomposition of organic matter yields N, P, and S, and promotes the extraction of plant nutrients from minerals through the formation of organic and inorganic acid. Moreover, there can be a considerable fixation of nitrogen from the air by non-symbiotic bacteria, which obtain their energy from decomposing dead plant tissue, and by symbiotic bacteria, which get it from the cell sap of such legumes as alfalfa, clovers, peas, and beans. Finally, the humus component of organic matter significantly increases the CEC of the whole soil because the CEC of the humus component is two to thirty times the CEC of mineral colloids. For additional details the reader is referred to the relevant chapters in RUSSELL (1954) and BEAR (1964).

the soil macro-fauna

Besides the various micro-organisms, the soil - especially if well supplied with fresh organic matter - contains a large number of animals, such as rodents, insectivora, insects, millipeds, mites, spiders and earthworms. Most of these animals use more or less undecomposed plant tissues (from litter and dead roots) as food. They thus serve to incorporate much organic matter into soils and to initiate the decomposition processes that are continued by the micro-organisms. Futhermore, considerable quantities of soil are mixed, transported, and granulated, especially by worms, and the holes left by the various animals serve to increase the aeration and internal drainage of the soil.

2.3 PHYSICAL PROPERTIES OF MINERAL SOILS

The combined influence of inorganic and organic components, plus the prevailing chemical conditions in the soil (A-factors), determine soil porosity and structure (see Fig.2). The A-factors, in combination with the quantity of water present, determine soil consistence, while, together with the prevailing regime of soil aeration and temperature, they determine soil colour.

2.3.1 SOIL POROSITY

Here, only the total pore space will be considered; the pore size distribution will be discussed in Sect. 2.3.2. To calculate the pore space of soils, the particle density and the bulk density have to be known.

particle density (specific gravity)

The particle density is the mass per unit volume of soil particle. It is usually expressed in grams per cm^3 of soil, but one may find it expressed in pounds per cubic foot (pcf). Note that 100 pfc corresponds with 1.6 g/cm^3. Instead of particle density, the term specific gravity (s.g.) is often used. It is normally defined as the ratio of the weight of a single soil particle to the weight of a volume of water equal to the volume of the particle. It is a dimensionless quantity. Since one cm^3 of water weighs one gram at normal temperature, the two terms have the same numerical value. The particle density is sometimes referred to as true density.
The average specific gravity of some soil components is: organic matter 1.47; sand 2.66; clay 2.75. The specific gravity of mineral soils usually varies from 2.6 to 2.9 - 2.65 being considered a fair average.

bulk density

The bulk density (volume weight or apparent density) is the dry weight of a unit volume of soil in its field condition, or in other words, it is the mass of a dry soil per unit bulk volume, the latter being determined before drying. It is expressed in grams per cm^3. It has the same numerical value as the apparent specific gravity, which is defined as the ratio of a unit bulk volume of soil to the weight of an equal volume of water. The bulk density of uncultivated soils usually varies between 1.0 and 1.6; compact layers may have a bulk density of 1.7 or 1.8. Generally, the finer the texture of the soil and the higher the organic matter content, the smaller the bulk density.

Note: moisture content on weight basis × bulk density = moisture content on volume basis.

The pore space of a soil is that portion of unit bulk volume which is occupied by air and/or water (volume of voids). The volume of pore space depends largely on the arrangement of the solid particles. The porosity, n, is the volume percentage of the unit bulk volume not occupied by the solid particles, i.e.

$$n = 100\ (1 - \frac{\text{bulk density}}{\text{particle density}}).$$

Example: if the bulk density is 1.4 and the particle density 2.65, the porosity equals $100\ (1 - \frac{1.4}{2.65} = 47\%)$.

Usually the porosity of mineral soils varies between 35% for compacted soils and 60% for loose topsoils.

The specific volume of the solid phase is the volume occupied by one gram of solids, i.e. the reciprocal value of the particle density. An average value is $1/2.65 = 0.38\ cm^3/gram$.

The specific volume of the soil is the bulk volume occupied by 1 gram of soil, i.e. the reciprocal value of the apparent density. (The latter is often called shortly the specific volume).

The shrinkage of soils can be computed directly from the specific volume. For example, a sediment which originally has a specific volume of $0.95\ cm^3/gram$ and, after drainage, of 0.72, shrinks 23 percent, or in other words, a layer of 10 cm shrinks 7.7 cm after drainage.

In soil engineering, the void ratio, e, is often used instead of the porosity. It is defined as the ratio of the volume of voids to the volume of solids,

$$e = \frac{V_v}{V_s} = \frac{V_v}{1 - V_v}$$

Note that the denominator in the expression for void ratio, e, remains constant when the soil as a whole changes its volume.
A porosity of 35% corresponds with a void ratio of e = 0.35/0.65 = 0.54.
A porosity of 60% corresponds with a void ratio of 1.5. Peats and mucks may have void ratios as high as 4 or 5.
The reduction in void ratio due to water leaving the soil pores is termed consolidation. The reduction in void ratio due to air being forced out of the soil by mechanical means is termed compaction.

2.3.2 SOIL STRUCTURE

The term "soil structure" refers to the 3-dimensional arrangement of primary soil particles (sand, silt, clay) and/or secondary soil particles (micro-aggregates) into a certain structural pattern (macro-aggregates or peds). The aggregates of textural elements are held together by colloids (mineral and organic) and separated from one another by cracks and large pores.
In a soil without structure the primary soil grains would be arranged in a more or less random fashion, approximating a dense packing such as might occur in a mixture of spheres of different sizes. It would have no systematic pattern of planes of weakness along which it could be broken apart to form distinct peds.
Structure is an important morphological characteristic of the soil. As such it is not a plant-growth factor in itself, but it influences almost all plant-growth factors such as water retention, water movement, soil aeration, root penetration, micro-biological activities, resistance to soil erosion, etc.
In a structured soil the size and shape of aggregates governs the pattern and spacing of cracks and macro-pores, including the total surface area of aggregates. Movement of water takes place principally through major cracks and large pores, and the water that is most readily available to plants is stored in the macro-pores between and within the aggregates. These are also the spaces which plant roots explore most intensively for water and nutrients.
There are four main aspects of soil structure:

- The visible macrostructure, based on field investigation and described in terms of shape, size of the aggregates, and the grade (distinctness of individual aggregates) of structure.
- The spaces in between and within the macro- and micro-aggregates or the total pore space and the pore size distribution.
- The structure stability, especially of the topsoil or plough layer.
- The structure profile, or the kind, thickness, and sequence of the various

structural horizons or layers.

macro structures

Macro structures may be divided into:
- Simple structures, coherent or non-coherent, in which natural cleavage planes are absent or indistinct (structureless):
  - . single grain, common in loose sands and silts with low organic matter content (e.g. beach sands, recent volcanic ash), and
  - . massive, common in sandy loams, loamy sands, silt loams, etc. The soil clings together because of small amounts of clay and organic matter, but shows no preferred and pre-existing lines of cleavage.
- Aggregate structure in which natural cleavage planes are distinct. An individual aggregate is called a ped (in contrast to a "clod" caused by disturbance such as ploughing or digging, and a "fragment" caused by rupture of the soil mass across natural surfaces of weakness).

There are four primary types of structure, which are based on the relative length of vertical/horizontal axes and the contour of the edges (Fig.5):
- Platy: Horizontal dimensions greater than vertical dimensions. Horizontal cleavage plane dominate (medium class: 2-10 mm).
- Prismatic: Vertically elongated aggregates in the shape of prisms. Prisms rounded at the top are designated columnar (medium class: 20-55 mm).
- Blocky: About the same horizontal and vertical dimensions (medium class: 10-20 mm).
  Angular blocky: faces flattened, edges sharp.
  Subangular blocky: mixed rounded and flattened faces, some rounded edges.
- Granular: More or less rounded granules uniform in shape and size. The term "crumb" indicates a granular aggregation which is more porous and more irregular in size and shape (medium class: 2-5 mm).

Further details and illustrations can be found in the SOIL SURVEY MANUAL (1951). In literature on soils one may come across such references as "weakly structured", "well-structured soil", "strongly developed soil structure", "good structure" or "bad structure", etc. The first three refer to the grade of structure development and indicate whether or not the aggregates are distinct and retain their size and shape upon disturbance of the soil. The last two refer to qualities relevant to crop production. Good structure means that there are many stable small aggregates which optimize water movement, storage, and aeration, and increase the surface area of aggregates upon which exchange and uptake of nutrients takes place. For example, a weak, fine granular structure indicates a risk of the structure being destroyed by cultivation. A strong, coarse prismatic structure is agriculturally bad. A well-developed fine crumb structure, on the other hand, is a very favourable type for crops.

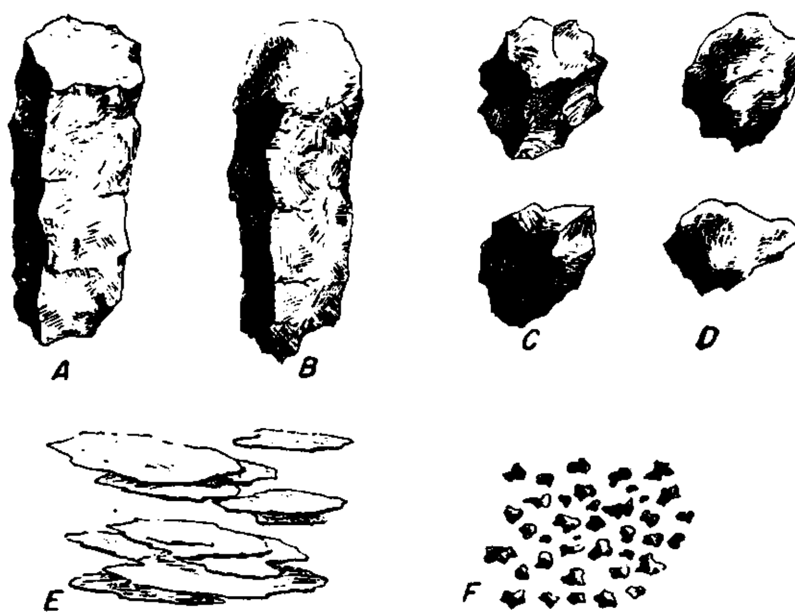

Fig.5. Drawings illustrating some of the types of soil structure:
A. prismatic, B. columnar, C. angular blocky, D. subangular blocky, E. platy, F. granular. (Soil Survey Manual, 1951).

pore size distribution

When the water-transmitting and water-holding properties of the soil are being considered, it is the pore size distribution rather than the total pore space that is important.

There are two main genetic types of soil pores: firstly the inter-aggregate spaces, which are pores resulting from the aggregation of the soil particles, and secondly, the biopores resulting from plant growth (rootlets and root-hairs)

and soil fauna (worms, insects, etc.). Accordingly, pores vary in quantity, size, shape, and continuity.
With respect to the size and the function of the pores, the following distinctions can be made:

Descriptive term	Size	Main function
macro pores	100 microns	aeration and drainage (gravity flow)
meso pores	30-100 „	water conduction (rapid capillary flow)
micro pores	30-3 „	water retention (slow capillary flow)

The macro pores are visible to the naked eye. Other names for them are aeration pores and non-capillary pores.
The meso pores are visible at 10 x magnification.
The micro pores are not visible, but their presence can be deduced from the faces of the aggregates. When these aggregates have a rough surface, there are many micro pores.
Water is present not only in the pores but also on the surface of the soil particles (hygroscopic and film water).

structural stability

Aggregates vary greatly in the extent to which they can withstand the impact of raindrops, the flow of irrigation or runoff water, and water-logged conditions. Plant growth (seedling emergence and stand), aeration, runoff and erosion all depend greatly on the structural stability of the topsoil. Such stability is mainly determined by the organic matter content of the soil (quantity and kind), the silt and very fine sand content in relation to the clay content, the chemical components associated with the clay (Ca^{+2}, Na^{+}) and mineral cementing materials (iron-, aluminium-,silicium-oxydes and hydroxydes).
A soil is likely to be structurally unstable when it has a low content of organic matter, a high content of silt and fine sand, and a moderately high clay content. Such soils are often greyish or yellowish in colour. The process by which a dry soil mass disintegrates upon wetting is called "slaking" or the soil is said to "run together". On drying, this results in the formation of a surface crust (surface sealing). If mechanical forces (raindrops, trampling by cattle, ploughing of wet soils), cause the soil aggregates to break down the term "puddled" is used. Rice fields are usually puddled on purpose; this destroys the macro pores and creates a dense, more or less impermeable layer, which prevents excessive losses of water.

the structural profile

The structural profile refers to the kind, thickness and sequence of various soil structures. The water-holding and water-transmitting properties of a soil profile consisting of sand over clay or of clay over sand are quite different. The position of a dense layer (bulk density 1.6 to 1.8) within the soil is of particular importance in matters of water percolation, aeration, and root penetration. Such layers, or an abrupt change in texture and structure (pronounced stratification), largely determine the effective soil depth or root zone depth.

2.3.3 SOIL CONSISTENCE

Soil consistence refers to the manifestation of the physical forces of cohesion and adhesion within the soil at various moisture contents (dry, moist, wet), as evidenced by the behaviour of that soil toward mechanical stress or gravity. Soil consistence will decide the duration of the period suitable for ploughing, the required traction force, and whether it is easy or difficult to prepare a good seedbed.
Indirectly, the consistence can provide an experienced field surveyor with a great deal of useful information on soil texture, structure, and permeability. The phenomena of soil consistence and friability, plasticity, and stickiness, as well as resistance to compression (suitability for foundation), and shear (side slope of cavity). The last two phenomena belong in the field of soil mechanics and will not be discussed here.

Friability characterizes the ease of crumbling of moist soils. Descriptive terms are: loose, very friable, friable, firm, very firm, extremely firm.
Dry soil consistence is described in terms of: loose, soft, slightly hard, hard, very hard, extremely hard.

Plasticity refers to capacity of the wet soil, within a certain range of moisture contents, to change its form when subjected to outside forces and to retain this new form (capacity for moulding). Soils containing less than 15 to 20 percent of clay are generally non-plastic. The engineering classification of soil is based on texture (coarse grained, fine grained) and plasticity.
For this classification two consistence limits (Atterberg limits) are defined: the liquid limit and the plastic limit.
The liquid limit is the minimum water content at which a soil-water mixture

changes from a viscous liquid to a plastic solid. It has about the same consistency as the saturated soil paste used for salinity investigations. The liquid limit is determined by placing a soil sample in a standard machine, separating the sample into two halves with a standard grooving tool. If the groove in the soil-water mixture closes under the impact of 25 standard blows, the mixture is at its liquid limit.

The plastic limit is the lower water content boundary of the plastic range for a soil. It is arbitrarily defined in the laboratory as the smallest water content at which the soil can be rolled into a 3 mm diameter thread without crumbling.

The plasticity index or plastic number (liquid limit minus plastic limit) defines the range of moisture contents at which the soil has the properties of a plastic solid.

At a moisture content above the plastic limit the soil will puddle if handled or worked, which means that soils should be ploughed at a moisture content below the plastic limit (LAMBE, 1961).

Stickiness refers to the degree of adhesion of the wet soil material to other objects. It is determined by noting its adherence to the skin when pressed between thumb and finger.

Descriptive terms are: non-sticky, slightly sticky, sticky, and very sticky.

Another aspect of soil consistence is soil compaction, which denotes a combination of firm or hard consistence and close packing of particles, resulting in a low porosity. It is measured by the resistance to penetration of the moist soil. This is in contrast to cementation, which refers to a hard, brittle consistence which does not soften appreciably on prolonged moistening.

Consistence and structure are closely related. Whilst soil structure is the result of the forces in the solid phase, consistence is an indirect measure of these forces.

This implies that the structure can sometimes be inferred from the consistence and that the converse is also true.

2.3.4 SOIL COLOUR

Colour is the most obvious and most easily determined soil characteristic. When the colour of a soil is considered in conjunction with other observable features - structure, texture and consistence - a great deal can be inferred as to the soil's physical and chemical conditions.

causes of soil colours

Colour is dependent on the nature of the parent material from which the soil has been formed, on internal and external drainage, and on the prevailing soil temperatures.
Colour is due primarily to coatings on the surface of mineral particles. In aerated soils the colours can be dark brown, almost black when humus particles predominate, or they can be yellow to red due to coatings of more or less hydrated iron compounds. In waterlogged soils grey-greenish colours occur, due to the reduction of ferric iron to ferrous iron. This condition is indicated by the term "gley".
A horizon may be uniform in colour or it may be mottled (marked with spots). The term "gley mottling" refers to the occurrence of patches or red, yellow and other colours due to oxidation after a period of reduction, under a regime of a fluctuating water table (temporary waterlogged conditions). Another important diagnostic factor which may indicate a temporary waterlogged condition is the colour of the soil in the vicinity of the roots. When the channels of living roots are characterized by lighter colours than the surrounding soil mass (e.g. in a brown soil the roots are outlined in grey or green) and the channels of dead roots are outlined in rusty yellow and browns, this may be taken as the criterion of impeded aeration.
Pronounced red and yellow colours are generally associated with humid tropical or sub-tropical soils, whereas in arid climates greyish-yellow colours prevail, indicating little chemical weathering and low organic matter content.

colour description

In soil survey reports the colours are described using the Munsell colour system, which is a colour designation system specifying the relative degrees of the three simple variables of colour: hue, value, and chroma.
Hue is a quality that distinguishes one colour from another. Main hues are yellow (Y), red (R), green (G), blue (B), and purple (P).
The detailed soil hues in the range red to yellow are: 10 R, 2.5 YR, 7.5 YR, 10 YR, 2.5 Y and 5 Y.

Value is a measure of the lightness or darkness of any colour, Value 1 being very dark and Value 8 very light.

Chroma is a measure of the strength or weakness of a colour, Chroma 8 being very strong and Chroma 1 very weak.

The Munsell notation should always be accompanied by a verbal description of the colour, for example: 10 YR 6/4 (light yellowish brown).

2.4 SOIL MOISTURE

The soil-water relations can be divided into two main categories:
- the water-transmitting properties of the soil
- the water-retaining properties of the soil.

2.4.1 WATER-TRANSMITTING PROPERTIES

The rate of water movement is governed by gravity or capillary forces, or both, and by the soil permeability.

soil permeability

The term soil permeability is used in a general, quantitative sense and means the readiness with which a soil conducts or transmits water. To define soil permeability more precisely, a distinction is made between:
- the surface intake rate, which determines the relation between water absorption and runoff,
- the sub-surface percolation rate, which determines the internal soil profile drainage, and
- the hydraulic conductivity, which is the proportionality factor in Darcy's law (see Chap.6). This soil characteristic is of particular importance to sub-surface drainage flow (Fig.6).

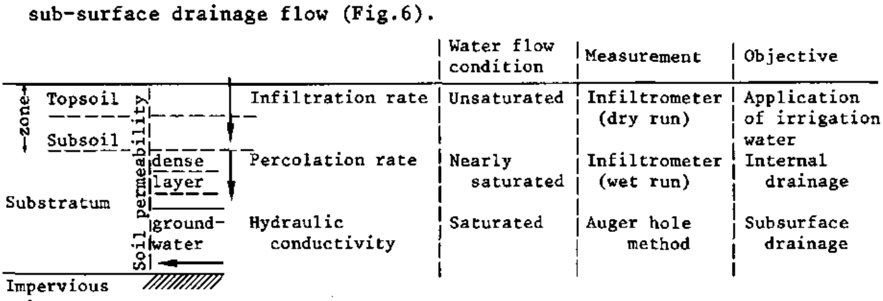

Fig.6. Three aspects of soil permeability.

The intake and percolation rates both refer to vertical permeabilities under unsaturated conditions. The two terms, however, are not synonymous. The intake (or infiltration rate) refers specifically to the entry of water into the soil surface (i.e. transmission plus storage), whereas percolation rate refers to the water movement through the soil and may be defined as the quantity of water passing through a unit area of cross section per unit time at a given depth in the soil mass.

internal drainage of the rootzone

The term internal drainage refers to that property of the rootzone that permits excess water to flow through it in a downward direction. Poor internal drainage is revealed when infiltrating water becomes stagnant in the rootzone, e.g. on a poorly pervious layer. Whether such a layer will generate a so-called perched water table, i.e. a water table forming the boundary of a saturated zone below which an unsaturated soil layer is found, depends not only on the percolation rate of this layer, but also on the rate at which surface water infiltrates, and on the soil storage capacity.

2.4.2 TOTAL READILY AVAILABLE MOISTURE

The capacity of a soil to retain moisture that is readily available for plant growth is an important factor in land use planning. This applies not only where there is adequate rainfall, but also in irrigation projects, where irrigation water has to be applied at the right time and in the right quantity.
Available Moisture (AM) is the moisture-holding capacity of a given undisturbed soil sample between field capacity (upper limit) and wilting point (lower limit) expressed in volume percentage. It is a physical characteristic of a given soil layer.
The wilting point, also called permanent wilting point, is the moisture content at which most plant roots are no longer capable of taking up water from the soil, and the plants suffer irreversible wilting. Experience has shown that the moisture tension at wilting often equals about - 15 atm. Hence the fifteen atmosphere percentage, which means the percentage of water contained in a soil that has been saturated and subsequently equilibrated with an applied pressure of 15 atm in a pressure membrane apparatus, is often used in preference to the wilting point. This measurement is much easier to apply and is more reproducible than when wilting point is determined by a series of experiments on live plants.
The field capacity is the percentage of water remaining in a soil two or three

days after having been saturated and when free drainage has practically ceased. Experience has shown that for many deep, homogeneous and freely draining loams the moisture tension at field capacity equals about - 1/3 atm. Hence the moisture content of a soil sample equilibrated at 1/3 atm pressure is commonly used instead of the field capacity.
There is often a good relationship between soil texture and AM-values, particularly when the soils have the same clay mineralogy, adsorbed ions, structure, etc. (see Fig.4).

Total Available Moisture (TAM) is the sum of the AM-values for each layer of the actual or potential rooting depth of the soil profile, i.e. the effective soil depth.
The term effective soil depth refers to the depth of soil which plant roots can readily penetrate in search of water and plant nutrients. The character of any layer limiting the effective depth will also affect the internal drainage of the soil. Limiting layers are: compact or distinctly indurated layers, bed rock, gravel, coarse sand or any abrupt and pronounced discontinuity within the profile.

Not all the total available water (AM x effective root depth) may be considered readily available. A rule of thumb is that the TRAM-value, Total Readily Available Moisture, is about two-thirds of the TAM-value. As an example: when the effective soil depth = 1.20 m and AM = 10%, then TAM = 12 cm and TRAM = 8 cm.
In sandy soils the available water is generally so low as to be a major problem.

2.5 SOIL AIR

Plant roots and most soil micro-organisms utilize oxygen (O_2) from the soil air, and give off or respire carbon dioxide (CO_2). A continuous supply of oxygen is needed for this respiration process. An insufficient supply will limit plant growth, particularly in medium to fine textured soils in a humid climate and in irrigated soils. Improving the soil aeration is one of the main objectives of drainage.

2.5.1 COMPOSITION OF SOIL AIR

The pore space of a soil - about 40-50 percent by volume - is occupied by water and gases. A small proportion of the gases is dissolved in the soil water and the rest constitutes the soil air.
When comparing soil air and atmospheric air, the nitrogen content of both is about the same (79 percent), the oxygen content of soil air is lower (atmosphere 20.97%) and its content of carbon dioxide and water vapor is higher (CO_2 atmosphere: 0.03%).
The CO_2-content of soil air is usually between 0.2 - 0.5%, but this may increase to over 1%, and may even become as high as 15%. There is a general inverse relation between O_2 and CO_2 contents: when O_2 decreases, CO_2 increases. With some notable exceptions, the sum of the soil air's CO_2 and O_2 is very near that of the atmospheric air.
In places where gas interchanges are prevented, as happens under waterlogged conditions or where anaerobic biological activities predominate, such products as methane (CH_4) and hydrogen sulfide (H_2S) may accumulate.
The composition of soil air shows a marked seasonal variation. It represents a dynamic equilibrium between two competing processes: the production of CO_2 (respiration of roots and microbes) and its removal.

2.5.2 VOLUME OF SOIL AIR

There is an inverse relationship between soil-air and soil-water. An excessive amount of water implies a shortage of soil air.
Of special importance is the air content of the soil one or two days after heavy rainfall or irrigation, when most of the gravity water has been removed. The air-filled pore space under these conditions is often referred to as aeration porosity, as aeration capacity, or as non-capillary porosity. It can be defined in terms of soil water tension, pore size diameter, or volume percentage.
As a rule of thumb, it can be said that a soil is well aerated if it has an aeration porosity of 10% on a volume basis.

2.5.3 RATE OF OXYGEN SUPPLY

Two distinctive mechanisms are involved in the interchange of gases between the soil and the atmosphere: diffusion and mass flow (convection). Diffusion is the most important means by which the soil air is renewed. In this process the individual gases move in response to their own partial pressure differences or gradients. Due to the respiration processes of roots and microbes, the partial pressure of oxygen is reduced below that of the atmosphere, resulting in a movement of atmospheric oxygen, whilst the partial pressure of the CO_2 rises above its normal atmospheric content, resulting in an outward movement of CO_2. Diffusion must take place through air-filled pores, since air cannot diffuse readily through a water layer. The rate of diffusion is determined by the total volume, and especially by the continuity, of the air-filled pores. The sizes of the pores have little effect on the rate of diffusion, whereas they are of major importance for the water-transmitting properties of the soil. Experience has shown that a compacted surface soil layer or a soil crust has a strongly negative influence on the aeration of the soil, especially under wet conditions and high temperatures.

Mass flow results when the flow of gases into and out of the soil is a consequence of the gradients in the total pressure between the soil air and the atmosphere. Pressure differences of this type arise mainly from differences in temperature and barometric pressure. Compared with diffusion, the mass flow is a minor factor in soil aeration. For mass flow the size of pores is decisive, the rate of mass flow being proportional to a power of the pore size.

2.5.4 PLANT REQUIREMENTS

A prerequisite for vigorous plant growth is an ample supply of oxygen in the rootzone. A plant's aeration requirements, however, and its tolerance for poor aeration conditions, vary considerably. Its stage of growth can also be significant. There is a lack of information on the exact aeration requirements of different plants and of quantitative data specifying the aeration status of soils. Practical experience has established only the relative need for aeration. Plants with high oxygen needs are tomatoes, potatoes, sugarbeets, peas, and barley.

Poor aeration conditions impede a plant's uptake of water (physiological drought) and nutrients, and curtails its growth of roots.

2.5.5 AERATION CONDITION AND SOIL PROCESSES

In many ways soil aeration also exerts an indirect influence on plant growth through the effect it has on the soil's biological processes and chemical conditions.

Nitrogen fixation by aerobic microbes is of great importance in soil and is strongly influenced by aeration. The lack of sufficient air prevents the oxydation of nitrogen and sulfur into forms that plants can readily utilize. The amounts of soluble iron and manganese are as strongly influenced by the oxygen concentration of the soil air, as they are by the pH of the soil. When anaerobic processes are periodically replaced by aerobic reactions, iron and manganese may accumulate in the soil in the form of concretion. Under anaerobic conditions both inorganic and organic toxic substances may develop.
In general, a high CO_2 content increases the solubility of phosphorus and calcium carbonate. The latter is of importance in the reclamation of calcareous sodic soils.

2.5.6 SOIL AERATION AND DRAINAGE

The principal objective of subsurface drainage is to promote favourable soil-water-air relations, and in this a distinction should be made between crop-drainage and soil-drainage.
The purpose of crop-drainage is to promote an aerated rootzone during the growing season of the crop.
During the time that there are no crops on the land, the so-called soil-drainage is required. Soil-drainage has two aims:

- to maintain the soil's structure, temperature and nitrogen supply in a state favourable to future plant growth,
- to maintain soil trafficability and workability (ploughing, seedbed preparation).

2.6 SOIL TEMPERATURE

Along with water, air, and nutrients, another important growth factor for plants is the temperature of the soil. Microbiological activity, seed germination, and root growth are all greatly affected by the soil temperature.

2.6.1 SOIL TEMPERATURE AND PLANT GROWTH

The process of germination depends on the temperature of the soil rather than that of the air. The temperature favourable to seed germination varies with the species of plants. The required minimum daily soil temperature in the top 5 cm is about 10° C for alfalfa, 16° C for corn, and 22° C for cotton.

The subsoil temperatures are particularly important for root growth in early spring. Soils with well-drained subsoils warm up more quickly and to a greater depth than do soils with a higher water content, hence the importance of good drainage in early spring in temperate or mediterranean climates.
Microbiological activity is very restricted below a temperature of 10^{o} C. Above 10^{o} C the activity increases greatly, with a corresponding increase in the availability of nitrogen, phosphorus, and sulfur, brought about by the decomposition of fresh organic matter.

2.6.2 SOIL TEMPERATURE AND DRAINAGE

Wet soils have a higher heat capacity (thermal capacity or specific heat) than do dry ones.
The specific heat of any substance is defined as the number of calories of heat required to raise the temperature of one gram of that substance by 1^{o} C.
The specific heat of water is 1.00 cal/g. The specific heat of a dry mineral soil is usually about 0.20 cal/g. To compare the heat capacity of dry and wet soils, it is preferable to use the volumetric heat capacity (cal/cm^3). One cm^3 of dry soil with a pore space of 50% has a heat capacity of 0.5 x 265 x 0.2 = 0.26 cal/cm^3. Such a soil, when all its pores are filled with water, would have a heat capacity of 0.26 + (0.5 x 1.0) = 0.76 cal/cm^3. If only half the pore space were filled with water, the heat capacity would be: 0.26 + (0.25 x 1.0) = 0.51 cal/cm^3. Furthermore, if excess water does not percolate through the soil, most of it will be removed by evaporation, which has a pronounced cooling effect.
The temperatures of poorly drained soils are 4^{o} to 8^{o} lower than those of comparable well-drained soils.
Besides the specific heat of the soil, the heat conduction should also be taken into account, but this does not alter the general conclusion that badly drained soils are cold soils.

2.7 SOIL FERTILITY AND PRODUCTIVITY

2.7.1 DEFINITIONS

The term "fertility" is used in various senses. It may refer to:

a. The inherent capacity of the soil to supply nutrients to plants in adequate amounts and in suitable proportions. Other terms used in this sense are "chemical fertility" and "nutrient status". The nutrient-supplying capacity of the soil can be divided into

(i) the actual fertility, being the nutrient supply per unit of time, otherwise called the short-term nutrient-supplying capacity, and

(ii) the potential fertility, which is a function of the total reserves of plant nutrients in the soil (weatherable minerals , organic matter, exchangeable bases) and/or the soil's response to the use of fertilizers.

b. The ability of the soil to yield crops. In this sense fertility is a function of both chemical and physical soil properties (soil-water and soil-air relationships).

Crop yields, however, depend not only on soil conditions, but also on the prevailing climatic conditions and on farm practices as regards soil, water and crop management (use of fertilizers, control of erosion, weeds, pests, drainage, irrigation, variety used, etc.). These aspects are covered by the term "productive capacity" of the soil, which refers to the yields of crops adapted to that particular soil and climate under a given set of management practices.
In land classification the term "soil productivity" is used. This term means the productive capacity, expressed in terms of produce.
Under certain conditions a soil can have a high productive capacity but a rather low productivity. Ultimately, it is only the soil productivity that counts. It is measured in terms of output (yields) in relation to inputs (water, fertilizer, insecticides, machines, etc.) for a specific kind of soil under a defined system of management.

2.7.2 THE SUPPLY OF NUTRIENTS

The major elements absorbed by plant roots from the soil are nitrogen (N), phosphorus (P), potassium (K), calcium (Ca), magnesium (Mg), and certain trace elements, i.e. elements which are needed in minute quantities, such as iron (Fe), manganese (Mn), copper (Cu), zinc (Zn), boron (B), molybdenum (Mo), cobalt (Co).
The elements may occur in the soil in three main forms: in the minerals (temporarily unavailable), in the exchangeable form, and in the bulk of the solution.

Nitrogen is fixed from the atmosphere through bacterial action (nitrogen fixation) or is released from the organic matter in the soil after decomposition (nitrification). Phosphorus is derived from decomposition of organic matter and weathering of certain minerals. The other elements are primarily supplied by the weathering of inorganic solid phase. N, P and K are commonly supplied to the soil as commercial fertilizers and/or as farm manure.

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INTRODUCTORY SUBJECTS

3. SALTY SOILS

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PURPOSE AND SCOPE

Brief account of the problems of salty soils, their origin, occurrence, and reclamation. The drainage of salty soils is discussed in detail in Chapters 9 and 11 (Vol.II).

CONTENTS

		page
3.1	Origin and occurrence	77
3.2	Salt types and their distribution	78
3.3	Effects of salinity on crops and soils	79
3.4	Classification	80
3.5	Reclamation	84
Literature		89

3.1 ORIGIN AND OCCURRENCE

All soils, even those in humid areas, contain some soluble salts. These are usually calcium salts and their concentration is often no more than 0.4 g per litre soil moisture. The salt content of soils in arid zones, though usually higher than in humid areas, can still be considered low when compared with actual, salty, soils. Salty soils show high contents of various kinds of salts and/or a high percentage of exchangeable sodium. Heavily salinized soils may even show efflorescences or complete salt-crusts, formed by such salts as gypsum ($CaSO_4$), common salt (NaCl), soda (Na_2CO_3), or more complex salts.
Some salty soils came into being because the parent material was itself salty. Others became salinized by being flooded with sea water, by wind-borne salt spray or dust, by irrigation with water that contained salt or that was contaminated by saline industrial waste waters. The majority of salty soils, however, have developed as a result of the upward capillary flow of water exceeding its downward movement.
A considerable capillary transport of groundwater to the surface is only to be expected when water tables are high for prolonged periods of time. Such a situation is often found in irrigated areas which have inadequate drainage. High phreatic levels also occur in regions where the groundwater reservoir is fed from natural sources, which means that salinization is caused by the evaporation of water that fell on another place. Hence, salty soils are mostly found in or near depressions and valleys in arid or semi-arid regions (Fig.1).

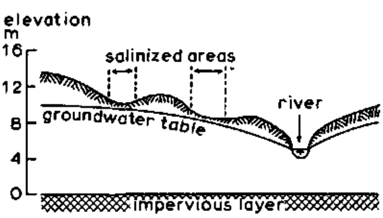

Fig.1. Relation between depth of groundwater and salinization.

The extent of capillary salinization and the depth at which salts accumulate are governed by the rate of capillary rise and the salinity of the groundwater, counteracted by the leaching intensity (by rain or irrigation water). The rate of water transport to the soil surface depends on the depth of the groundwater table, on the potential gradient between groundwater and soil surface, and on the capillary conductivity of the soil in relation to the moisture content

(Chap.5). The reduction of the salt content in the soil brought about by irrigation water or rain depends on the quantity and quality of water percolating through the soil, on the physical characteristics of the soil,and on its moisture content.

3.2 SALT TYPES AND THEIR DISTRIBUTION

Salty soils vary considerably in their salt content, their types of salts, their structure, and their reclamability.

Dominant anions are chlorides, sulphates, and carbonates, sometimes nitrates. Sodium salts occur most frequently, but calcium and magnesium compounds are common too, while mixtures of various salts and complex minerals are not exceptional. Fig.2 shows a typical pattern of the various soluble components found in a saline sodic soils.

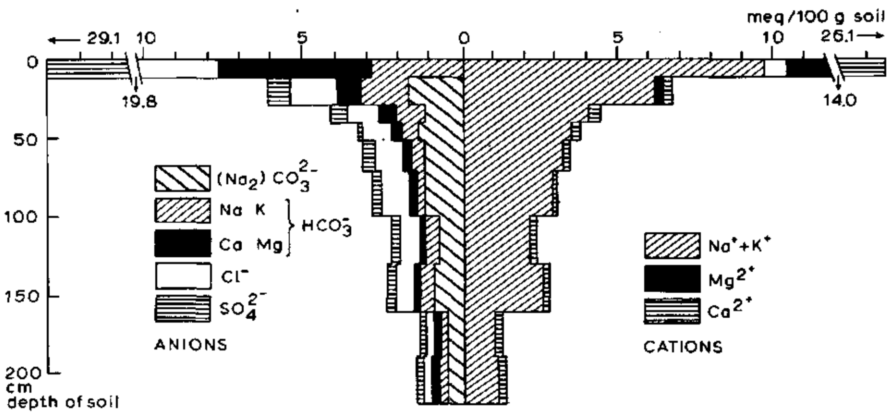

Fig.2. Distribution of various soluble salts in the upper 2 m of a saline sodic soil. Analysis performed in 1:5 extract. (Field guide of the Excursion of the Symposium on Sodic Soils (not published) 1964).

As mentioned before, the non-saline soil solution usually contains mainly calcium salts. A relation exists between the cations in the soil solution and those bound, in an exchangeable form, to the clay particles. In normal soils calcium generally forms 80% or more of the exchangeable cations. Magnesium, potassium and sodium make up the major part of the remaining exchangeable cations, sodium remaining below 5% (often even below 1%) of the total cations.

The soil solution in saline soils is not only much more concentrated, but the kinds of salts are different from those in non-saline soils. That means that the cations adsorbed at the surfaces of the clay particles also show another composition. The percentage of exchangeable calcium is lower and the values of potassium, magnesium and, in particular, sodium are higher.

Characteristics of salty soils is the non-uniform salt distribution. Within short distances the salt content may vary greatly, owing to slight differences in level, soil composition, permeability, plant growth, etc. (Fig.3). The vegetation on salty soils often shows a strikingly patchy growth. The wide variation in salinity, both horizontally and vertically, greatly hampers adequate sampling.

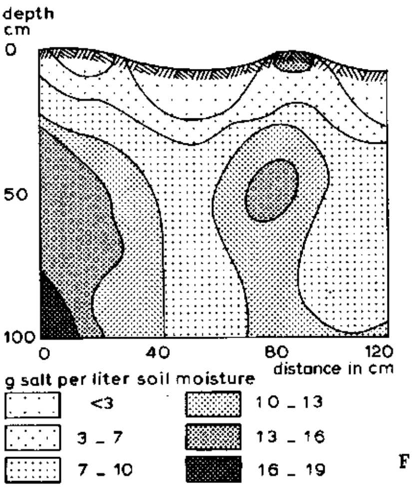

Fig.3. Differences between salt figures over short distances.

3.3 EFFECTS OF SALINITY ON CROPS AND SOILS

Salt affect crops through specific toxic ions, though the toxic effect is often less significant than that of the increased osmotic pressure of the soil solution, which results in a reduction of the plants' capacity to withdraw water from the soil. There may be an indirect adverse effect on crops caused by the unfavourable structure of salty soils. The characteristics of clay soils (shrinkage, swelling, pore space distribution, total pore space, structural stability) are influenced substantially by the strength of attraction between the clay particles. This attraction depends largely upon the exchangeable cation composition. Divalent and trivalent cations (Ca, Mg; Al) are more strongly attracted to clay particles than are monovalent cations (Na, K) and allow these particles to condense into stable larger aggregates, which can result in better structured agricultural soils. A low salt concentration, coupled with a predominance of sodium among the exchangeable cations, causes deterioration of the structure of soils that contain significant amounts of clay. A high salt concentration in the soil solution compresses the layer of adsorbed cations and hence physical qualities of the soil are good. These effects can be predicted from the Gouy-Chapman Diffuse Double Layer Theory for exchangeable cations. This theory describes the thickness of the mantle of bound water in which the adsorbed exchangeable ions are distributed around the clay particles. After the leaching of the excess of salts, the clay particles of a sodium soil disperse; fine particles may be washed down to the subsoil where they form an impervious layer; the swelling on wetting becomes more pronounced; permeability for air and water is greatly reduced; crust formation is favoured; the soils are sticky when wet and hard when dry; they become unsuitable for cultivation and are hardly fit for plant growth. In soils containing sodium carbonate, organic matter may go into solution, colouring the surface of the soil black upon evaporation of the soil moisture. Many soils with high exchangeable magnesium values show bad structures too.

3.4 CLASSIFICATION

Many local names are used to identify salty soils and their characteristics: Reh, Usar (India), Sabbagh (Iraq), Tir (Morocco), Brak (S.Africa) and Szik (Hungary). The term white and black alkali have been widely used in the USA. Internationally known and much used are the Russian names Solonchak and Solonetz. White alkali and Solonchak refer to soils containing an excess of soluble salts, usually visibly accumulated at the ground surface. Black alkali and Solonetz soils contain an excess of exchangeable sodium in the absence of considerable amounts of soluble salts. The Russian classification is, in part, based on soil profile development. Thus, as a result of some leaching, by rainfall, a Solonetz shows a compact, prismatic or columnar, clay-enriched subsoil layer (B-horizon). Probably the most practical classification is the one used by the US Salinity Laboratory (RICHARDS, ed., 1954). This classification is based on two charac-

teristics: the salinity of the soil (i.e. the amount or concentration of water-soluble salts in the soil) and the exchangeable sodium-percentage. The soil salinity is the predominating factor for plant growth, whereas the exchangeable sodium level determines the possible decline of structure.
As a parameter for the soil salinity the electrical conductivity of the saturation extract (EC_e) is used, expressed in mmhos, mho being the reciprocal value of ohm. The saturation extract is the solution extracted from a saturated, disturbed soil paste which is a glistening mixture of soil and water generally just sliding freely from a spatula. It has a water content of about the liquid limit. To calculate the exchangeable sodium percentage (ESP), the amount of exchangeable sodium (ES) and the cation-exchange capacity (CEC) have to be determined:

$$ESP = \frac{100 \times ES}{CEC}$$

The ESP can also be estimated from the amounts (Ca^{++} + Mg^{++}), Na^+ and K^+ found in the saturation extract (Fig.4).

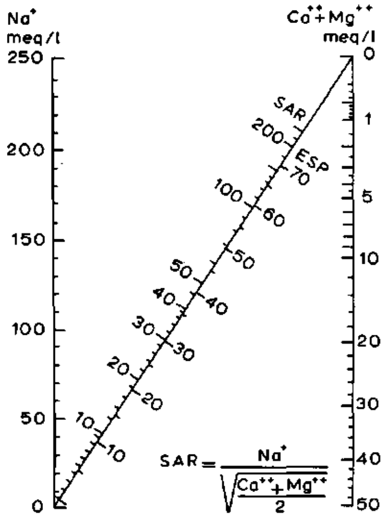

Fig.4. Nomogram for determining the SAR value of a saturation extract and for estimating the corresponding ESP value of soil at equilibrium with the extract. (According to Richards, ed., 1954).

Based on EC_e and ESP values, the classification is as follows:

<u>saline soils</u>

(a) EC_e higher than 4 mmhos/cm at 25° C

(b) ESP lower than 15

(c) pH generally below 8.5

Such quantities of soluble salts adversely affect the majority of crops. White salt crusts may be found on the surface of these soils. The chief anions are Cl^- and SO_4^{--} and to a lesser extent HCO_3^- and NO_3^-. Insoluble carbonates and sulphates may be present. Na^+, as a rule, comprises less than 50% of the soluble cations.

saline-sodic soils

(a) EC_e higher than 4 mmhos/cm at 25°C

(b) ESP higer than 15

(c) pH seldom higher than 8.5

Crop growth on these soils is seriously impeded. Their structural condition is usually fair, but may deteriorate considerably upon leaching. The soil may then become strongly alkaline; soil particles will disperse; the permeability will diminish markedly, and its suitability for tillage will reduce.

non-saline sodic soils

(a) EC_e lower than 4 mmhos/cm at 25°C

(b) ESP higher than 15

(c) pH usually between 8.5 and 10, but lime-free soils pH-values as low as 6 may occur.

In general, the important anions are Gl^-, SO_4^{--} and HCO_3^- but carbonates too are often present. Na^+ is the main cation in the soil solution, Ca^{++} and Mg^{++} being largely precipitated. The structure of non-saline sodic soils is usually very bad.

discussion

The above classification has the advantage that it is simple and is based on characteristic quantities. However, it does not completely cover the wide variety in field conditions (no classification could!) and should thus be handled judiciously. For this reason, the choice and use of the parameters EC_e and ESP will be briefly discussed.

Relationships between EC_e-values in the root zone and crop growth do exist

(Table 1), but these relations are not strict because the salt tolerance of crops is also dependent on weather conditions, moisture regime of the soil (irrigation), the kind of salts present, and the extent to which certain salts predominate. For a proper use of the data obtained, depth and time of sampling should always be noted in view of the variability of salinity with depth and time.

Table 1. Relation between salt content in the root zone and plant growth (medium textured soils)

Crop response	EC_e	Salt Content (% on a dry matter basis)
Salinity effects negligible	0 - 2	0.05 - 0.1
Yields of very sensitive crops may be restricted (e.g. beans and most fruit crops)	2 - 4	0.1 - 0.2
Yields of many crops restricted	4 - 8	0.2 - 0.4
Only tolerant crops yield satisfactorily (in lower range cotton, rape, sugar beet, barley, most grasses, and some clovers; in higher range some salt-resistant grasses)	8 - 16	0.4 - 0.8

The ESP value, which is often used in calculating the amount of gypsum required for the reclamation of salty soils, should be handled carefully. The critical value of ESP = 15 does not apply to all soils. Soils high in organic matter may have a far better structure than would be expected from their ESP-values. On the other hand, some soils with a poor Solonetz-like structure show ESP-values far below 15. Moreover, the ratios of exchangeable K to exchangeable Na and of exchangeable Mg to exchangeable Na may influence the effect of the exchangeable sodium on soil properties.

The thickness of the diffuse (double) layer of exchangeable cations depends on its composition and the salt concentration of the soil moisture. Exchangeable Ca-ions reduce the dimensions of the layer whilst Na-ions increase them. Moreover, the thickness decreases with increasing concentration of the soil solution. This layers of exchangeable cations usually produce favourable physical properties whereas, from an agricultural point of view, thick layers provoke an unfavourable behaviour of clay soils. For this reason, saline-sodic soils with a concentrated soil solution do not show specific structural problems (the fine particles of the soil being kept coagulated by the high concentration of electrolytes). After most of the soluble salts are leached, however, the structure of the soil can deteriorate.

3.5 RECLAMATION

The general principle of reclaiming saline soils comprise:

a. the prevention of further salinization,

b. the leaching of salts,

c. the replacement of exchangeable sodium by exchangeable calcium.

Preventing further salinization may require the prevention of flooding, but most times it is a matter of reducing the capillary rise of soil moisture. A lowering of the groundwater table is usually the solution. This process generally requires the construction of a drainage system. Groundwater tables are often lowered to a depth of 1.5 - 2 m below ground surface. The depth required on such factors as climate (a lesser depth is admissible in humid regions), the kind of soil (capillary rise in sand differs from that in clay soils), the quality of groundwater and irrigation water (the better their quality, the higher the groundwater table may be), and the irrigation practices.

In arid zones the leaching of the salts requires irrigation; in semi-arid areas precipitation will sometimes be sufficient to leach the soil, provided that the groundwater table is lowered to an adequate depth; in humid regions rainfall usually frees the soil of salt within a reasonable lapse of time.

Leaching is seldom simply a matter of replacing the saline soil solution by percolating fresh water through the soil. The water moving downwards will mix with the soil moisture. The thinner the layer of water required for complete mixing, the higher the leaching efficiency. This efficiency depends on the moisture content of the soil, the rate of leaching, the pore size of distribution, the spatial arrangement of various pore sizes, and the vertical distribution of the salt through the profile.

During leaching the equilibrium between adsorbed ions and ions in the free soil solution is disturbed and exchanges take place. Ro replace exchangeable Na-ions by Ca-ions, enough Ca-ions must be present in the soil moisture, and leaching is required to wash down the product of the exchange process. The natural concentration of Ca-ions present in the soil solution of calcareous soils is often too low to provide a rapid exchange. Large applications of organic matter, designed to increase the solubility of the lime (production of CO_2), are some-

times successful when used in combination with irrigation. Another technique is to transform the lime in the soil into a more soluble Ca-salt by applying H_2SO_4 or S. The latter will be oxidized by micro-biological action and will then react with the lime in the soil.
A common practice is to apply a fairly soluble Ca-salt, usually gypsum, directly to the soil. Some salty soils already contain gypsum and here leaching alone may suffice.
The water used for leaching saline soils should have a low salt content and a favourable Na/Ca ratio.

reclamation of saline soils

The reclamation of saline soils is largely a matter of leaching excess soluble salts from the root zone. Such leaching, however, will serve no useful purpose if measures are not taken to prevent resalinization after reclamation, i.e. to eliminate the cause of salinization or, at least, to reduce it. This may be done by lowering the groundwater table to a depth deemed necessary and feasible under the local conditions of groundwater, soils, topography and climate.
Great quantities of water (as much as 100 to 150 cm) are often needed for leaching. The leaching itself will create few adverse side-effects since, as observed above, the structure of the soil will nog be affected seriously and permeability does not decrease markedly during leaching. Soluble plant nutrients - particularly nitrates - are lost from the soil along with the excess salts during leaching and sometimes measures have to be taken to restore soil fertility after leaching.
It may be profitable in the first instance to leach only the extent that salt-tolerant crops can be grown, and to continue leaching during and after the first and following crops.
It may be safely be said that a combination of deep drainage and adequate irrigation will be sufficient for the reclamation of most saline soils.

reclamation of saline-sodic soils

Reclamation of these soils is more complicated because special measures have to be taken during leaching to prevent a decline of structure. In an unreclaimed state the physical appearance of these soils closely resembles that of saline soils, but as soon as the bulk of the salt has been leached, the detrimental effect of the high ESP becomes evident through a deterioration of structure.

For saline-sodic soils, too, the prevention of further salinization by lowering groundwater tables is a prerequisite for successful reclamation.

Leaching with irrigation water containing calcium may prevent a decline of structure. Leaching with water of a gradually decreasing salt content, if available, may also help to prevent or reduce collapse of structure. On saline-sodic soils which already have an unfavourable structure before reclamation, rice is often a useful initial crop. Sometimes, the very first step in reclamation is the construction of fish ponds; in this stage some salt may be removed by diffusion.

The gypsum used for restoring soil structure or preventing its decline may be either dissolved in irrigation water or spread over the land. There is no advantage to be gained in psreading more gypsum than can be dissolved in the applied sheet of irrigation water. It is often wise to give an initial dressing that is lower than required amount. It may be that this application increases the permeability to such an extent that natural processes will begin the replacement of exchangeable Na-ions by Ca-ions.
The gypsum requirement can be determined by treating a soil sample with a saturated gypsum solution and measuring the number of Ca-ions which have been used for replacement of other exchangeable cations (except Mg). A rough approach to the amount of gypsum required is given by a formula of the type

$$x_z = \frac{ESP_a - ESP_f}{100} \cdot CEC \cdot y_z$$

where

x_z = the amount of gypsum required per ha for restoring structure in a layer z cm thick

ESP_a = actual exchangeable-sodium-percentage

ESP_f = permissible final exchangeable-sodium-percentage

CEC = cation-exchange-capacity in meq per 100 g of dry soil

y_z = the amount of gypsum required per ha to replace 1 meq of Na per 100 g of dry soil in a layer z cm thick and of specified bulk density.

For a soil with a bulk density of 1.4 the theoretical value of y is 1200 kg of gypsum for a 10 cm layer. In practice more gypsum will be needed because of the non-uniform distribution of the amendment, because part of the $CaSO_4$ will displace other cations, and because some gypsum may be washed to the subsoil.

On the other hand, the supply of Ca-ions by the soil itself has not been included in the calculation.

When amendments other than gypsum are applied, Table 2 (RICHARDS, ed., 1954) may be used for converting amounts of gypsum to quantities of other chemicals.

Table 2. Tons of structure-improving amendments equivalent to one ton of gypsum ($CaSO_4.2\ H_2O$).

Sulphur	0.19
Sulphuric acid	0.61
Iron sulphate ($FeSO_4.\ 7\ H_2O$)	1.71
Aluminium sulphate ($Al_2\ (SO_4)_3.\ 18\ H_2O$)	1.37
Limestone ($CaCO_3$)	0.62

reclamation of non-saline sodic soils

The only measure required with these soils is an improvement of their structure, but this is often hard to achieve. Large quantities of gypsum or other suitable amendments may be needed.
Infiltration of the amendments into the soil is slow due to a poor permeability, and the downward movement of clay particles creates dense and impermeable subsoil horizons. Sometimes deep tillage combined with subsoiling may improve the structure since it loosens impervious strata. Deep ploughing will bring subsoil containing lime and/or gypsum to the top. This may, however, unfavourably affect crop yields for the first year or two.

preventing resalinization

It must be kept in mind that, after having been reclaimed, salty soils will be threatened with resalinization through salts moving into the root zone by upward capillary transport. Low groundwater tables, made possible by deep drainage systems, will keep the capillary rise within limits, but some extra irrigation will be required to wash down the salts brought up between applications of water. The higher the groundwater table for a given soil, the more leaching will be needed. When plenty of irrigation water is available for leaching, even comparatively high groundwater tables may permit an acceptable salt-equilibrium. For the permanent use of land endangered by salinization, leaching intensity and depth of groundwater have to be in harmony.

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INTRODUCTORY SUBJECTS

4. PLANT GROWTH IN RELATION TO DRAINAGE

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PURPOSE AND SCOPE

The relation between plant growth, root development and soil moisture conditions are described in general terms, and, in somewhat more detail, the effects of high water tables and drainage on soil conditions and crop response are discussed.

CONTENTS

		page
4.1	Drainage and agriculture	93
4.2	Plant and soil-water relations	94
	4.2.1 Water and air in the soil	94
	4.2.2 Adaptation of plants to soil moisture conditions	96
4.3	Drainage and physical soil conditions	100
	4.3.1 Soil structure	100
	4.3.2 Soil aeration	101
	4.3.3 Soil organic matter	101
	4.3.4 Soil temperature	102
4.4	Drainage and cultivation practices	103
	4.4.1 Tillage	103
	4.4.2 Weed control	104
4.5	Drainage and nutrient supply	105
4.6	Drainage and salinity	109
4.7	Drainage and diseases or pests	111
4.8	Groundwater level and crop production	112
	4.8.1 Grassland	113
	4.8.2 Arable crops	115
	4.8.3 Fruit trees	118
Literature		120

4.1 DRAINAGE AND AGRICULTURE

Drainage of agricultural land is the natural or artificial removal of excess water from in or on the soil. Water is 'in excess' when the amount present adversely affects the production of crops by reducing the soil volume accessible to roots.
Excessive soil moisture also prevents the carbondioxide formed by plant roots and other organisms from being exchanged with oxygen from the atmosphere, a process known as aeration. Without aeration the root development and uptake capacity for water and nutrients of most plants is reduced. The effect of drainage on soil aeration and consequent root development and plant ecology will be discussed in Sect.2. Drainage also affects physical soil conditions (Sect.3), cultivation practices (Sect.4), nutrient supply (Sect.5), soil salinity or alkalinity (Sect.6) and diseases or pests (Sect.7). In Sect.8 the reactions of various groups of crops (grassland, arable crops, and fruit trees) to the depth of the groundwater table will be reviewed.

Drainage will often result in new areas being brought under cultivation or in the agricultural pattern of an area being changed because conditions have become favourable for a greater or different range of crops. When waterlogged or saline land is reclaimed by drainage, the usual types of monoculture (e.g. extensively exploited grass- or hayland, or, in tropical monsoon areas, a continuous cultivation of rice) will often make way for a wider variety of crops. Most arable crops, e.g. cereals, root crops, fibre crops, and fruit trees, require well-drained soils. Deeply aerated crop lands need regular supplements of organic matter and nitrogen fertilizer. Here, the cultivation of leguminous plants will be helpful. Since leguminous plants may also be valuable as high protein fodder, their introduction into the crop rotation may stimulate livestock farming, leading to a type of mixed farming.

Surface drainage in erosive areas is often accomplished by such methods as grassed waterways or strip cropping with dense-growing cover crops. These methods, too, will move arable (row crop) farming into the direction of mixed farming. Even fish ponds may be part of such farms.
If hydrological, topographical, and soil conditions prevent the drainage of areas with a shallow water table, these areas should be used for crops that can benefit from such conditions. This will mean either grassland (livestock) or

vegetable farming, and in tropical regions the cultivation of rice. (If the surface drainage is sufficient, other crops such as sugar cane can alternate with rice).

To conserve organic soils and to preserve them from shrinkage, water tables are preferably maintained high. On many organic soils, therefore, cultivation is limited to fodder crops, rice, vegetables, or other crops suited to high water-level conditions.

In the drainage of excess water, a distinction can be made between the drainage of:
- the soil surface
- the root zone
- the groundwater.

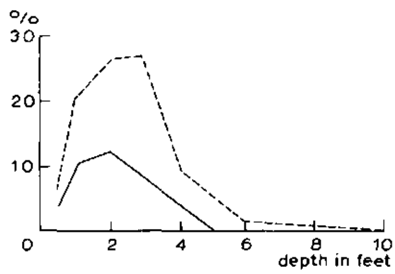
In the absence of shallow groundwater tables, drainage problems on the soil surface or in the root zone will be due on the one hand to high precipitation or irrigation intensities, or on the other to unfavourable soil structures that cause the water to infiltrate or percolate too slowly. Such problems can be solved either by installing surface drainage or by improving the soil conditions through good soil management. Where high groundwater tables exist, drainage can be tackled by evacuating the groundwater through a subsurface drainage system. Drainage and soil conditions often influence one another. For example, a lowering of the groundwater table may result in a better structure of the topsoil, an increased infiltration rate and porosity, and consequently a reduced surface drainage problem.

4.2 PLANT AND SOIL-WATER RELATIONS

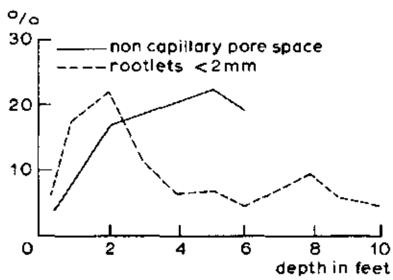
4.2.1 WATER AND AIR IN THE SOIL

Roots require oxygen for respiration and other metabolic activities; they absorb water and dissolved nutrients from the soil, and produce carbondioxide, which has to be exchanged with oxygen from the atmosphere. This aeration process, which takes place by diffusion and mass flow, requires open pore space in the soil. If roots are to develop well, water plus nutrients and air must be available simultaneously. In the root zone, the interstices between soil particles (the pore space) comprise 40 to 60% of the total soil volume. The subterranean plant parts - roots, stolons etc. - and micro-flora and soil fauna grow and develop in these interstices. If the pore space is mainly occupied with water for an appreciable length of time, the soil is said to be waterlogged. Waterlogging is an undesirable condition for most crops, since it causes a deficiency of oxygen. Both soil texture and structure are a means of describing the size of pores (see Chap.2). Capillary and non-capillary pores are distinguished. The capillary pores, which are small, are important for the storage of water for the plant. The non-capillary pores, which are large and readily emptied, function under adequate drainage conditions as channels for the exchange of gases.

With regard to soil moisture content, two important conditions are distinguished: field capacity and permanent wilting point (Chap.2). The amount of water in the soil between field capacity and permanent wilting point indicates the availability of soil moisture for plant growth. Field capacity is considered the upper limit of available soil moisture. It is the amount of water which, under good drainage, is retained against the force of gravity. At field capacity, the capillary pores are filled with water and the non-capillary pores are filled with air. In most soils aeration is sufficient at this point, i.e. the air-filled pore space is sufficiently large for the exchange of gases. In some heavy soils, however, although the pore space may be 60% or more, almost all pores are of capillary dimensions. Such pores remain filled with water and cannot be readily drained. In this case the soil is essentially waterlogged at field capacity (Fig.1). Sandy soils, on the other hand, often have too little capillary porosity. Consequently they do not hold much water against gravity and have a low moisture availability for the plant. Well-distributed rainfall or frequent irrigations will then be necessary to satisfy the water requirements of the crop.
The wilting point is the lower limit of available soil moisture. At this point soil moisture is depleted to such a degree that plants dry out and fail to recover when placed in a dark, humid atmosphere: the plant has wilted.



A) An unfavourable soil. Roots confined to shallow layer because of poor aeration in deeper layer.



B) A favourable soil. Deep aeration permits a more uniform distribution of roots.

Fig.1. Distribution of rootlets (DAUBENMIRE, 1953).

4.2.2. ADAPTATION OF PLANTS TO SOIL MOISTURE CONDITIONS

Plants are differently adapted to the water availability in their environment. Natural vegetation reacts sharply to different soil moisture regimes as shown in Fig.2 (BARON, 1963).

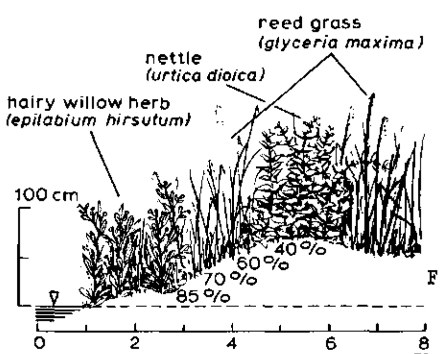
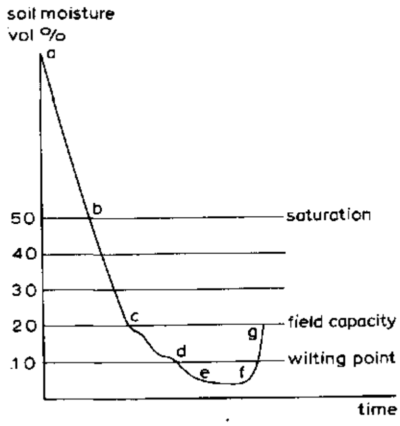


Fig.2. Cross-section showing soil elevation and plant growth in the Itchen valley (Hampshire, U.K.).
(% indicates water content of top-soil)

Accordingly, various ecologic classes of plants are distinguished, e.g. hydrophytes, xerophytes and mesophytes, denoting plants of wet, dry and moist habitats, respectively. The relation of these groups to a soil's moisture content has been demonstrated by WHITE (1956) (Fig.3).



hydrophytes a-b-c
mesophytes c-d
xerophytes c-d-e-f-g
(e-f dormant stage)

Fig.3.
Relation of growth of hydrophytes, mesophytes and xerophytes to moisture conditions in a loam soil.

Hydrophytes include aquatic plants (growing in water) and swamp and bog plants (inhabiting soils saturated with water). These plants develop peculiar internal structures to facilitate the aeration of their roots. Aquatics occur under conditions a-b (in Fig.3); bog plants under b-c. Hydrophytes (e.g. rice) grow under anaerobic soil conditions. During germination they first develop shoots which elongate rapidly and soon pierce through the water surface into the air. During the growth of the shoot, internal air channels (lacunae) develop, making possible the passage of air from shoot tip to base as soon as the shoot has reached the air. From that moment on the root starts to grow, simultaneously developing air-filled intercellular spaces through which oxygen is transported. Hydrophytes often have a shallow root system, or with some species, no root system at all. Roothairs are generally not formed. On the other hand in wet habitats some aquatic plants (e.g. low-land rice) may have a maximum development since mechanical impedance to root growth in the liquid mud is virtually nil.

Xerophytes are plants of relatively dry habitats. By morphological or physiological means these plants escape or endure recurrent drought. Xerophytes often develop an extended, but shallow root system, since in dry regions moisture usually penetrates only into the superficial soil layers.

To the group of mesophytes belong those plant species that cannot inhabit water or wet soil, nor can they survive in habitats where water is significantly depleted (DAUBENMIRE, 1959). The majority of crop plants belong to this group. They generally have a moderately deep root system. Mesophytes and xerophytes occur under conditions indicated by the points c to d in Fig.3. Mesophytes die at moisture contents below point d and xerophytes become semi-dormant or dormant. In arid regions one finds not only xerophytes but also some specially adapted mesophytes: the phreatophytes and ephemerals. Phreatophytes are able to develop deep root systems, with which they tap the phreatic water at greater depths than most other plants!

Ephemerals, in contrast with phreatophytes, have a rather shallow and diminutive root system. Their chief physiological adaptation is their ability to complete their life cycle during the brief rainy season. These plants set seed before the soil dries out. They tolerate atmospheric drought, but not soil drought.

At the onset of germination, seeds absorb large quantities of water. This initiates the further development of the young plant. Unlike hydrophytes, the development of mesophytes and xerophytes almost invariably begins with the root system. In most circumstances water in the immediate environment of the seed is soon exhausted and the development of the roots enables a larger soil volume to be explored for water. If, during their development, the roots meet a waterlogged soil with reduced aeration, their growth will be suppressed. DAUBENMIRE (1959) describes the root development under reduced conditions as follows:

- Roots are shorter and root systems occupy less space and become shallow and sometimes root branches extend upward into the atmosphere.
- Roots may be less numerous, root branching less complex, and roothair formation is usually suppressed.
- Sometimes development of adventitious roots from the base of the stem is stimulated.
- The respiration of the roots changes from aerobic to anaerobic with a consequent accumulation of toxic by-products and a reduced release of energy from the same amount of carbonhydrates.
- The rate of absorption of water and nutrients and the rate of transpiration are reduced.

As a consequence of these adverse soil conditions other plant parts are affected as wel:

- Shoot leaf areas are reduced and leaves are discolored.
- Reproductive processes are delayed and repressed, flowers or young fruit may drop prematurely.

Good aeration and moisture conditions throughout the greater part of the soil profile stimulate growth and development of roots in all directions. The resulting extensive, deep root system explores a large soil volume for water and nutrients. This is enhanced by the intensive contact of roothairs, which are formed more profusely under the stimulus of an adequate external supply of oxygen (ROGERS and HEAD, 1970). In well-drained soils, the deep root system may even advantageously withdraw water from the capillary fringe of the groundwater. Above the groundwater table two zones can be distinguished: a nearly saturated zone and a zone with moisture content near field capacity into which groundwater rises by capillary force. The latter zone is called the capillary fringe. The height to which this capillary fringe extends depends on the depth of the groundwater table and the texture and structure of the soil. In the capillary fringe both aeration and water supply are favourable and the water requirements of the plant may be partly or totally fulfilled by this source. In some parts of the world the growth of crops is wholly based upon this type of water provision.

Plants that develop a shallow root system because of waterlogging during the initial growth phases may suffer from water shortage at later periods of drought, although the groundwater table may not be very deep. Thus paradoxically the prevalence of excess moisture in the soil during the early part of the growing season may seriously intensify the adverse effect of drought occurring later in the season. The depth to which the roots of a number of field crops penetrate in a well-drained soil with an adequate moisture supply is presented in Table 1. This table gives average values only. Deviations from these values are often found, due to differences in soil types and crop varieties. The root volume is not evenly distributed over the ultimate depth attained but decreases with depth. For a great number of crops, especially annuals, about 70% of the root volume is found in the first 30 to 60 cm below soil surface.

Table 1. Average depth of root penetration of crops under optimum soil moisture conditions.

Crops	Depth	
	feet	cm
Bulb crops, onions, lettuce	1 - 2	30 - 60
Pasture grasses, cabbage, spinach, beans strawberries, potatoes, carrots, egg plants	2	60
Capsicum spec., squash	2 - 3	60 - 90
Coconut, oilpalm, datepalm	2 - 4	60 - 120
Cotton, lima beans	4	120
Maize, flax, small grains, sugar beet, melons	5 - 6	150 - 180
Alfalfa, sorghum, sudan grass, steppe grasses, sugar cane, deciduous orchards, citrus orchard	5 - 7	150 - 210

4.3 DRAINAGE AND PHYSICAL SOIL CONDITIONS

Physical soil conditions influenced by drainage are structure, aeration, organic matter, and temperature, which will be discussed in turn in this section.

4.3.1 SOIL STRUCTURE

Good structure (aggregation and arrangement of soil particles) means favourable conditions for simultaneous aeration and storage of soil moisture, and also that mechanical impedance to root growth is reduced and stable traction for farm implements is provided. Drainage affects soil structure through its influence on the groundwater level. In soils with a groundwater table at 40 to 60 cm below the soil surface HOOGHOUDT (1952) found a deterioration of structure leading to a more compact and sticky topsoil than was found in soils with deeper groundwater. In surface layers of poorly drained soils, therefore, many large clods are found (HOOGHOUDT, 1952; NICHOLSON and FIRTH, 1958), whereas with well-drained soils small crumbs predominate. Drainage may also increase the pore space, thus promoting cracking and aeration of the soil (WESSELING and VAN WIJK, 1957). Figure 4 shows the influence of groundwater depth on soil-water-air ratios (VAN HOORN, 1958). In Van Hoorn's experiments the percentage of large pores decreased with a shallow water table and the hydraulic conductivity of the 50-90 cm layer decreased (during winter) from 2.5 to 0.35 metres per day.

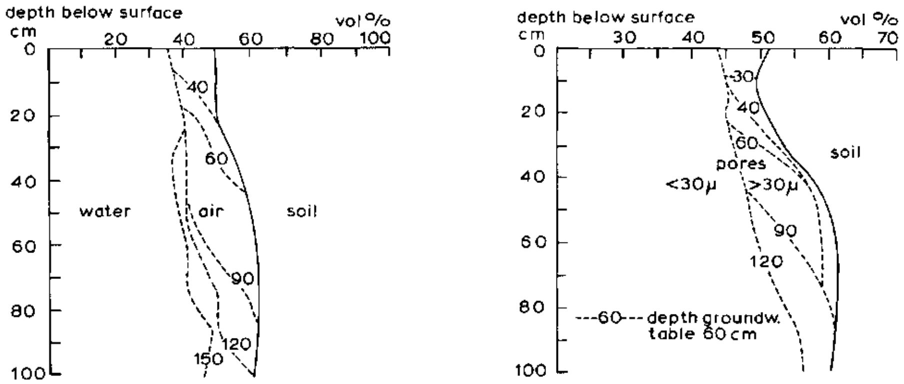


Fig.4. Influence of groundwater depth on water and air content and pore distribution.

It appears that maintaining the soil-water level at a greater depth exerts a beneficial influence on structure and on structurally determined soil properties.

4.3.2 SOIL AERATION

The volume of air in the soil varies inversely with the water content of the soil and is very low in waterlogged or flooded soils. When a soil is permanently flooded oxygen disappears within a few days. In a well-drained soil, as opposed to one with a shallow water table, air not only penetrates into deeper soil layers, but the volume of air in the surface layers is much greater (see Fig.4). WILLIAMSON et al. (1969) found that low millet yields from tanks with shallow water tables were primarily attributable to the low oxygen level. The diffusion rate of oxygen in a soil with deeper groundwater tables was found to be greater and more favourable for plant growth.

4.3.3 SOIL ORGANIC MATTER

Organic matter is important for soil structure as well as for the supply of nutrients. During the decomposition of organic matter, important substances for the build-up of soil aggregates are formed, while at the same time plant nutrients are released. Great losses of organic matter will have a bad influence on soil structure which will impede internal drainage; the soil will become compacted, which has an adverse influence on root penetration. Organic matter must then be supplied either in the form of stable manure, compost or

green manure (including leguminous crops in the rotation). This is especially required in modern farming systems, which return less and less organic matter to the soil, while, because of the greater working depth of mechanical ploughing, the available organic matter is mixed with subsurface soil which is practically devoid of humus.
In organic soils (peat and muck) drainage may lead to subsidence of the land surface. The subsidence is caused by shrinkage due to irreversible drying, oxidation, and compaction. Subsidence may often be accelerated by the customary burning and wind erosion during dry spells.

4.3.4 SOIL TEMPERATURE

The outflow of water and inflow of air brought about by drainage will result in a lowering of the specific heat (Chap.2) of the soil. This means that the soil will warm up sooner, but also that it will lose its warmth sooner. Water requires five times more heat to raise its temperature than dry soil does. Consequently, waterlogged soil, with approximately 50% moisture, requires about 2½ times more heat to warm up than a dry soil does. In addition, the cooling effect of the greater evaporation from a wet soil delays a temperature rise. Both effects cause a delay of growth in spring. In general it can be stated that when the soil is drained the soil surface climate is favourably changed, which will promote early planting in areas with cold winters.

In soils with high water tables, freezing (which implies expansion of the water volume) may cause root damage and heaving of crops that cover the soil during winter (e.g. alfalfa, clover, grass, or winter cereals). Heaving and other damage by frost can be reduced by proper drainage of the soil surface layers.

Sometimes, wet soils have favourable temperature effects, for example, in hot climates or in climates with an occasional frost during the growing season. In hot climates a wet soil may have a lower, more suitable, temperature than a dry soil. Since, however, irrigated fields in hot and arid climates require deep drainage to prevent salinization, the optimum soil moisture content has to be realized by good irrigation practices rather than by restricted drainage.
Concerning late frost, HARRIS et al. (1962) report a 50% stand reduction, due to a frost in June, of maize, potatoes, and peppermint on fields with a groundwater table at 40 inches (100 cm) below the soil surface, whereas almost no damage was observed in fields where the water table was 16 inches (40 cm) deep.

In the latter case the soil was wetter, and could give up heat to the surrounding air in greater quantities and at a faster rate than a drier soil. Thus, with regard to temperature effects, it can be said that differences in groundwater table depths may be favourable or unfavourable, according to circumstances.

4.4 DRAINAGE AND CULTIVATION PRACTICES

This discussion of cultivation practices will be restricted to tillage and weed control.

4.4.1 TILLAGE

With adequate drainage the moisture content of the surface soil layer will, on the average, not rise above field capacity. This is important because there is a rather narrow range of soil moisture contents suitable for tillage operations (Chap.2). The optimum moisture content is below field capacity. Care should therefore be taken to avoid tillage of the soil soon after rainfall or irrigation. Working the soil at higher moisture contents will, in many clayey soils, cause breakdown of aggregates, dispersion of soil particles and, to some extent, puddling of the soil (McGEORGE, 1937). In extreme cases the almost complete destruction of aggregates will result in a compacted soil i.e. a soil devoid of pore space. Such soils are extremely hard when dry. It will take years to build up new aggregates and give the soil a favourable structure again.

As a result of compaction (plough-sole, tractor-sole or traffic layer) and crust formation, both the infiltration rate and hydraulic conductivity are low. This impedes internal drainage, and subsurface drainage systems will not be able to function properly. Subsurface treatment of the soil will then be necessary to restore the internal drainage conditions. Favourable tillage conditions are often met immediately after harvest. By that time evaporation from the surface soil and transpiration through the plant have brought the moisture content of the root zone to far below field capacity. Waiting times after a wet winter from the beginning of a dry period in spring to the moment the soil is tillable have been estimated for The Netherlands by WIND (1963), assuming a daily evaporation of 1-2 mm and groundwater table depths varying from 20 to 160 cm below the soil surface. With deep water tables, the waiting time was only

1 or 2 days. With a water table at 20 cm, the waiting time for a clay soil was 25 days when the evaporation was 1 mm, and 7 days when the evaporation was 2 mm. For a silty loam the waiting times were 69 and 20 days respectively. If it is possible to grow a cover crop in the winter or off-season, preferably a leguminous crop like clover, tillage can start earlier because of the drying effect of this crop. When such a crop is wholly or partly used as green manure, it will at the same time improve the soil structure. The nutrient status of the soil will be improved accordingly.

To derive the maximum benefit from drainage, the soil should receive an adequately level tillage to avoid differences in relief.

4.4.2 WEED CONTROL

A large proportion of tillage operations is devoted to the control of weed growth. Weed growth is most abundant and troublesome with high soil moisture contents. Good drainage therefore reduces both the need for tillage and the hazard of soil structure deterioration. Going from wet soils with high groundwater tables to drier soils (water level 100 cm below the soil surface in sands, or 200 cm in clays), the kinds of weeds will change from hydrophytes - marsh vegetation like Scirpus (bull rush), Typha (cattail), Spartina (cordgrass), Carex (sedges) - to various Cyperaceae intermixed with a number of grasses. With the soil water still deeper more grasses appear, which may ultimately be suppressed by various broadleaved plants. With very deep water tables in arid regions, weeds with xerophytic or phreatophytic characteristics, or both, will survive.

Most crops are mesophytic, showing their best growth and development on soils with moderate moisture contents. The weeds associated with these crops are predominantly broadleaved plants in well-drained soils and some grass species in poorly drained soils. Broadleaved weeds are more easily eradicated - whether by hand, implements or weedicides - than most grasses are. A careful tending of the crop, and the application of the rotations that are possible on well-drained soils also contribute to the effective control of weed growth. To illustrate the competition between alfalfa and weeds may be mentioned. Alfalfa will suffer from high groundwater tables, especially when the growth of grasses is stimulated by these wet conditions. On the other hand, with adequate drainage, not only will grasses disappear, but the vigorous growth of the alfalfa will control growth of broadleaved weeds like Cirsium arvense (Canada thistle). Flood-fallowing (FOLLET-SMITH and ROBINSON, 1936) as practised in Guiana on sugar cane land, seems to be rather exceptional in this connection and may serve to show that under certain conditions a different approach is possible in the improvement of soil tilth and the control of weeds. In Guiana, at the end of the cropping cycle,of sugar cane, the structure of the soil is rather poor and there are a lot of weeds. The land is then flooded to a depth of 1 foot for 6 months. During this time thrash decomposes and anaerobic reactions occur. Meanwhile heavy growth of water weeds suffocates the canefield weeds. When the fields are drained oxidation occurs. The result is a marked improvement in tilth and the eradication of the cane field weeds.

4.5 DRAINAGE AND NUTRIENT SUPPLY

Various processes activated by bacteria, fungi, or other micro and macro organisms depend on good aeration and drainage. Nitrogen fixation and nitrification by micro organisms may be mentioned as two of the principal aerobic processes exerting an important influence on plant growth and development. The deeper the roots can penetrate, the more nutrients there are available for absorption. The advantage of drainage and a consequent deep root zone is even more pronounced when nutrients have been displaced to the deeper layers. That the uptake of nutrients (N, P, K, Ca and Mg) depends on the depth of the groundwater table is illustrated in Fig.5 for two citrus varieties grown under field conditions in the UAR (MINESSY et al., 1971).

Good drainage will also increase the microbiological decomposition of organic matter, as a result of which plant nutrients like nitrogen and phosphate are released and become readily available. This process may be either favourable (when organic matter is amply present) or unfavourable when loss of organic matter means deterioration of soil structure.
Good drainage prevents the production of harmful reduced substances that might form under anaerobiosis. For example, it will prevent high concentrations of dissolved manganese to which alfalfa is sensitive. Harmful products may also occur when rice is a regular crop in the rotation. The continuous flooding of rice fields causes a reduction of the soil and an accumulation of toxic products such as H_2S. An occasional drainage of the fields will result in a reoxidation of the reduced substances.

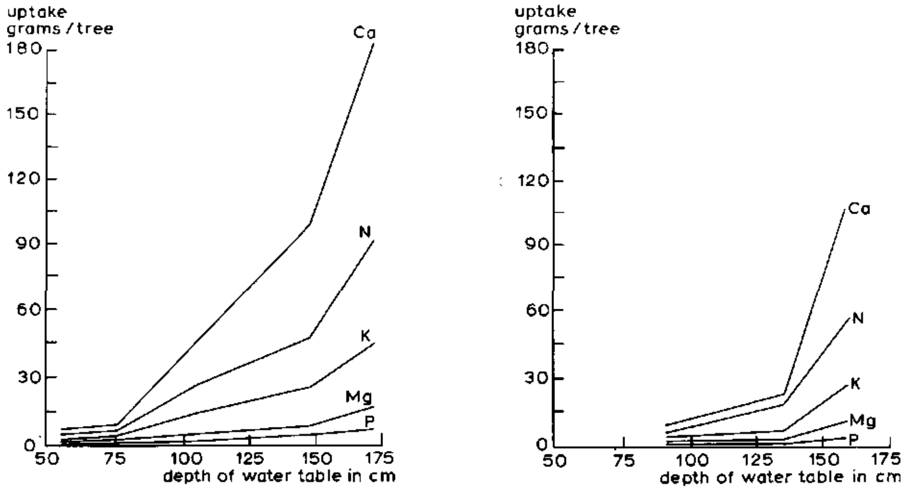


Fig.5. The effect of the depth of water table on the uptake of some plant nutrients by newly produced leaves per year in Washington navel orange (A) and Balady mandarin (B) (avs., 1965 and 1966).

With respect to the influence of depth to groundwater on nitrogen availability, VAN HOORN (1958) found that with high water tables the colour of plants was often yellowish, indicative of a shortage of nitrogen. It is interesting that the leguminous pulse crops show a different reaction. Because of symbiosis of pulse crops with Rhizobium radicicola there is an autotrophic nitrogen supply, which appears to reduce the influence of high groundwater levels. Fig.6 shows the influence of depth of the groundwater table on the amount of nitrogen supplied by the soil to cereals (VAN HOORN, 1958).

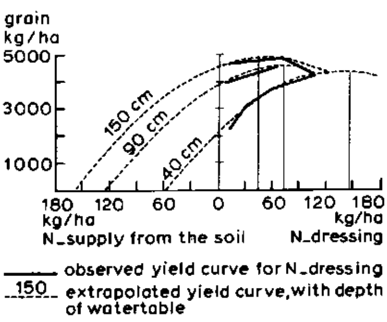


Fig.6. Influence of depth of water table on nitrogen supplied by the soil.

When the groundwater level is at 150 cm, the nitrogen supplied by the soil appears to be 150 kg N per ha. When the groundwater level is at 40 cm, less than 60 kg N per ha is supplied. Thus, with groundwater at 40 cm, 100 kg N per ha has to be applied in the form of nitrogen fertilizer to obtain a yield comparable with the maximum yield obtained with groundwater at 150 cm.

HARRIS et al. (1962) report that for crops grown on a muck soil with groundwater at 40 cm, yields were only 63% of those obtained when groundwater was at 100 cm. With an application of nitrogen fertilizer, the yield could be increased by an average of 36%, which made it the same as that obtained with groundwater at 100 cm. This indicates that a shallow groundwater table interferes with the nitrogen supply. SHALHEVET and ZWERMAN (1962) also describe an experiment on the relations between drainage and nitrogen supply (Table 2).

Table 2. Yields of maize (kg/ha) in relation to drainage conditions and nitrogen fertilizer (SHALHEVET and ZWERMAN, 1962).

Fertilizer	Drainage conditions		
	Good	Intermediate	Poor
NO_3^-	2800	2036	1190
NH_4^+	3320	1895	591
none	2843	931	249

With good drainage conditions only those fertilizers containing ammonia were effective; this may be due to the fact that ammonia, in contrast with nitrate, is first adsorbed to the soil and then only gradually released. With poor drainage, yields on the whole were lower; nitrate fertilizer, however, produced a remarkable effect (not achieved by ammonia, in contrast with its performance under good drainage conditions). For intermediate drainage conditions the effect of nitrate fertilizer was approximately equal to that of the ammonia fertilizer. Also of interest with regard to the relations between yield, nitrogen supply, source of nitrogen, and groundwater table depth are the experiments described by HOOGERKAMP and WOLDRINGH (1965) with sugar beet on heavy river clay soil (Table 3).

Table 3. Yields of sugar beet in relation to groundwater table depth and nitrogen fertilizers (HOOGERKAMP and WOLDRINGH, 1965).

	Groundwater table depth in cm							
	40		65		95		140	
fertilizer 140 kg/N/ha	$NO_3^-+NH_4^+$	NH_4^+	$NO_3^-+NH_4^+$	NH_4^+	$NO_3^-+NH_4^+$	NH_4^+	$NO_3^-+NH_4^+$	NH_4^+
yield of roots kg/100 m^2	329	167	354	220	353	262	358	331
yield of leaves kg/100 m^2	260	188	269	167	247	197	251	244
yield of sugar kg/100 m^2	57.5	28.2	63.6	39.1	65.8	47.9	66.0	59.0
sugar %	17.47	16.88	17.96	17.84	18.63	18.33	18.46	17.89

Table 4 presents the relative sugar yields obtained with the two types of fertilizer at different groundwater table depths.

Table 4. Relative yields of sugar with different groundwater tables (HOOGERKAMP and WOLDRINGH, 1965).

groundwater table depth	40	65	95	140
relative yield with NO_3+NH_4	89.6	93.4	99.1	100
relative yield with NH_4	53.1	61.1	73.1	90.7

With both types of fertilizer the best results were obtained with the deepest groundwater table. Generally, the effect of the nitrate fertilizer was better. With this type of fertilizer the yield obtained when the groundwater table depth was at 40 cm was virtually the same as that obtained with the ammonia fertilizer when groundwater was at a depth of 140 cm. WILLIAMSON and WILLEY (1969) found with tall fescue grass that the beneficial effect of nitrate fertilizer lasted for only a few weeks. Denitrification and leaching of nitrate soon followed when water tables were high.

The better results obtained with nitrate fertilizer at shallow water table depths may be attributable to the favourable effect of nitrates on aeration.

It appears that for each cm the water table is lowered, about 1 kg N/ha becomes available for plant production. The fact that, under certain cropping systems, nutrient losses (N, etc.) are sometimes observed in tile drainage effluents, would seem to contradict the theory of a gain in nutrients, e.g. N by a lowering of the groundwater table through drainage. It is, however, understandable that under good aeration conditions in well-drained soils, where nitrification is promoted, easily soluble nitrates are carried away by the water that percolates through the soil to the drainpipes. The higher the amount of flowing water, the greater the losses. BOLTON et al. (1970) found that greater nutrient losses from continuous fertilized maize, for example, as against losses from blue grass sod, were associated with total annual, average effluent flows of 155.7 and 64.5 mm for these respective treatments. The annual gain in N should therefore be seen as the balance between N becoming available under good drainage conditions and that lost through drainpipes.

4.6 DRAINAGE AND SALINITY

Soil salinity refers to the presence of high concentrations of soluble salts in the soil moisture of the root zone. These concentrations of soluble salts, through their high osmotic pressures, affect plant growth by restricting the uptake of water by the roots. All plants are subject to this influence, but sensitivity to high osmotic pressures varies widely among plant species. Salinity can also affect plant growth because the high concentration of salts in the soil solution interferes with a balanced absorption of essential nutritional ions by the plant.

Halophytes, which are plants adapted to saline conditions, may be used as indicators of the salinity level of the soil. Representative of this group are: Atriplex hastata, Atriplex vesicaria, Salicornia spp., Salsola spp., Chenopodium album and Portulaca oleracea.

High concentrations of sodium (alkalinity) will also affect soil physical conditions, by the dispersion of clay particles (Chap.3). The result is a deterioration of soil structure. This reduces the infiltration and percolation capacity of the soil so that the movement of water into and through the soil, as well as the diffusion and exchange of gases, will be impeded. The loss of soil structure also results in a crust formation and compaction of the soil, which will obstruct the emergence of young seedlings and the development of roots.

The main effects of salinity on plant growth and crop production are:

- slow and insufficient germination of seeds, a patchy stand of the crop,
- physiologic drought, wilting, and desiccation of plants,
- stunted growth, small leaves, short stems and branches,
- bluish green leaf colour,
- retarded flowering, fewer flowers, sterility, and smaller seeds,
- growth of salt-tolerant or halophilous weed plants,
- as a result of all these unfavourable factors, low yields of seeds and other plant parts.

The term "salt tolerance" indicates the degree of salinity a plant can withstand without being appreciably affected in its growth or development. The salt tolerance of plants can be conveniently related to the electrical conductivity of the saturation extract (EC_e in mmho) of the soil in the root zone of the plants. In field experiments with some principal crops BERNSTEIN (1964) determined the salinity levels causing yield reductions of 10%, 25% and 50%.
It appeared that most field crops (e.g. wheat, oats, rice,and rye) have a salt tolerance of EC_e 4-8 mmho. Some field crops (barley, sugar beet, cotton), vegetables (garden beets, kale, spinach, asparagus), and fruit crops (date palm, mulberry, olive, pomegranate, jujube) have a higher salt tolerance of EC_e 8-16 mmho. Some grasses such as Sporobolus, Pucinellia, Cynodon dactylon (Bermuda grasses), Chloris gayana (Rhodes grass) and Agropyron elongatum (tall wheatgrass) also have a high salt tolerance (EC_e 8-16 mmho). Field beans are salt-sensitive, having a salt tolerance of EC_e 2-4 mmho.
For comparison, rice showed a reduction in yield of 10%, 25% and 50% at an EC_e of 5, 6 and 8 respectively, whereas the same yield reductions for barley were found at higher EC_e values of 12, 16 and 18 respectively.

If the land is liable to become saline or alkaline, adequate drainage will remove or reduce these dangers (Chap.4), thus ensuring a better crop production. If the land is already saline or alkaline it can be reclaimed with a good combination of drainage and irrigation. If the land is alkaline, chemical amendments, e.g. gypsum, may be applied. Often the introduction of a reclamation crop will accelerate the process of reclamation. For example, lowland rice is often used during reclamation in sub-tropical and tropical climates. The flooded condition of the fields promotes a continuous leaching of salt from the soil, and also a dilution of the salt solution in the soil water. Moreover the fact that rice seedlings may be grown in nurseries, under less saline conditions, make it possible to grow rice during the early phase of reclamation. Grasses (Bermuda grass or tall wheatgrass) and barley may be chosen as reclamation crops in climates less favourable for rice production.

4.7 DRAINAGE AND DISEASES OR PESTS

Drainage, through its influence on soil conditions and plant growth, may either favourably or unfavourably influence the incidence of diseases and pests. A few examples will serve to illustrate this statement. A high groundwater table may have a favourable effect on potato production, but this advantage may largely be offset by blight (Phytophthora infestans). Blight occurs more frequently when groundwater tables are high since high water levels are required for the dispersion and germination of spores (GRABLE, 1966). Also, because the cork development of the tuber is reduced at low oxygen concentrations, high groundwater tables favour the infection of potato tubers by Bacillus atrosepticus. On the other hand, Actinomyces scabies or potato scab (scabies) can be checked with a restricted oxygen supply. An abundant nitrogen supply encourages the multiplication of the green peach louse (Myzus persicae), the vector of potato virus diseases. As nitrogen is in more plentiful supply when groundwater tables are deep, (Sect.5) drainage may thus exert an indirect influence on the occurrence of virus diseases. Mildew infections of winter wheat are positively related to high water tables, and are suppressed by improved drainage. However, the improved nitrogen supply resulting from a deeper groundwater level may promote the incidence of rust diseases in a similar way as for virus diseases in potatoes. This is also the case with tobacco. With a higher nitrogen supply the more succulent soft leaves are susceptible to diseases like Phytophthora infestans, Peronospora tabacina and virus diseases. A high supply of nitrogen also has an unfavourable effect on the quality of tobacco leaves. With cotton, wet conditions lead to excessive vegetative growth, retarded maturation, opening of the bolls, and boll-weevil damage; boll rot also increases. If, however, these wet conditions are reduced, the increased availability of nitrogen will promote the incidence of vascular wilt (Fusarium oxysporum var, vasinfectum). Collar rot or trunk cancer in apple trees, and white root diseases (Fomes lignosus) in Hevea rubber trees occur under bad drainage conditions. For bananas, good drainage is an absolute must for the production of high quality fruit, while waterlogging or flooding may enhance the occurrence of Panama disease (Fusarium oxysporum f.cubense). On the other hand, it has been found that flood fallowing for a year or more is necessary to "reclaim" soils infested by the fungus.

Growing flooded rice on organic soils kills certain fungous diseases and certain nematodes that are harmful to vegetable crops. In such a case wet conditions are favourable.

4.8 GROUNDWATER LEVEL AND CROP PRODUCTION

By keeping the groundwater table at a more or less fixed depth, information can be obtained as to the effect that drainage to such a depth has on crop production. In practice, however, the groundwater level is generally not constant but fluctuating, even in fields with an adequate drainage system. There may be a number of days when the groundwater table is higher than the average. These excessive soil water levels, even though of a relatively short duration, may exert an influence on production, depending on the growth phase of the plant when they take place. Annual crops are especially sensitive during germination and in the reproductive phase. In the dormant phase, many plants (e.g. deciduous trees or meadow grasses) are generally not adversely affected by too much water or too little air.

In a number of field experiments on drain spacing and depth conducted in the Northeast polder (Netherlands), the fluctuation of water levels was recorded over a period of 10 years. SIEBEN (1965) found a relation between yield reduction of wheat and frequency of high groundwater tables. He introduced the SEW-value (sum exceedance value in winter), which is obtained by summing all daily values (in cm) by which the groundwater levels (midway between drains) exceed a reference level in winter (1 November to 1 March). For example, if the reference level is 30 cm below the soil surface and the water table on three successive days shows levels of 25, 10 and 5 cm below the soil surface, then the contribution to the SEW-value is (30-25) + (30-10) + (30-5) = 5 + 20 + 25 = 50 cm. Sieben found that the SEW-value - as a parameter for the groundwater regime in the upper layer of the soil - showed, above a certain minimum, negative relations with crop yields. It was relatively unimportant which reference level,

within certain limits, was chosen. Thus, above a certain minimum SEW-value, yields reduced. Hence, it is concluded that even during a rather inactive phase of growth as found with wheat in winter, excessive water may exert a bad influence on plant production.
BERTRAM (1931) in an experiment with barley, found that a one-time rise in the water table from 80 cm to 10 cm below soil surface during the reproductive phase reduced the yield to 20%.
Applying, on the other hand, SES values (sum exceedance value in spring) below a certain maximum, positive relations with crop yields were found, i.e. soils with higher groundwater tables, within certain limits, are favourable for crop production. The explanation is that during dry spells crops growing on soils with deep groundwater tables will suffer more from drought than crops on soils with less deep water tables, especially when the plants are in an active phase. In general it may be said that the seriousness of the effect of fluctuations in the groundwater table will depend on whether plants are in an active phase (e.g. the reproductive phase) or in a less active or latent phase (during winter). In an active phase both an optimal aeration and moisture content are important.

In the following the effect of the depth of the groundwater level on the production of different crops, viz. grassland, arable crops and fruit trees, will be discussed.

4.8.1 GRASSLAND

Interesting in this connection are the results of experiments conducted on heavy river-clay soils in The Netherlands. (HOOGERKAMP and WOLDRING, 1965, 1967). Their purpose was to study the influence of the depth of the groundwater level on the grass yield of old grassland. It was found that groundwater levels at 25, 40, 65, 95 and 140 cm below the soil surface affect grass production in various ways. In periods with excess precipitation over evapotranspiration, the highest yields were found on the most intensively drained plots; in dry periods the reverse was true. This is in general agreement with findings reported elsewhere (VAN 't WOUDT, 1957). In The Netherlands the highest average yields of grassland are found when groundwater table depths are between 60 and 80 cm for fine-textured soils and between 40-60 cm for sandy soils.

Many pasture grass species, having ordinarily a superficial root system, require or tolerate a higher groundwater table than most arable crops. Some other grass

species, however, have a much deeper-penetrating root system and grow better on drier soils with deeper water tables. In the experiments described by HOOGERKAMP and WOLDRING (1965, 1967) the species Alopecurus pratensis was stimulated by lowering the water table, whilst Agrostis stolonifera was stimulated by raising it. In an experiment on sandy soils, the percentage of the grass species Festuca pratensis and Phleum pratensis was much higher with groundwater levels at 30 and 50 cm than at deeper levels.
For the grass Lolium perenne, the reverse was true. The clover Trifolium repens was most abundant with intermediate groundwater depths of 70 and 90 cm; other clover species appeared spontaneously when the water table was deeper. Very shallow water tables favour undesirable plants such as rushes, sedges, unpalatable grasses, which tend to become dominant under these wet conditions (HUDSON, et al., 1962).

In general it is found that with adequate drainage the production of grass starts earlier in spring and proceeds longer in autumn. The protein content of grass and hay is often higher with deeper groundwater tables. This may be due partly to the gradually increasing influence of the clovers and partly to an increased nitrogen supply by the soil as the depth of the groundwater level increases.
Earthworms are important for the soil structure and good drainage promotes their presence. EDMUND (1963) found 15 earthworms per square foot in undrained grassland against 60 in well-drained land with better soil structure.

When pastures with shallow groundwater tables are grazed, the structure of the soil surface deteriorates (HUDSON et al., 1962). Plants are trampled and will often be killed because of root damage, plant displacement, or burial in mud. The result is that other less palatable and less valuable, but more resistant, grass species will become dominant. Livestock, too, will suffer from the wet conditions resulting from shallow groundwater tables. HUDSON, et al. (1962) mention the occurrence of footrot, chills, and internal parasites with cattle, and losses of lambs by chills when pastures are wet.

Grass and clover plants can withstand long periods of inundation fairly well when they are dormant or during the early stages of growth (McKENZIE 1951, RHOADES 1967). Some grass species appeared to be tolerant to more than 20 days of inundation. Clovers are, with a few exceptions, less tolerant to inundation

(McKENZIE 1951, HOVELAND and WEBSTER 1965).

4.8.2 ARABLE CROPS

Important arable crops like cereal crops (maize, wheat, rice, barley, rye, oats, sorghum), root and tuber crops (cassave, sugar beet, sweet potatoes), bulb crops (onions, tulips), fibre crops (cotton, flax, jute, kenaf), pulse crops (peas and beans) and sugar cane often have different drainage requirements. A common characteristic of these crops, except for irrigated lowland rice, is that for optimum root development they require a well-drained friable soil in which both water plus nutrients and air exchange are adequate. The optimum depth of the groundwater table, however, depends on the depth of the root system, on soil characteristics, and on climate, and consequently will differ according to plant species and the growth phase of the plant. When two or more crops are grown in rotation on the same field, the depth and spacing of a drainage system should offer a compromise between the different requirements of the crops.

Flooding or waterlogging of arable crops will usually be detrimental, although some varieties of sugar cane and leguminous crops may tolerate such conditions. In Florida sugar cane is grown on organic soils where the groundwater table is purposely kept high in order to keep oxidation, subsidence, and wind erosion to a minimum. HUMBERT (1968) mentions that sugar cane can survive long periods of inundation if the tops of the plants are above water and the water is moving. Injury to cane by waterlogging will, however, occur when the air temperature is very high. This type of damage is often referred to as scalding.

Leguminous crops - clovers, pulse crops - perform well with intermediate groundwater depths of about 50 cm. Alfalfa may adapt to shallow non-fluctuating water tables, but on the other hand, if subjected to dry conditions from the start, may ultimately penetrate to depths or more than 2 meters and behave like a phreatophyte. Plants such as grasses and clovers, grown for their vegetative parts (leaves and stems), and vegetables like kale, lettuce and potatoes, produce favourably with shallow water tables of 30-50 cm (VAN 't WOUDT and HAGAN, 1957).

A long-term groundwater level experiment with a number of arable crops was conducted by VAN HOORN (1958) in Nieuw Beerta (The Netherlands) on a clay soil (48% < 2μ, 20% from 2-16μ). Groundwater levels were maintained during summer at

40, 60, 90, 120 and 150 cm below the soil surface. The results are presented in Table 5.

Table 5. Yields of various crops with groundwater levels from 40-150 cm. The maximum yield is taken as 100% (VAN HOORN, 1958).

Crop	Number of years	Relative yield of grain, roots or tubers at groundwater depths (cm) of:					Yield at 100% level in kg/ha	Relative yield of straw at groundwater depths (cm) of:					Yield at 100% level in kg/ha
		40	60	90	120	150		40	60	90	120	150	
Wheat	6	58	77	89	95	100	4600	59	75	84	92	100	8600
Barley	5	58	80	89	95	100	4100	57	76	84	93	100	5150
Oats	3	49	74	85	99	100	5000	60	82	89	98	100	5850
Peas	4	50	90	100	100	100	2750	67	94	100	100	100	3550
Beans	3	79	84	90	94	100	3100	86	95	100	100	100	4500
Caraway	3	80	96	98	100	100	1700	93	98	97	100	100	5100
Rape seed	2	79	95	95	98	100	2500	70	84	92	97	100	6400
Sugar beet seed	1	75	82	90	96	100	4250	78	94	95	100	100	6500
Sugar beet	2	71	84	92	97	100	40500						
Potatoes	1	90	100	95	92	96	26000						

The percentages in Table 5 indicate that with the exception of potatoes the best yield of all crops was obtained with the deepest groundwater table.
The yield of peas with watertable depths of 90 and 120 cm and the yield of careway with a water table depth of 120 cm were the same as those with a water table depth of 150 cm below the soil surface. The yields of the other crops were a few percent lower when groundwater depths were at 90 and 120 cm. Potatoes showed the highest yields in plots with a groundwater depth of 60 cm. Over the whole range of water table depths, however, the variation of potato yields was only 10 percent or less. Throughout their development potatoes require a continuous and plentiful supply of water in the root zone. With the intermediate rooting depth of the potatoes, an intermediate depth of the groundwater table will guarantee this regular water supply from the capillary fringe. In a great number of farm fields located on river ridge soils, FERRARI (1952) found the highest yields with intermediate water table depths of 50-75 cm.
HARRIS, et al. (1962) described the relations between yield and groundwater depth obtained from experiments on a muck soil. Some results are presented in

Table 6.

Table 6. Relative yields of crops at different groundwater table depths (HARRIS et al., 1962).

Crop	Number of years	Relative yield in % at groundwater depth of:				Yield at 100% level per acre
		16 inches 40 cm	14 inches 60 cm	32 inches 80 cm	40 inches 100 cm	
Potatoes	12	46	94	97	100	327 bu
Maize	9	71	100	103	100	126 bu
Peppermint	13	48	91	100	100	32 lb. oil
Onions	11	63	109	113	100	3335 lb
Sweet corn	4	61	100	92	100	5.1 tons
Carrots	4	59	93	96	100	2.7 tons
Average		63%	98%	100%	100%	

In these experiments the yields of potatoes and carrots were highest with the deepest water table of 100 cm below soil surface. The yields of maize and onions were best when groundwater was at 80 cm. With groundwater at 60 cm the yield of onions was still more than that at 100 cm. The yields of maize and sweet corn were the same at 60 cm as at 100 cm water table depth. In North Carolina (USA) DISETER and VAN SCHILFGAARDE (1958) found that yields of maize did not esentially differ when drain depths were 2, 3 or 4 ft with a spacing of 160 ft (50 m). These yields, however, were about 34 bu/acre (2,000 kg/ha) higher than those from poorly drained land with a drain spacing of 310 ft (100 m). In a three-year experiment with maize, SCHWAB et al. (1966) found an average yield of about 2,500 kg/ha in fields without drainage and more than 4,000 kg/ha in fields with a drainage system that consisted of a surface drainage system or of a tile drainage system (depth 3 ft and spacing 40 ft). The effect of applied nitrogen fertilizer was considerably higher with drainage. With 112 kg N/ha the yield in the undrained plots increased by 500 kg/ha and attained a level of 3,000 kg/ha. With the same amount of fertilizer the yield on drained plots increased by 2,000 kg/ha and attained a level of 6,000 kg/ha. In this particular field the effect of nitrogen fertilizer did not eliminate the adverse influence of bad drainage conditions as in the cases mentioned in Sect.5.

4.8.3 FRUIT TREES

The relatively inflexible root systems of trees with active roots developing only at relatively great depths appear to be easily affected by high groundwater tables. Walnut trees, for instance, prefer a water table depth of 8 to 10 feet (VAN 't WOUDT and HAGAN, 1957). A tree's susceptibility to poor drainage, however, depends on its age and the season. In summer, for instance, the roots of apple trees will suffer greater damage from poor drainage than in winter when the trees are leafless and inactive. PENMAN (1938) observed citrus trees to remain healthy for the first 8 or 10 years of their life with a water table within 4 feet of the soil surface. Beyond that age they require a deeper water table. This is illustrated in Fig.7 and 8 for citrus varieties (MINESSY et al., 1971).

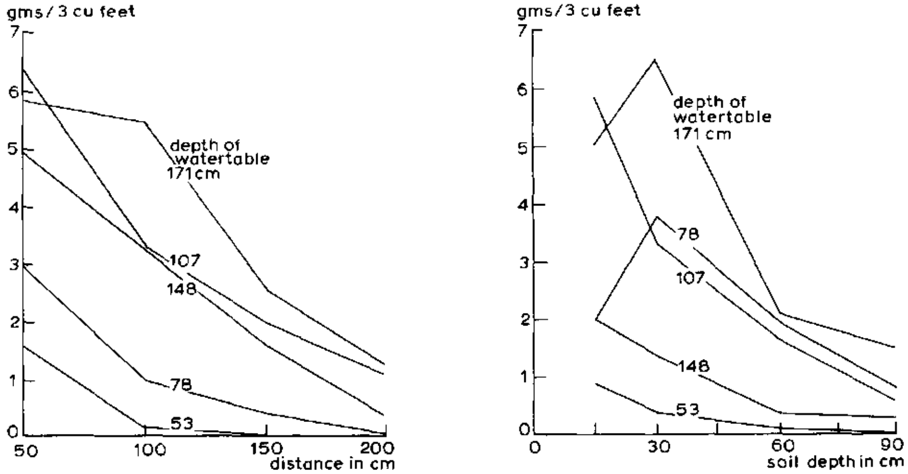


Fig.7. Quantity of roots as a function of water table depth and distance from tree (A) and depth (B) for Washington navel orange.

Extension and depth of roots increase with deeper groundwater tables. Yields of fruits are only obtained from trees growing on soils with a water table deeper than 1 m. The highest yield (41 kg fruit per tree) is obtained on soils with the lowest water table. The same was found for plum trees (VISSER, 1947). In orchards, palm groves, and rubber estates, cover crops may function to lower the water table during times of excessive water.

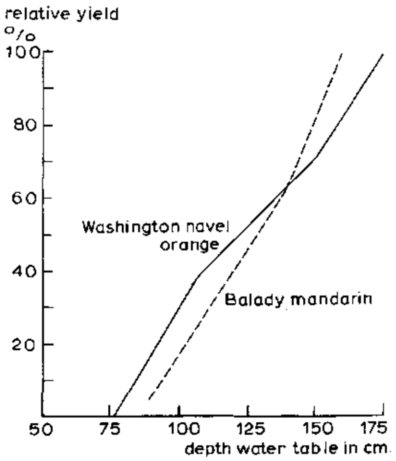


Fig.8. The effect of the depth of water table on the relative yield of Washington navel orange and Balady mandarin.

Coconut palms, date palms, and oil palms have rather coarse and shallow root systems, devoid of root hairs. When growing under anaerobic conditions they develop air cavities in the roots, which may facilitate survival under temporarily anaerobic conditions.

Coconuts often flourish in places where there is underground water within the root's range, the roots "touching" the water. Coconuts are growing in coastal regions where the influence of the tides causes temporary waterlogging. If, however, the soil becomes permanently waterlogged, the leaves become sickly and yellow. In Malaya the oil palm yields more than 1.5 tons of oil per acre on the relatively rich alluvial clays (70% < 2μ) of the western coastal plains (GRAY, 1963). The shallow rooting oil palm is well suited to the high water tables prevailing in these areas. This is also true for tropical fruit trees like Zalacca edulis (also a palm) and Garcinia mangostana (mangosteen), which can stand excessive water in the soil without a decrease in yield (TERRA, 1948). Planted on soils high in organic matter, crops like oil palm, coconut, and, to some extent, rubber and fruit trees may suffer severe damage when such soils are drained and subsidence occurs. The shrinkage of the soil may cause the rootmats to appear more than 2 or 3 feet above the soil surface. The trees may then fall down and their stems may be seen growing in all directions.

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INTRODUCTORY SUBJECTS

5. PHYSICS OF SOIL MOISTURE

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PURPOSE AND SCOPE

The forces acting on water in an unsaturated soil are discussed. Retention and movement of soil water are described by equations based on the potential concept.

CONTENTS

		page
5.1	The presence of water in the soil	125
5.2	The retention of water by the soil	125
	5.2.1 Field capacity and wilting point	125
	5.2.2 The mechanisms of retention of water by soil	127
	5.2.3 Retention curves	129
5.3	Forces and potentials	131
	5.3.1 The concept of potential	131
	5.3.2 The component and total soil water potentials	134
5.4	Transport of water in the soil	140
	5.4.1 Transport equations based on the potential concept	140
	5.4.2 Infiltration	143
	5.4.3 Capillary rise	148
Literature		151

5.1 THE PRESENCE OF WATER IN THE SOIL

Water can be present in the soil as a solid (ice), a liquid, or a vapour. In this chapter we will mainly be concerned with the liquid phase. Liquid water serves within the soil as a leaching agent, a solvent, a reactant, as a medium for chemical reactions, and as a plasticizing agent. Soil water always contains many dissolved substances. Even in soils considered to be non-saline, the total electrolyte concentration of soil water varies between 1 and 20 meq/liter, and it may vary considerably within the soil profile as a result of evaporation and leaching. Those properties of soil water that result from its chemical composition will not be considered further in this chapter.

The physical behaviour of soil water depends to a considerable extent on the soil properties. Soil, which is a porous medium, exhibits a large surface accessible to water. This solid surface ranges from about 1000 cm^2/g for coarse sands to over 1,000,000 cm^2/g for clay soils. Soil particles are generally hydrophillic, i.e. water tends to adhere to the solid surfaces. In soils that are not fully saturated, soil water also shares a large interfacial area with the gas phase. These two types of interfacial interactions, i.e. at the solid-liquid and the liquid-air interface, determine the retention of water by the soil and the movement of water through the soil.

The total pore space of a soil can be filled with air, water, or both. The last-mentioned situation is the most desirable for soils on which agriculture is practised. The air-filled porosity, then, is the fraction of the bulk volume occupied by air. The water-filled porosity, or fraction of the bulk volume occupied by water, is often called the soil water ratio (soil water content on a volume basis). Soil water content is more commonly expressed as the "dry weight soil moisture fraction", i.e. the ratio of the mass of water to the mass of dry soil. Hence, the soil water ratio multiplied by the ratio of water density over bulk density is equal to the dry weight soil moisture fraction.

5.2 THE RETENTION OF WATER BY THE SOIL

5.2.1 FIELD CAPACITY AND WILTING POINT

From the fact that water will continue to enter the soil, it cannot be concluded

that the pore space is not filled and that it is possible to store more water in the soil: when a certain maximum amount of water is stored the remainder will drain away. This has led to the concept of field capacity. The amount of water stored in the soil at field capacity is that amount of water which a soil will hold against gravitational forces.
This water content is not a unique value that normally occurs in the soil. Soil is a dynamic system of water removal by drainage, evaporation, and absorption by plant roots, and the addition of water by rain, dew, irrigation, or by capillary rise from a groundwater table. When there is a shallow water table but no strong upward movement of water in the soil resulting from evaporation at the soil surface, an equilibrium situation will be reached between upward movement as a result of capillary rise and downward movement because of the gravitational pull. This led to a second definition of field capacity as the amount of water near the soil surface when in equilibrium with a groundwater table at a depth of 1 m. This is not necessarily the same amount of water that a well-drained soil will hold against gravitational forces.
In the latter case field capacity refers to a range of water contents where the rate of water removal from the soil, following irrigation or heavy rain, begins to reduce. Movement of water downward in the soil does not cease when the field capacity has been reached but continues for a long time at a reduced rate.

Field capacity has been used to indicate the upper limit of water available to plant growth, assuming that the amount in excess of field capacity drains away too quickly to be of any use. This is somewhat misleading, however, because all water that is not held tightly in the soil is available for plant growth as long as it is in contact with the roots.
The permanent wilting point, the lower limit of available water in the soil, is the water content of the soil at which the leaves of plants growing in it show wilting and fail to recover when placed overnight in a near saturated atmosphere. Like the field capacity, this is a range of water contents. It is interesting to note that at permanent wilting point the air within the soil pores in equilibrium with the soil water has a relative humidity of 98.8%.

The amount of water in a soil is in itself no effective indication of its availability; a better indicator is the force with which the water is kept by the soil.

5.2.2 THE MECHANISMS OF RETENTION OF WATER BY SOIL

Several mechanisms are active in the adsorption of water by soil particles. Those resulting from the electrostatic charge of soil particles and the presence of adsorbed counterions only act over a short range. They cause a strong binding of a very thin film but are of little consequence at those higher water contents we are considering.

That those larger amounts of water can be retained in soil results from the presence of air-water interfaces similarly as exist in blotting paper or sponges. Surface tension acting at the air-water interface provide the mechanism of water retention.

Surface tension is caused by the mutual attraction of water molecules (cohesion). Within a waterbody, the mutual attraction of water molecules is the same in all directions; hence the net attraction is zero. At an air-water interface, however, there remains a net force pulling the molecules inwards into the bulk of the water. This force results in a surface tension (σ) which acts to reduce the air-water interface.

The energy required to enlarge an air-water interface(against the surface tension) by 1 cm^2 is 72 erg at 25^oC. Thus σ equals 72 erg/cm^2 or 72 dyne/cm. Water molecules in contact with solid surfaces are attracted (this is called adhesion). Surface tension also plays a role when water adheres to solid surfaces. The adhesion of liquids to solids can be described by the amount of mechanical work required to separate them when they are pulled apart at right angles from one another. This work of adhesion (W) is related to the liquid-solid contact angle (α) by the equation

$$W = \sigma(1 + \cos\alpha) \qquad (1)$$

where σ is the surface tension. Those surfaces which have an angle of contact with water of 0^o (e.g. clean glass, quartz) have a work of adhesion of 2 σ erg/cm^2. Water adheres to those surfaces as strongly as it does to itself, since the work of cohesion is also 2 σ(two new air-water interfaces are created). Thus water wets the walls of a glass capillary tube (see Fig.1), and the water adhering to the walls pulls the body of the liquid up to a height h. This phenomenon is called capillary rise.

The force tending to pull the liquid up in a capillary with radius r is $2\pi r\sigma \cos \alpha$ and the force tending to pull the liquid down, caused by the weight of the water column above the free water surface is $\pi r^2 h \rho_w g$ in which g is the acceleration due to gravity and ρ_w the density of water. At equilibrium with a zero contact angle we obtain:

$$\rho_w gh = \frac{2\sigma}{r} \tag{2}$$

The difference between the hydrostatic pressure in the tube on a level with the free water surface (P_2) and the hydrostatic pressure immediately below the interface (P_1) is equal to $\rho_w gh$.
Hence,

$$P_2 - P_1 = \frac{2\sigma}{r} \tag{3}$$

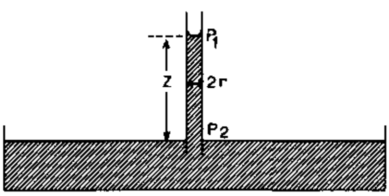

Fig.1. Capillary rise of water table into a tube.

The hydrostatic pressure above the interface differs only by a very small amount ($\rho_a gh$, where ρ_a is the density of air) from that in the free water surface, P_2, so that Eq.3 describes the lowering of the hydrostatic pressure across the curved air-water meniscus. The amount of pressure lowering is commonly referred to as suction. The retention of water between soil particles can be thought of as a capillary phenomenon in which the soil particles are usually assumed to exhibit a contact angle $\alpha = 0$.
The lowering of the hydrostatic pressure apparently increases with a decrease in diameter of the pore (Eq.3). In other words, water is more tightly held in narrow pores than in wide ones.
Physical quantities and their dimensions are given in Table 1. The units are presented in the cm-gram-sec (c.g.s.) and the meter-kilogram-sec (m.k.s.) systems. The former is used more frequently but the latter is recommended by the International Society of Physics and leads to more practical numerical values.

Table 1. Physical units.

Symbol	Physical quantity	Dimensions	c.g.s. units	m.k.s. units
L	length	L	centimeter (cm)	meter
M	mass	M	gram (g)	kilogram (kg)
t	time	T	sec	sec
A	area	L^2	cm^2	m^2
V	volume	L^3	cm^3	m^3
$\bar{V}$	specific volume	L^3M^{-1}	cm^3/g	m^3/kg
ρ	density	ML^{-3}	g/cm^3	kg/m^3
v	velocity	LT^{-1}	cm/sec	m/sec
a	acceleration	LT^{-2}	cm/sec^2	m/sec^2
F	force	$M\ L\ T^{-2}$	dyne = $g.cm/sec^2$	newton = $kg.m/sec^2$
F_k	specific force	LT^{-2}	cm/sec^2	
W	work	$M\ L^2T^{-2}$	erg = dyne.cm	joule = newton.m
E	energy	ML^2T^{-2}	erg	joule
ϕ	specific energy (potential)	L^2T^{-2}	erg/g	joule/kg
ϕ'	pressure	$M\ L^{-1}T^{-2}$	$dyne/cm^2$	$newton/m^2$

<u>Some additional units</u>:

g = acceleration of earth's gravitational field = 980 cm/sec^2

σ = surface energy or surface tension equals potential energy per unit surface area of liquid MT^{-2} (for water σ = 72.7 $dyne.cm^{-1}$ at 10^oC)

η = viscosity = tangential force on 1 cm^2 of liquid exerted by a gradient of velocity in normal direction of 1 cm/sec per cm (units poise = $1g.cm^{-1}.sec^{-1}$; η of water at 20^oC is about 0.01 poise = 1 centipoise)

R = gas constant = 8.317×10^7 erg/degree/mole; RT = 25×10^9 erg/mole, at 300^oK or 27^oC.

5.2.3 RETENTION CURVES

The forces of retention mentioned so far, i.e. the adsorption forces and the adhesion-cohesion forces resulting in concave air-water interfaces, are taken together as matric forces, because both types arise from the presence of the soil matrix. The existence of matric forces can be demonstrated by means of a tensiometer (Fig.2).

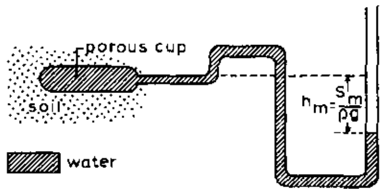

Fig.2. Tensiometer.

A tensiometer is a piezometer adapted for the measurement of negative pressures (suctions). Water inside the porous cup of the tensiometer attains equilibrium with the water in the surrounding soil. The matric suction, S_m, is thus given by the position of the water level in the open end of the manometer. For convenience, a Bourdon manometer or a mercury manometer may be connected to the open end of the tensiometer.

The matric suction is a function of soil water content. This can clearly be seen from Eq.3 as applied to soil-water retention, since this relationship implies that the greater the lowering in hydrostatic pressure, the smaller the radius of the water-filled pores will be. Or, in other words, the greater the applied suction, the lower the resulting water content will be.

This directly indicates the two main ways in which the interdependence of water content and suction can be determined:

- by applying suction to a porous plate on which moist soil is placed,
- by applying excess pressure to moist soil on a semipermeable membrane through which water, but not soil particles, can pass.

Details about the determination of the retention curves, i.e. the graphical representation of the function relating matric suction to soil water content, are given in Chap.23, Vol.III. If the suction is expressed in cm water and plotted on a logarithmic scale against the volumetric water content, the water retention curve is called pF curve.

Soil water retention curves are used to indicate the amount of water that can be retained by the soil and that is available for plant growth. This amount of water was defined earlier as the difference in water content at field capacity and at the permanent wilting point. For convenience, field capacity is then usually defined as the amount of water retained by the soil against a suction of 100 cm, and permanent wilting point as the amount of water retained against a suction of 15 atmospheres.

Although it is realized that these definitions are somewhat arbitrary in terms of the actual amount of available moisture, they still provide us with a suitable criterion for comparing the water-holding characteristics of different soils. In this way the water retention curve (pF curve) becomes a valuable tool in irrigation. Water retention curves are also used in determining the amount of water that is released or taken up by the soil when the water table drops or rises. This is of importance in the design of drainage systems (Chap.8, Vol.II).

5.3 FORCES AND POTENTIALS

5.3.1 THE CONCEPT OF POTENTIAL

BUCKINGHAM (1907) introduced the idea that the flow of water through the soil could be compared to the flow of heat through a metal bar or the flow of electricity through a wire. The driving force comparable to the temperature difference or the electrical potential was visualized as the difference in attraction for water (retention) between two portions of the soil which are not equally moist.

This may be easy enough when there is only one force acting on the water. It would then immediately be obvious whether the water was flowing and, if so, in what direction. However, besides the matric forces there are two types of forces acting on the soil water. These are: osmotic forces(caused by dissolved solutes) and body forces (inertial forces and gravitational force).

When it is desired to know differences in retention in order to obtain the direction of flow or to establish whether an equilibrium situation exists, it is necessary to obtain the vector sum of all forces acting on the water. This means that it is necessary to calculate the magnitude of the residual force and its direction, which in turn requires that the magnitude and direction of each force acting on the water are known. The equilibrium condition, of course, is that the residual force is zero, i.e. the vector sum of all component forces equals zero. It is, however, far more convenient to assign a corresponding potential to each force field.

The potential of water is defined as the work required to transfer a unit quantity of water from a standard reference state, where the potential is taken zero, to the situation where the potential has the defined value.

Without going into great detail, potential can be understood to give an indication of the energy status and hence the availability of soil water. The lower the potential, the lower is the availability of the water.

The unit quantity in the definition of potential may be unit mass, unit volume, or unit weight. In all cases only the differences in potential are significant and not the absolute value of the energy, even assuming this can be defined.

Potentials are scalars, not vectors (i.e. they have magnitude only and no direction) and thus the equilibrium condition is reduced to the requirement that the algebraic sum of these component potentials is constant. This sum of the component potentials is called the total potential. The driving force for the movement of water is then simply the gradient of total potential.

Potentials can be expressed in terms of energy per unit mass ϕ (i.e. in the c.g.s. system erg/g). It is often more useful to express potentials as energy per unit volume ϕ', which term can directly be reduced to a pressure term (i.e. erg/cm^3 or dyne/cm^2).

The conversion between the energy per unit mass and energy per unit volume term makes use of the specific volume (volume per unit mass, $\overline{V}$) or density (mass per unit volume, ρ) of water ($\phi' = \rho\phi$).

Finally, potentials can also be expressed on a weight basis, h : if the potential equals ϕ erg/g (or $\rho\phi$ ergs/cm^3), then on a weight basis it is equal to $h = \phi/g$ cm.

The dimensions of the latter are obtained as follows (see Table 1):

$$|\text{erg g}^{-1}/\text{cm sec}^{-2}| = \frac{L^2T^{-2}}{LT^{-2}} = L$$

Units and conversion factors for water potential are listed in Table 2.
In the c.g.s. system the factor ρg for water is close to 1000, with slight variations due to differences in temperature or salt content (Chap.6, Vol.I). The potential expressed on a mass basis and on a volume basis ($\phi' = \rho gh$ in dynes/cm^2) have therefore the same numerical values. It is the convention to assign to free and pure liquid water a potential value of 0. Since soil water

is less available, i.e. it has a restricted energy status as compared with the reference pure liquid water, the value of its potential is negative.

Table 2. Conversion of energy units.

specific energy units		pressure units		
Erg/gram	joule/kg	bar	atmosphere	cm of water
1	1×10^{-4}	1×10^{-6}	0.99×10^{-6}	1.02×10^{-3}
1×10^{4}	1	1×10^{-2}	0.99×10^{-2}	1.02×10^{1}
1×10^{6}	1×10^{2}	1	0.99	1.02×10^{3}
1.01×10^{6}	1.01×10^{2}	1.01	1	1.03×10^{3}
0.98×10^{3}	0.98×10^{-1}	0.98×10^{-3}	0.97×10^{-3}	1

Note: 1 joule = 10^7 erg
1 gram.cal = 4.186 joule
1 cm mercury (Hg) = 13.6 cm H_2O
1 atmosphere = 14.7 lbs/inch^2 (psi)
1 bar = 10^6 dynes/cm^2

Total potential, ϕ, is thus defined as the work necessary to move a unit quantity of water (e.g. 1 gram) from a chosen standard system with free and pure liquid water to the point under consideration in the soil:

$$\phi = \Sigma_k(- \int \vec{F}_k \, d\vec{s}) \tag{4}$$

where the summation is performed over all component potentials resulting from forces $\vec{F}_k$, expressed as force per gram mass, and $d\vec{s}$ is the displacement in the force field (with reference to a chosen standard location). The arrows indicate that these quantities are vectors. The following component potentials are considered in the next section:
- gravitational potential, ϕ_g
- matric potential, ϕ_m
- osmotic potential, ϕ_o

Under conditions where additional force fields are imposed on the soil, other component potentials should be included in the total potential. For example, soil in a pressure membrane apparatus is subjected to an additional external pressure (Chap.23, Vol.III), which would lead to a so-called external pressure potential. Likewise soil in a centrifuge would experience an external pressure potential resulting from the centrifugal force field.

5.3.2 COMPONENT AND TOTAL SOIL WATER POTENTIALS

the gravitational potential

The gravitational potential is defined by:

$$\phi_g = -\int_{z_o}^{z_1} \vec{g}\,d\vec{z} \qquad (5)$$

The integration is performed from the reference position z_o to the point under consideration z_1, where z is distance above the reference level. How does the gravitational potential depend on location in the profile? We take a column of an "ideal" soil, in which water is not subjected to osmotic forces. The water in the column is in equilibrium with a free water table. No evaporation occurs; hence the water in the soil is in a state of equilibrium. We now move an infinitesimal mass of water dM from z to a point z + dz, where dz is an infinitesimal distance. The work done by the applied force against the retaining forces in removing water from the substance at z is $\phi_g dM$. The work done by the applied force as water unites with the soil at the new level z + dz where the potential is $\phi_g + \frac{\partial \phi_g}{\partial z} dz$ amounts to $-(\phi_g + \frac{\partial \phi_g}{\partial z})\,dM$.
Hence the net work done by the applied force is

$$W_1 = -\frac{\partial \phi_g}{\partial z}\,dz\,dM \qquad (6)$$

Moving the mass of water from z to z + dz involves working against gravity:

$$W_2 = g dM\,dz \qquad (7)$$

At equilibrium

$$W_T = 0 = W_1 + W_2$$

which leads to

$$\frac{\partial \phi_g}{\partial z}\,dz\,dM = g dM\,dz = 0 \qquad (8)$$

or

$$g = \frac{\partial \phi_g}{\partial z} \qquad (9)$$

Integration of Eq.9 gives

$$\phi_g = gz + B \tag{10}$$

where B is the integration constant, which we may evaluate as follows: at the water table z = 0 and $\phi_g = 0$ by choice of our reference level; inserting these values in Eq.10 gives B = 0.
Hence we have

$$\phi_g = gz \tag{11}$$

which states that the gravitational potential is proportional to the height above the reference level where $\phi_g = 0$, which in this example was chosen at the water table.
Expressing the gravitational potential per unit weight (h_g), it is found that $h_g = z$; per unit volume (pressure potential ϕ') it gives $\phi'_g = \rho gz$.

<u>the matric potential</u>

The matric potential is formally defined as

$$\phi_m = - \int_{x_o}^{x_1} \vec{F}_m d\vec{x} \tag{12}$$

Since the matric forces are not known quantitatively, their total effect is derived from the value of the matric suction, S_m in dyne/cm^2 or h_m in cm water column, which can be measured experimentally. The matric potential is then obtained by multiplying S_m with $- \overline{V} = \frac{1}{\rho}$ (the specific volume) or h_m with $- g$ (the acceleration due to gravity).

Hence

$$\phi_m = - \frac{1}{\rho} S_m = - gh_m \tag{13}$$

Above the groundwater table soil water is "under suction", i.e. the matric forces have the effect of a negative hydrostatic pressure.
Below the water table the hydrostatic pressure in positive. The corresponding pressure potential ϕ_p is then equal to:

$$\phi_p = \frac{1}{\rho} \phi'_p = gh_p$$

where ϕ'_p is the hydrostatic pressure in dyne/cm^2 and h_p is the hydrostatic pressure in cm water column.

the osmotic potential

The osmotic potential is formally defined as:

$$\phi_o = - \int_{x_o}^{x_1} \vec{F}_o d\vec{x} \tag{14}$$

Dissolved solutes in the water cause an osmotic pressure, Π, which has the effect of a negative hydrostatic pressure. Like the value of the matric potential, the value of the osmotic potential is obtained from the product of the corresponding pressure and the reciprocal of the density:

$$\phi_o = - \frac{1}{\rho} \Pi \quad (\mathrm{erg/g}) \tag{15}$$

It can also be calculated directly from the concentration and the dissociation constant:

$$\phi_o = - \rho g R T d C_o \quad (\mathrm{erg/g}) \tag{16}$$

where

RT is the product of the universal gasconstant and the absolute temperature (25×10^9 erg/mole at about 25^oC),

d is the factor by which the number of dissolved particles is increased by dissociation,

C_o is concentration in mole/g of water.

For example, if the concentration of the dissolved sa ts in the soil water present in the root zone equals 10^{-2} molar and the salt is mainly NaCl, its osmotic potential can be calculated by means of Eq.16, as follows:

$$\phi_o = - (25 \times 10^9 \times 2 \times 10^{-5}) = - 5 \times 10^5 \text{ erg/g.}$$

Differences in osmotic potential only play a role in causing movement of water when there is an effective barrier for salt movement between the two locations at which the difference ϕ_o was observed. Otherwise, the concentration of salts will become the same throughout the profiles by the process of diffusion, and the difference in ϕ_o will no longer exist. Barriers for the movement of salts are formed by the surface of roots, while a densely compacted clay layer may also serve as a somewhat imperfect semi-permeable membrane (i.e. a membrane permeable to water but not to salt).

the total potential and hydraulic potential

The total potential is obtained by combining the relevant component potentials:

$$\phi = \phi_g + \phi_m + \phi_o + \phi_{(external)} \tag{17}$$

Equilibrium, which is defined as the situation where mass transfer of water in the liquid phase is absent, is obtained when the value of the total potential is constant. Usually, sufficient condition is that the sum of the component potentials; ϕ_o being ignored, is constant. The equilibrium condition states then that in the absence of external force fields, i.e. in the absence of an external gas pressure, different from atmospheric pressure,

$$\phi_h = \phi_g + \phi_m = \text{constant} \tag{18}$$

or, for saturated conditions

$$\phi_h = \phi_g + \phi_p = \text{constant}$$

where ϕ_h is called the hydraulic potential.

In Fig.3 this condition is applied to a vertical soil column in equilibrium with a water table. No water movement occurs in the column. The groundwater table is taken as the reference level for the gravitational potential.

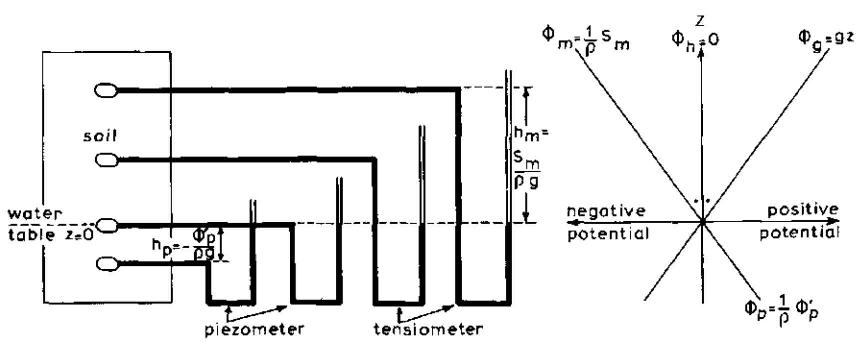

Fig.3. Equilibrium soil water condition $\phi_h = 0$.

Two piezometers (one at the water table) indicate values of the hydrostatic pressure and two tensiometers values of the matric suction. Since the matric potential is zero at the water table, it is found that the matric potential balances the gravitational potential throughout the profile. Apparently in the equilibrium situation a plot of ϕ_h as a function of position in the profile (z) is a straight line through the origin.

$\phi_h(z)$ functions can also be determined for non-equilibrium situations from soil water suction data obtained with tensiometers installed at different depths in the profile. For example, a tensiometer located at a depth of 50 cm showed a suction of 60 cm water; another at a depth of 75 cm showed a suction of 40 cm. Will water move upwards or downwards in the profile? If we take the position of the lower tensiometer as reference level for the gravitational potential we find that its value for the tensiometer at 50 cm depth, expressed on a weight basis, equals 25 cm; hence the hydraulic potential for that tensiometer, also on a weight basis, equals 25 - 60 = -35 cm.
For the lower tensiometer, the hydraulic potential equals 0 - 40 = -40 cm. The direction of flow is towards the position with the lowest water potential and is therefore downwards.

A different situation is depicted in Fig.4. Tensiometer readings were obtained at several depths in the profile. It was found that below a depth of 15 cm, the tensiometers indicated the same value of 100 cm suction (field capacity) while at a shallower depth the soil was drying under the influence of evaporation. The lowest tensiometer was again taken as reference level of the gravitational potential. Note that the value of the hydraulic potential calculated according to Eq.18 is not the same throughout the profile. Above a depth of 15 cm the direction of water movement is upwards, and below 15 cm water tends to move downwards. Whether water actually moves downwards depends on the conductivity of the unsaturated soil for water movement (see Sect.4 and Fig.11).

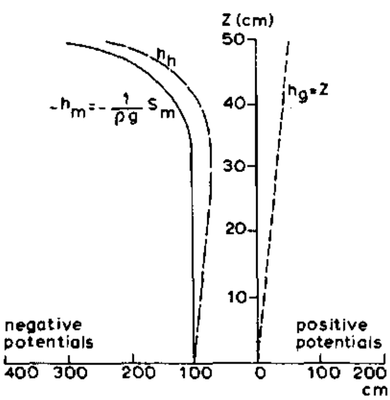

Fig.4. Soil water potential profiles with part of the soil at field capacity.

In Fig.5 a number of possible hydraulic potential profiles is shown. Experimentally determined $\phi_h(z)$ curves, as for example the curve in Fig.4, can be broken up into sections with shapes similar to those in Fig.5. From the shape and curvature of the $\phi_h(z)$ curve it is possible to determine the direction of flow (since water moves from a position of high hydraulic potential to a position of low hydraulic potential) and to assess likely changes in water content. These are also indicated in Fig.5. Obviously it is necessary to install more than two tensiometers in the soil to detect a curvature in the ϕ_h profiles. The position of the $\phi_h(z)$ curve with respect to the zero mark on the x-axis is immaterial and depends only on the reference level for the gravitational potential. Positive values of ϕ_h do not necessarily indicate saturated flow as can be seen in Fig.6. The slope of the curve indicates the direction of flow.

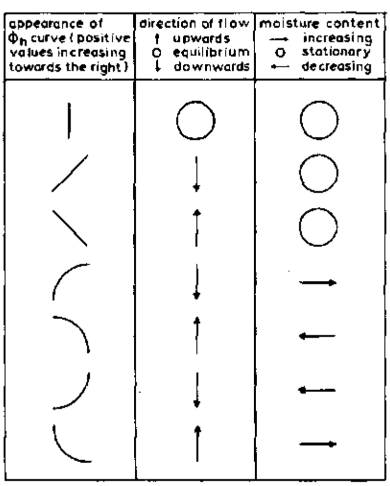

Fig.5. Direction of flow and change in soil water content for various ϕ_h profiles.

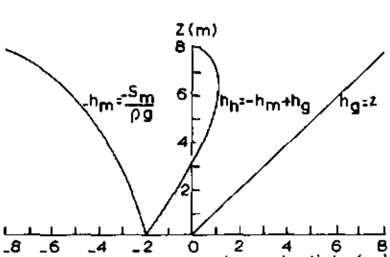

Fig.6. Soil moisture potential profiles with positive values of hydraulic potential h_h(m).

5.4 TRANSPORT OF WATER IN THE SOIL

5.4.1 TRANSPORT EQUATIONS BASED ON THE POTENTIAL CONCEPT

The concept of potentials is well suited for the analysis of flow of water in soils, since all flow is a consequence of potential gradients. Hence we can write as one general flow equation:

$$v = -k \frac{d\rho\phi_h}{ds} \qquad (19)$$

where (dimensions in the c.g.s. system):

v = flow velocity (cm/sec)

k = conductivity coefficient (dimensions will be determined below).

$\frac{d\rho\phi_h}{ds}$ = hydraulic potential gradient (dyne/cm^3) in which the hydraulic potential $\rho\phi_h = \phi_h'$ is expressed in erg/cm^3 or dyne/cm^2 (the hydraulic pressure).

The minus sign in Eq.19 indicates that the flow is in the direction of decreasing potential.

The units of k can be found from the dimensions of $d\rho\phi_h/ds$ (dyne/cm^3) and v(cm/sec) as cm^4/dyne.sec or cm^3 sec/g. The flow of water through saturated soils, v, is generally described by Darcy's law:

$$v = -Ki \qquad (20)$$

where i is the hydraulic gradient, and K is the constant hydraulic conductivity (see Chap.6, Vol.I). The hydraulic gradient is the loss of hydraulic head, h, (i.e. the hydraulic potential expressed on a weight basis, hence it has the dimensions of a length) over a unit length of flow path, s, ($i = dh/ds$). The hydraulic gradient is therefore a dimensionless unity, so that v and K have the same dimensions (cm/sec).

Comparing Eqs.19 and 20, one finds that $K = \rho g k \simeq 1000\ k$.

Equations 19 and 20 can also be used for the analysis of flow through unsaturated soil when it is understood that k, now called the capillary conductivity with symbol k_w, then depends not only on the pore geometry of the soil but also on

the water content. It should also be realized that ϕ_h' in saturated flow systems refers to the positive hydrostatic pressure, and in unsaturated flow systems to the negative hydraulic pressure with its two components, i.e. those due to matric forces and gravity.
The general flow equation for unsaturated flow

$$v = - k_w \frac{d\phi_h'}{dx} \tag{21}$$

expresses the relationship between the three variables, v, k_w, and ϕ_h'; for its solution two additional equations incorporating the same variables are required.

The first additional equation is obtained from the principle of conservation of matter. This equation is called the equation of continuity. Let us consider a small cube of soil with length of sides Δx, Δy, and Δz (see Fig.7).

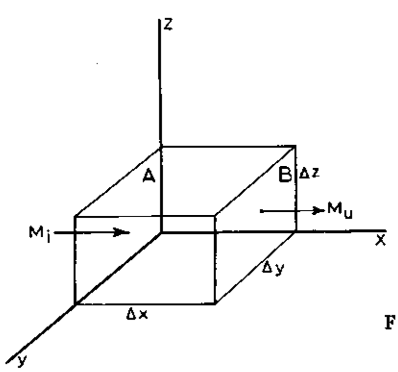

Fig.7. Model for the derivation of the continuity equation.

A mass of water, M_i, flowing into the cube in the x-direction, through face $\Delta y \Delta z$ at side A is given by

$$\frac{\partial M_i}{\partial t} = \rho v_x \Delta y \ \Delta z \tag{22}$$

where v_x is the flux of water in the x-direction. A mass M_u flowing out at side B is given by

$$\frac{\partial M_u}{\partial t} = (\rho v_x + \frac{\partial \rho v_x}{\partial x} \Delta x) \Delta y \ \Delta z \tag{23}$$

The net flux leaving or accumulating is the difference between Eqs.22 and 23.

The net change in flux of water moving through the volume of soil in all directions is then given by:

$$\frac{\partial M}{\partial t} = - (\frac{\partial \rho v_x}{\partial x} + \frac{\partial \rho v_y}{\partial y} + \frac{\partial \rho v_z}{\partial z}) \Delta x \; \Delta y \; \Delta z \tag{24}$$

M, the mass of water in the volume $\Delta x \; \Delta y \; \Delta z$ is given by the product of the bulk density B, the mass fraction of water in the soil, θ, and the volume of the cube:

$$M = B\theta \; \Delta x \; \Delta y \; \Delta z \tag{25}$$

If Eq.25 is differentiated and combined with Eq.24 we get:

$$\frac{\partial B\theta}{\partial t} = - (\frac{\partial \rho v_x}{\partial x} + \frac{\partial \rho v_y}{\partial y} + \frac{\partial \rho v_z}{\partial z}) \tag{26}$$

If the densities B and ρ are constant, Eq.26 can be simplified and written for flow in the x-direction only as:

$$\frac{B\partial\theta}{\rho\partial t} = \frac{\partial w}{\partial t} = - \frac{\partial v_x}{\partial x} \tag{27}$$

where $B\theta/\rho$ is equal to the volumetric water constant, w.

The second additional relationship that is required for the solution of Eq.21 is found in the water retention curve. Sometimes the graphical representation of the water retention curve can be expressed by simple expressions like $w = a + b\phi'_m$, or $w = a(\phi'_m)^b$ which are sufficiently accurate over a limited range. These expressions are usually not satisfactory when hysteresis effects are considerable.

Equations 21 and 27 can be combined to yield for flow in the x-direction

$$(\frac{\partial w}{\partial t})_x = \frac{\partial}{\partial x} \; (k_w \frac{\partial \phi'_h}{\partial x}) \tag{28}$$

or with $\phi'_h = - S_m + \rho g z$:

$$(\frac{\partial w}{\partial t})_x = \frac{\partial}{\partial x} (- k_w \frac{\partial Sm}{\partial x} + k_w \rho g \frac{\partial z}{\partial x}) \tag{29}$$

For horizontal flow, vertical flow upward, and vertical flow downward, the value of $\partial z/\partial x$ is 0, 1, -1, respectively. In vertical flow the influence of the gravitational force is often rather small in comparison with that of the matric forces, in which case the last term of Eq.29 may be neglected. This equation can further be simplified by the introduction of the soil water diffusivity, D_w, also a function of water content, w, defined as:

$$D_w = - k_w \frac{\partial Sm}{\partial w}$$

Eq.29 then becomes:

$$\left(\frac{\partial w}{\partial t}\right)_x = \frac{\partial}{\partial x} \left(D_w \frac{\partial w}{\partial x}\right) \qquad (30)$$

Let us illustrate some of the principles of transport of water through soils with the following example: In a root zone in a sandy soil a tensiometer measured a suction of 800 cm H_2O. The concentration of dissolved salts, mainly NaCl, in the water present in the root zone was 10^{-2} molar. The groundwater table was at 100 cm below the root zone. The concentration of the dissolved salts in the groundwater was 5×10^{-2} molar. If we take the groundwater table as reference level for the gravitational potential, we find for the component potentials in, respectively, the root zone and at the groundwater table the following values:

Potential (erg/g)	Root zone	Groundwater table
ϕ_o (osmotic potential)	$- 5 \times 10^5$	$- 25 \times 10^5$
ϕ_g (gravitational potential)	$+ 1 \times 10^5$	0
ϕ_m (matric potential	$- 8 \times 10^5$	0

The total potential in the groundwater ($- 25 \times 10^5$ erg/g) is lower than in the root zone ($- 5 \times 10^5$ erg/g). Nevertheless, water will move from the groundwater table upwards into the root zone because the hydraulic potential is lower in the root zone than in the groundwater ($- 7 \times 10^5$ and 0 erg/g respectively). If we assume an average value of the unsaturated conductivity, k_w, for the soil layer between root zone and groundwater table, equal to 10^{-10} cm^{3}sec/g ($\simeq 10^{-2}$ cm/day), we find that this hydraulic potential gradient can maintain, according to Eq.21, an upward flow of:

$$v = - k_w \frac{\rho \Delta\phi}{\Delta z} = + \frac{10^{-5} \times 7 \times 10^5}{10^2} = 7 \times 10^{-7} \text{ cm/sec} \simeq 0.07 \text{ cm/day}$$

5.4.2 INFILTRATION

The solution of Eq.30 is well known from heat problems for the situation in which a constant value of the diffusivity, not depending on the water content, can be assumed. Unfortunately, however, D_w for soil water flow depends greatly on w. Nevertheless, Eq.30 can be solved for certain boundary conditions.

For example, BRUCE and KLUTE (1956) solved Eq.30 for horizontal infiltration of water into a column of soil, if the water content at the inflow boundary remained constant. The boundary conditions are:

$$
\begin{array}{lll}
t = 0 & x > 0 & w = w_n \\
t > 0 & x = 0 & w = w_o
\end{array}
\qquad (31)
$$

where w_n is the initial water content and w_o the water content at the inflow boundary. These boundary conditions can be satisfied in an experimental set-up shows in Fig.8.

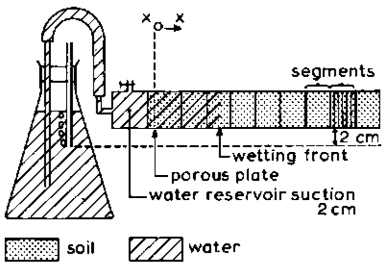

Fig.8. Experimental set-up for horizontal infiltration.

Equation 30 can now be solved under the additional condition that a plane of constant water content advances proportionally to the square root of the infiltration time. That this last condition is satisfied can easily be verified by plotting the advance of the wetting front (a plane of constant water content) as a function of the square root of infiltration time. Usually this plot is a straight line unless the soil exhibits considerable swelling when wetted.

The solution of Eq.30 under the conditions of Eq.31 is:

$$D(w) = - \frac{1}{2t} \frac{dx}{dw} \int_{w_n}^{w} x dw \qquad (32)$$

where t is the total time of infiltration.

The information necessary for the solution of Eq.32 can be obtained from the plot of w(x) at the end of infiltration (Fig.9), since $\frac{dx}{dw}$ is the slope of the curve relating x with w, and $\int$ xdw can be determined, for each value of w, from the shaded area below that particular value of w (see Fig.9), k_w can then be determined according to Eq.29 from the value D_w for each value of w and the slope of the water retention curve (dw/dS_m) at the same value of w. Resulting curves for k_w and D_w as functions of w are given in Fig.10.

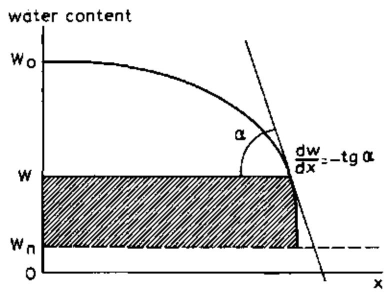

Fig.9. Water content distribution at end of infiltration test with set-up of Fig.8.

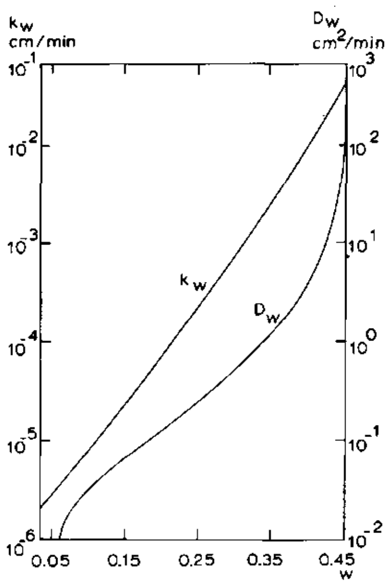

Fig.10. Experimental values of k_w and D as functions of w obtained from the w(x) function of Fig.9. Columbia silt loam (c.f. Davidson et al., 1963).

The unsaturated conductivity coefficient can also be expressed as a function of S_m through the water retention curve. Examples for soils of different textures are given in Fig.11. From the figure it appears that coarse textured material has a greater capillary conductivity that the finer textured soils at a low value of the suction, while at a greater suction (lower water potential) the situation is reversed.

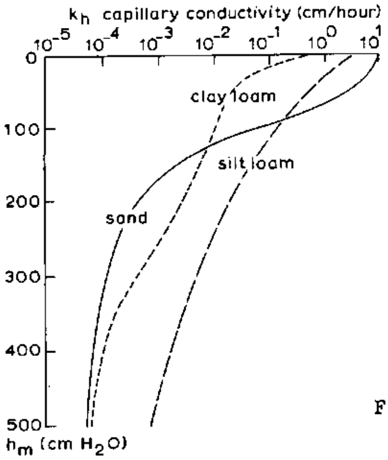

Fig.11. Relation between capillary conductivity k_h and matric suction h_m.

The practical implication is that redistribution of water in the profile, after infiltration has stopped, is considerably faster in fine textured soils than in coarse sands. The wetting front penetrates readily into a sandy soil as long as there is water infiltrating into the soil, but soon after infiltration ceases the movement of the wetting front will stop. This is shown in Fig.12. The experimental results of Fig.12 were obtained with gentle flooding of the surface with a head of water not exceeding 1 cm.

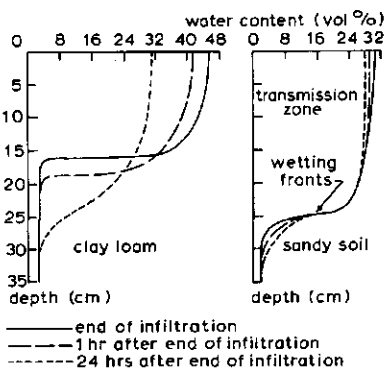

Fig.12. Infiltration and redistribution of soil water in a clay loam and sandy soil.

Often soil water profiles, for both vertical and horizontal flow, have sharp increases in water content near the infiltration boundary x = 0. As a result of the experimental procedure the water content in the saturated zone near the surface is higher than in the transmission zone, which is the zone of near constant water content lower in the profile. The value of the water content in this transmission zone decreases as the rate of water entering the soil decreases, and it is highest in the case of flood infiltration. This can be seen in Fig.13, where water content profiles are shown after 8 cm of water has infiltrated into a clay loam soil by three different methods of wetting: gentle flooding, and raining with two intensities, 1 cm/hr and 0.1 cm/hr. The rate of flow in the rain treatments is approximately equal to the capillary conductivity at the water content of the constant (transmission) region.

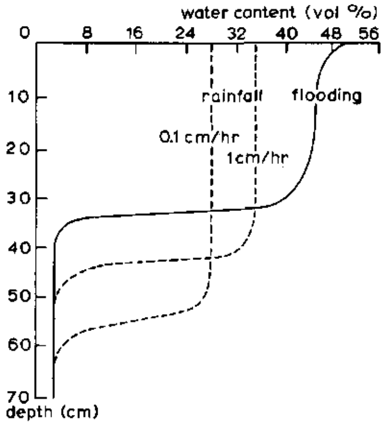

Fig.13. Infiltration into a clay loam with three different wetting methods.

Infiltration into stratified soil, i.e. when discontinuities in the conductivity function occur with depth in the profile, still largely escape our attempts at mathematical expression. An example of measured water content and suction profiles is presented in Fig.14. It clearly shows discontinuities in water content at the boundary between coarse and fine soil whereas the suction profiles are continuous functions across the boundary.

Empirically derived relations for the evaluation of infiltration are found in Chapter 15 (Vol.II).

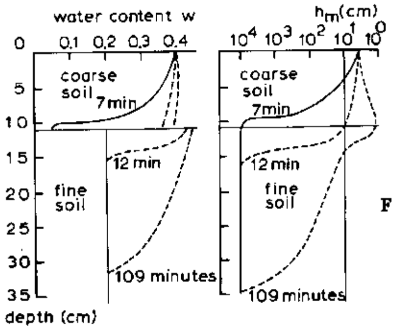

Fig.14. Water content and suction profiles at different stages of infiltration into stratified soil (c.f. Hanks and Bowers, 1962).

5.4.3 CAPILLARY RISE

The height to which water will rise from the groundwater table can be predicted for the situation in which water is in equilibrium with the groundwater and no movement occurs as a result of uptake by roots or evaporation. By analogy with the rise of water in glass capillaries, one obtains for h, the rise of water from the groundtable, for a contact angle of $\alpha = 0$ (see Sect.2.2):

$$h = \frac{2\sigma}{r\rho g} \tag{32}$$

in which r is now the effective pore radius of the soil.
The value of r cannot be determined independently so that r is usually evaluated from the measurement of h, the observed capillary rise. However, in many soils, and certainly in clay soils, the maximum h value is not always realized because of low conductivity characteristics in the soil. This could result in misleading values of r. Sometimes it is possible to assume a stationary flow, v, upwards from the groundwater table as a result of (near) constant removal of water by evaporation and/or uptake by plant roots. This implies that $\delta w/\delta t = 0$ at every location, $z \geq 0$ in the profile ($z = 0$ at the level of the groundwater table). It also implies that $\partial v/\partial z = 0$ for $t \geq 0$.

Eq.21 can be written in terms of the matric and gravity potential gradient, both expressed on a weight basis:

$$v = k_h\left(\frac{dh_m}{dz} - 1\right) \tag{33}$$

where k_h is the capillary conductivity as function of h_m (as, for example, given in Fig.11).

Integration of Eq.33 leads to:

$$z = \int_{o}^{h_m} \frac{k_h}{k_h + v} dh_m \qquad (34)$$

This integral can be solved when k_h is known (Fig.11), or when this relation can be described by an empirical expression. GARDNER (1958) has proposed

$$k_h = \frac{a}{(h_m)^n + b} \qquad (35)$$

where a, b, and n are constants which would have to be determined experimentally.

The solution of Eq.34 is presented in Fig.15 for a coarse textured soil for several flow velocities. From the figure it can be seen that a velocity of 1 mm/day can be maintained to a root zone about 90 cm above the groundwater table when the pF in the root zone is greater than about 2.5.

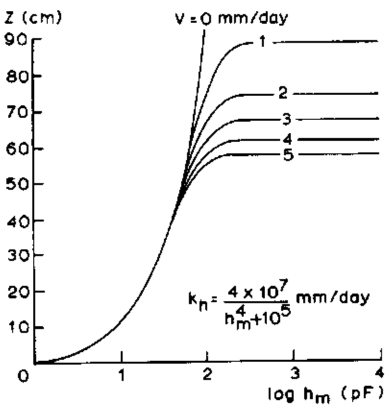

Fig.15. Potential profiles calculated for a coarse textured soil (m=4) under influence of capillary rise.

When the soil surface dries out, the capillary conductivity at the surface approaches a small constant value (i.e. the conductivity of the soil for water vapour). It appears then from Eq.33 that the evaporation rate, v, would approach a constant value. This value varies approximately with z_o^{-n}, where z_o is the depth of the groundwater table. However, it has been found that the evaporation rate goes through a maximum rather than approaching a constant value, with an increase in evaporative demand of the atmosphere. In Fig.16 the observed steady state evaporation is shown as a function of evaporative demand of the atmosphere for a fine sandy loam with two different depths of the groundwater table (HADAS and HILLEL, 1968).

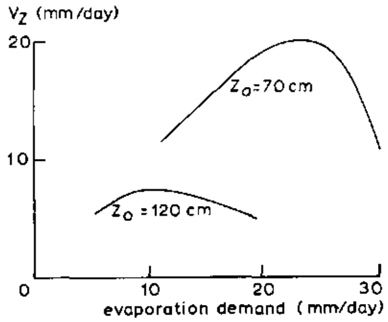

Fig.16. Relation of measured steady state evaporation rate to potential evaporativity for a fine sandy loam with water table at two depths.

It is obvious that the evaporation rate passes through a maximum value, which is higher for the shallow water table. The subsequent decrease in evaporation rate could not be due to salt accumulation, but might result from the occurrence of two layers in the profile with a large potential gradient between them, resulting from the drying process (HADAS and HILLEL, loc.cit.).

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INTRODUCTORY SUBJECTS

6. ELEMENTARY GROUNDWATER HYDRAULICS

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PURPOSE AND SCOPE

The fundamental physical laws governing the flow of groundwater, the basic hydraulic flow equations, and the underlying assumptions and approximate solutions as used in applied groundwater hydrology are discussed and illustrated by some examples.

CONTENTS

		page
6.1	Groundwater and water table defined	156
6.2	Physical properties, basic laws	156
6.2.1	Mass density of water	157
6.2.2	Viscosity of water	157
6.2.3	Specific weight	158
6.2.4	Law of conservation of energy	158
6.2.5	Potential of groundwater	160
6.2.6	Law of conservation of mass	161
6.3	Darcy's law	164
6.3.1	General formulation	164
6.3.2	The proportionality constant K in Darcy's law	166
6.3.3	Influence of temperature	168
6.3.4	Fresh-water head of salty groundwater	168
6.4	Some applications of Darcy's law	169
6.4.1	The falling head permeameter	169
6.4.2	Flow through stratified soils	171
6.5	Basic equations of groundwater flow	174
6.5.1	The Laplace equation for steady flow	174
6.5.2	The Dupuit-Forchheimer assumptions	176
6.5.3	Nonsteady flow	179
6.5.4	Replenishment	180
6.5.5	Flow in an unconfined layer between two water bodies	181
6.5.6	Steady flow to parallel ditches with a uniform recharge on the soil surface	182

		page
6.5.7	Steady flow towards a well	185
6.6	Some aspects of two-dimensional flow	187
6.6.1	Discontinuous hydraulic conductivity	187
6.6.2	Potential and stream functions	188
6.6.3	Exact methods of solution	192
6.6.4	Some other approximate methods of solution	193
6.7	Boundary conditions	195
6.7.1	Impervious boundaries	196
6.7.2	Planes of symmetry	196
6.7.3	Free water surface	196
6.7.4	Boundaries with water at rest or slowly moving water	197
6.7.5	Seepage surface	198
Literature		199

6.1 GROUNDWATER AND WATER TABLE DEFINED

The term groundwater refers to the body of water in the soil, all the pores of which are saturated with water. The locus of points in the groundwater body where the pressure is equal to the atmospheric pressure defines the phreatic surface, also called free water surface or groundwater table (Fig.1). It is found as the water level in an open bore hole that penetrates the saturated zone. Pressure is usually expressed as relative pressure p with reference to atmospheric pressure p_{atm}. At the groundwater table, by definition $p = p_{atm}$.

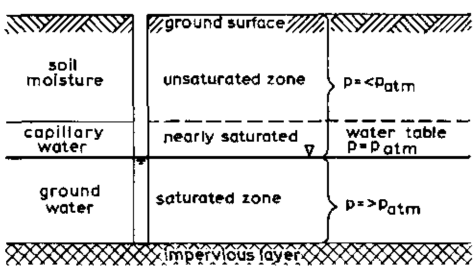

Fig.1. Scheme of the occurrence of water in the subsurface.

The mass of groundwater actually extends slightly above the groundwater table owing to capillary action, but the water is held there at less than atmospheric pressure. The zone in which capillary water fills nearly all pores of the soil is called the capillary fringe. Although occurring above the groundwater table, the water in the capillary fringe is sometimes included in the groundwater mass. The capillary water occurring above the capillary fringe belongs, together with the pendular water, to the unsaturated zone or zone of aeration in which the soil pores are filled partially with water and partially with air.

6.2 PHYSICAL PROPERTIES, BASIC LAWS

In drainage studies we are not only interested in the depth at which the groundwater table is found and the changes that occur in this depth; we are particularly concerned with the movement of the groundwater and its flow rate. This movement is governed by several well-known principles of hydrodynamics which, in fact, are nothing more than a reformulation of the corresponding principles of mechanics.
From a physical viewpoint, a complete system of hydrodynamics requires the form-

ulation of: the equation of continuity, the equation of state of the groundwater, and the dynamic equations of movement of the groundwater.
Since we are dealing with groundwater as a fluid, some of its physical properties and the basic laws related to its movement will be presented, after which the above-mentioned equations will be formulated.

6.2.1 MASS DENSITY OF WATER

The density of a material is defined as the mass per unit volume. The mass density may vary with pressure, temperature, and concentration of dissolved solids. For water, the mass density will be denoted by ρ. Its value is approximately 1000 kg/m^3 and, in the considerations that follow, it will be assumed a constant.

6.2.2 VISCOSITY OF WATER

In a state of laminar flow, i.e. a flow in which the path lines of water particles are parallel, a layer of fluid sliding over another exerts a frictional drag upon it, and this force is reciprocated. The faster-moving layer tends to drag the slower along with it; the slower tends to hold back the faster. This friction is called viscosity. The overall effect of the thin water layers, each of which is moving at a different speed from its neighbours, is observed as a velocity gradient in the direction y at right angles to the line of movement. At any given point where the velocity gradient is dv/dy, the viscous shear stress, F/A, in the plane at right angles to the direction of y is

$$F/A = \eta \, dv/dy \tag{1}$$

where η is the dynamic viscosity of the fluid. For water, the dynamic viscosity is approximately 10^{-3} kg/m s.

The kinematic viscosity ν is defined by the relation

$$\nu = \eta/\rho \tag{2}$$

For water, the kinematic viscosity is approximately 10^{-6} m^2/s.
The variation of viscosity and mass density of water with temperature is given in Table 1.

Table 1. Variation of mass density and viscosity of water with temperature.

Temperature ^{o}C	Mass density kg/m^3	Dynamic viscosity kg/m s	Kinematic viscosity m^2/s
0	999.87	1.79×10^{-3}	1.79×10^{-6}
5	999.99	1.52×10^{-3}	1.52×10^{-6}
10	999.73	1.31×10^{-3}	1.31×10^{-6}
15	999.13	1.14×10^{-3},	1.14×10^{-6}
20	998.23	1.01×10^{-3}	1.007×10^{-6}
25	997.07	0.89×10^{-3}	0.897×10^{-6}
30	995.67	0.80×10^{-3}	0.804×10^{-6}
40	992.24	0.65×10^{-3}	0.661×10^{-6}

6.2.3 SPECIFIC WEIGHT

The specific weight of water, γ , is obtained by multiplying its mass density by the acceleration of gravity, $g(= 9.81\ m/s^2)$

$$\gamma = \rho g \tag{3}$$

For water, the specific weight is approximately 9810 kg $m^{-2}s^{-2}$.

6.2.4 LAW OF CONSERVATION OF ENERGY

A fundamental law in hydrodynamics is the law of conservation of energy, which states that no energy can be created or destroyed in a closed system.
Assume a fluid particle moving during time Δt from point 1 to point 2 along a stream line of the fluid in the tube depicted in Fig.2. This fluid particle has the following three types of interchangeable energy per unit of vol·me

$1/2\rho v^2$ = kinetic energy per unit of volume
ρgz = potential energy per unit of volume
p = pressure energy per unit of volume

If, for the moment, it is further assumed that the flow tube of Fig.2 is not obstructed by solid material, there is no loss of energy due to friction. Since there is no gain of energy either, we may write

$$(1/2\rho v^2 + \rho gz + p)_1 = (1/2\rho v^2 + \rho gz + p)_2 = \text{constant} \tag{4}$$

This equation is only valid when a fluid particle is moving along a stream line under steady flow conditions, when the energy losses are negligible, and when the mass density of the fluid, ρ, is a constant.
Equation 4 is known as Bernoulli's equation, which in its general form is written

$$1/2\rho v^2 + \rho g z + p = \text{constant} \tag{5}$$

Since, in nature, velocities of groundwater flow (v) are usually low, the kinetic energy in Eq.5 may be neglected without any appreciable error. Hence, Eq.5 reduces to

$$\rho g z + p = \text{constant} \tag{6}$$

Since laminar flow was assumed (i.e. the stream lines are straight and parallel), the sum of the potential and pressure energies in the plane perpendicular to the direction of flow is constant. In other words: $\rho g z + p$ = constant for all points of the cross section.
In Eq.6 the energy is expressed per unit of volume. Expressing the energy per unit of weight, i.e. dividing by ρg, we convert the energy equation (6) into one of potential

$$\frac{p}{\rho g} + z = \text{constant} = h \tag{7}$$

where
$\frac{p}{\rho g}$ = the pressure head, z = the elevation head, and h = the hydraulic head.
The tube in Fig.2, in fact, is filled with sand, and a fluid particle travelling along a stream line has to overcome a resistance. In doing so, it loses energy, which is accounted for by a head loss Δh. For the example shown in Fig.2, Bernoulli's equation then reads

$$\frac{p_1}{\rho g} + z_1 = \frac{p_2}{\rho g} + z_2 + \Delta h_L \tag{8}$$

or

$$\Delta h_L = \left(\frac{p_1}{\rho g} + z_1\right) - \left(\frac{p_2}{\rho g} + z_2\right) \tag{9}$$

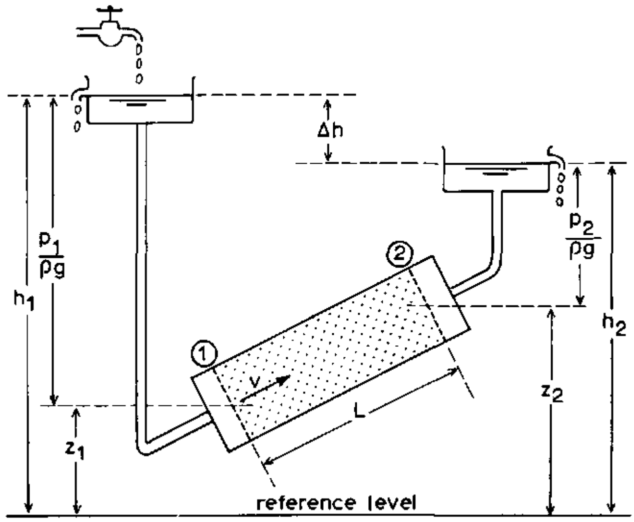

Fig.2. Pressure distribution and head loss in flow through a sand column.

The head loss may therefore be defined as the loss of potential and pressure energy per unit of weight when the fluid is moving from section 1 to section 2, the energy head being lost by frictional resistance.

6.2.5 POTENTIAL OF GROUNDWATER

The potential head or hydraulic head of groundwater at a certain point A is the elevation to which the water would rise in an open tube whose bottom end coincides with the point in question, the elevation being measured from an arbitrarily chosen plane of reference (Fig.3). It is made up of two items, namely pressure head, $p/\rho g$, and elevation head, z.
In the study of groundwater flow problems, it is common practice to express the potential and pressure energies on a unit weight basis (Chap.5, Vol.I), i.e. in length of water column, h. Hence

$$h = \frac{p}{\rho g} + z \qquad (10)$$

where
z = the elevation of the point under consideration above some plane of reference
p = the pressure of the groundwater at that point relative to some reference pressure
and the other symbols as defined earlier.

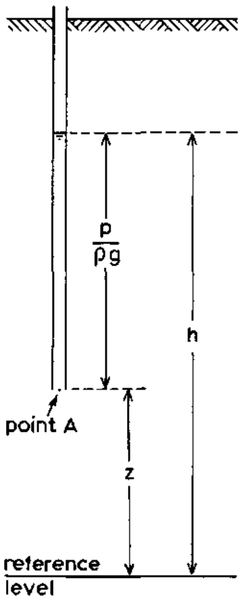

Fig.3. Potential, or hydraulic head, h at a point A located at a height z above a reference level.

Pressure is usually expressed as relative pressure, p, with reference to atmospheric pressure, p_{atm}. Thus in this context, p_{atm} equals zero pressure.
Mean sea level is sometimes used as reference level for height. But since for our purpose only relative values are used, the height component of the potential head is generally taken relative to an arbitrary level, for instance an impervious layer.

6.2.6 LAW OF CONSERVATION OF MASS

A second fundamental law in hydrodynamics is the law of conservation of mass, which states that in a closed system the fluid mass can be neither created nor destroyed.
A space element, dx dy dz, in which the fluid and the flow medium are both incompressible, will conserve its mass over a time dt. Therefore the fluid must enter the space element at the same rate (volume per unit time) as it leaves the element. The rate at which a volume is transferred across a section equals the product of the velocity component perpendicular to the section and the area of the section. If we assume a linear velocity distribution over the elementary distances dx, dy, and dz, we may write the average velocity components perpendicular to

the lateral faces of the space element as indicated in Fig.4.

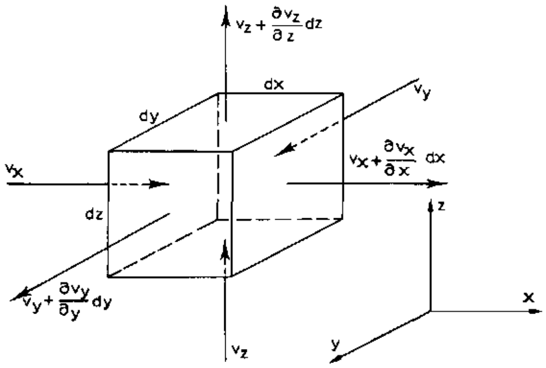

Fig.4.
Velocity distribution in a space element of fluid.

The difference of volume transferred per time dt in the x-direction equals

$$(v_x + \frac{\partial v_x}{\partial x} dx)\ dy\ dz\ dt - v_x dy\ dz\ dt \tag{11}$$

or

$$\frac{\partial v_x}{\partial x} dx\ dy\ dz\ dt \tag{12}$$

Analogous expressions can be derived for the differences of volume transferred per time dt in the y- and z-directions

$$\frac{\partial v_y}{\partial y} dy\ dx\ dz\ dt \tag{13}$$

and

$$\frac{\partial v_z}{\partial z} dz\ dx\ dy\ dt \tag{14}$$

According to the law of conservation of mass, the total difference of volume transferred in and out of the space element must equal zero. Hence

$$\frac{\partial v_x}{\partial x} dx\ dy\ dz\ dt + \frac{\partial v_y}{\partial y} dy\ dx\ dz\ dt + \frac{\partial v_z}{\partial z} dz\ dx\ dy\ dt = 0 \tag{15}$$

For time-independent flow this equation reduces to

$$\frac{\partial v_x}{\partial x} + \frac{\partial v_y}{\partial y} + \frac{\partial v_z}{\partial z} = 0 \tag{16}$$

which is the general form of the so-called continuity equation.

In fluid mechanics, it is common practice to select a coördinate system whose x-direction coincides with the direction of the flow vector at the point under consideration. In other words, the x-direction is parallel to the tangent of the path line at the considered point. Consequently $v_x = v$, $v_y = 0$, and $v_z = 0$. Since under these circumstances there is a transfer of volume in the x-direction only, the difference of volume transferred in this direction per time dt must equal zero. Hence

$$(v_x + \frac{dv_x}{dx} dx)\ dy\ dz\ dt - (v_x)\ dy\ dz\ dt = 0 \tag{17}$$

or

$$v_{x+dx}\ dA - v_x\ dA = 0$$

or

$$(vdA)_{x+dx} = (vdA)_x = dQ \tag{18}$$

Thus, the rate of flow dQ is a constant through two elementary cross-sections at infinitely short distance from each other. In fact, we considered an elementary part of a stream tube bounded by streamlines lying on the dx dy and dx dz planes. Equation 18 is valid for an elementary cross-sectional area of flow: $dA = dy\ dz$. If we now consider a finite area of flow, A, we may write the continuity equation as

$$Q = \int_A v.dA = \bar{v}A \tag{19}$$

where $\bar{v}$ is the average velocity component perpendicular to the cross-sectional area of flow under consideration.

6.3 DARCY'S LAW

6.3.1 GENERAL FORMULATION

The fundamental law describing the movement of groundwater through a soil was given by Darcy in 1856. The experiments that Darcy performed were of the type shown in Fig.2.

Darcy observed that the amount of water flowing through the sand sample in unit time (or in other words the rate of flow or the discharge) was proportional to the difference Δh between the fluid heads at the inlet and outlet faces of the sample (head loss $h_1 - h_2 = \Delta h$), and inversely proportional to the length of the sand sample (the flow path). This proportionality can be expressed mathematically as follows

$$Q = K \frac{\Delta h}{L} A \qquad (20)$$

where

Q = the rate of flow through the sample (L^3T^{-1})

Δh = the head loss (L)

L = the length of the sample (L)

A = the cross-sectional area of the tube (L^2)

K = a proportionality constant, depending of the nature of the sand and the fluid (water), (LT^{-1})

The quantity Q/A represents the discharge or flow rate per unit of cross-sectional area and is called apparent velocity, sometimes also called effective flow velocity or specific discharge. It is denoted by the symbol v. Hence

$$v = Q/A \qquad (21)$$

which is the continuity equation, see Eq.19.

The term $\Delta h/L$ represents the loss of head per unit length of flow path and is called gradient of hydraulic head. Denoting this hydraulic gradient by i and substituting it into Eq.20 yields what is known as Darcy's law or law of linear resistance in analogy with Ohm's law in electricity

$$v = -Ki \qquad (22)$$

Darcy's law states that the apparent velocity is directly proportional to the derivative of the hydraulic head in the direction of flow. The negative sign indicates that the flow is in the direction of decreasing head.
The dimension of v is (LT^{-1}), while i is dimensionless. Hence, the dimension of K is that of velocity (LT^{-1}).
The proportionality constant K is known as the coefficient of permeability or, preferably, hydraulic conductivity.
It should be noted that the flow velocity in the individual pores of the soil greatly exceeds the apparent velocity, which, in fact, is a hypothetical velocity that the water would have if flowing through the given flow column quite unobstructed by solid particles.
The actual velocity of the water particles, v_a, follows from

$$v_a = Q/nA = v/n \qquad (23)$$

where
n = the porosity of the soil (dimensionless). Since n is always smaller than 1, it can readily be seen that the actual velocity of the water is always greater than the apparent velocity.

Darcy's law is valid for laminar flow, and since we are dealing with unconsolidated alluvial sediments through which groundwater is moving at a low speed, laminar flow conditions prevail; consequently, Darcy's equation may be applied without any appreciable error.

As noted above, Darcy's law has a certain analogy with other physical laws, e.g. Ohm's law in electricity: $i = V/r$, where i is the current (ampères), V is the Voltage (volts), and r is the resistance (ohms). If we compare both laws it can be seen that 1/K is comparable with r in Ohm's law and, in fact, the inverse of the hydraulic conductivity represents a resistance. In other words, the smaller the value of K, the larger the value of 1/K and the larger the resistance of the flow, and vice versa.
The analogy of Darcy's law with laws describing other physical phenomena has important consequences because it enables solutions of groundwater flow problems to be found from similar problems in other branches of physics. Further, flow of groundwater can be simulated by other kinds of flow, a fact used in the study of models, e.g. electrical models, or conductive sheet analogues, see Chap.7.

Numerical example

Seepage under a road bed is intercepted by a ditch, as shown in Fig.5. The hydraulic conductivity of the pervious layer is 0.4 m/day. According to Eq.22 for a unit cross-sectional area

$$v = 0.4 \frac{4.5 - 3.2}{25} = 0.02 \text{ m/day}$$

Assuming a cross-sectional area A = 3 m^2 per metre length of ditch and a ditch length of 400 m, the quantity of water seeping into the ditch is

$$Q = vA = 0.02 \times 3 \times 400 = 24 \text{ m}^3\text{/day}$$

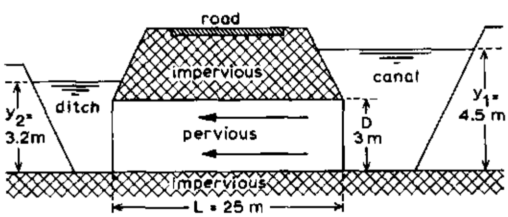

Fig.5.
Interception of seepage under a road bed.

6.3.2 THE PROPORTIONALITY CONSTANT K IN DARCY'S LAW

The proportionality constant K in Darcy's law, $v = - Ki$, represents the apparent flow velocity at unit hydraulic gradient. It is usually referred to as hydraulic conductivity or coefficient of permeability and depends on the properties of both fluid (groundwater) and porous medium (soil).
The flow of groundwater through the pores of the soil may be compared with the flow of a fluid through a narrow, circular tube of uniform radius R. For laminar flow of the fluid through the tube, the discharge can be expressed by the following formula, which is known as the Hagen-Poiseuille equation and was published in 1842

$$Q = \frac{\pi R^4 \rho g}{8\eta} \frac{\Delta h}{L} \tag{24}$$

where
Q = the rate of fluid flow (L^3T^{-1})
R = the radius of the tube (L)
η = the dynamic viscosity of the fluid $ML^{-1}T^{-1}$)

Δh = the head loss between the two points (L)

L = the length of the tube between the two points (L)

g = acceleration due to gravity (LT^{-2})

ρ = mass density of the fluid ML^{-3}).

Since the cross-sectional area of a circular flow tube, A equals πR^2 or $1/4\pi d^2$, where d is the diameter of the tube, Eq.24 may be written as

$$Q = \frac{d^2}{32} \frac{\rho g}{\eta} \frac{\Delta h}{L} A \tag{25}$$

Since, according to Eq.21, v = Q/A, Eq.25 can be written as

$$v = \frac{Q}{A} = \frac{d^2}{32} \frac{\rho g}{\eta} i \tag{26}$$

Laboratory and field experience shows that there is a close analogy between laminar flow in tubes and flow of water through soils. Comparing Eqs.22 and 26, it follows that

$$K = \frac{d^2}{32} \frac{\rho g}{\eta} \tag{27}$$

where d represents the mean diameter of the soil pores, being a parameter characteristic of the mean particle size.

Introducing a dimensionless constant depending on such physical properties as porosity, distribution and variation of the soil pores, particle shape, and packing, Eq.27 can be written as

$$K = Cd^2 \frac{\rho g}{\eta} = K' \frac{\rho g}{\eta} \tag{28}$$

where K' depends on the nature of the soil alone (and not on the properties of the fluid).

It was proposed by the Soil Science Society of America in 1952 that K' be termed intrinsic permeability, or simply permeability, whilst K (the proportionality constant in Darcy's law) be termed hydraulic conductivity. In practice, hydrologists will usually be dealing with the Darcy K, but both terms, permeability and hydraulic conductivity, are used interchangeably.

The methods used to determine the hydraulic conductivity and to evaluate its use in drainage studies are given in Chap.24, Vol.III.

6.3.3 INFLUENCE OF TEMPERATURE

As shown in Table 1 (Section 2.2) both the mass density and the viscosity of water are influenced by its temperature. In practice, the temperature dependency of the mass density is ignored, its value being taken as a constant, 1000 kg/m^3 However, as is obvious from this table, it is not always possible to ignore the influence of temperature on viscosity.
The hydraulic conductivity K at temperature x^{o}C can be obtained from K measured at y^{o}C by using the equation

$$K_{x^o} = K_{y^o} \frac{\eta_{y^o}}{\eta_{x^o}} \tag{29}$$

For example, if the hydraulic conductivity of a soil sample measured in the laboratory at 25^{o}is found to be 2 m/day, whilst the groundwater temperature is 10^{o}C, then

$$K_{10^o} = K_{25^o} \frac{\eta_{20^o}}{\eta_{10^o}} = 2 \times \frac{1.01 \times 10^{-3}}{1.31 \times 10^{-3}} = 1.5 \text{ m/day.}$$

6.3.4 FRESH-WATER HEAD OF SALTY GROUNDWATER

The potential, or hydraulic, head of groundwater was defined in Section 2.5, see also Fig.3. Hydraulic heads are measured in the field by means of a piezometer installed to the depth at which the water pressure is to be observed. Several techniques have been developed for the installation of piezometers and for the measurement of hydraulic heads and these will be discussed in Chap.20, Vol.III.

If measurements are taken in piezometers installed in a deep layer containing groundwater of different salt contents, i.e. the salt content varies laterally from very low to very high (from fresh water to saline water), the hydraulic heads measured in the salt water should, as a rule, be converted into fresh-water heads.
Expressing the fresh-water head, h_f, as (see Fig.6)

$$h_f = z + \frac{p}{\rho_f g}$$

and the salt-water head, h_s, as

$$h_s = z + \frac{p}{\rho_s g}$$

where ρ_f and ρ_s are the mass densities of fresh water and salt water, respectively, we obtain, after elemination of p/g

$$h_f = \frac{h_s \rho_s}{\rho_f} - z \frac{\rho_s - \rho_f}{\rho_f} \tag{30}$$

If the reference level coincides with the bottom end of the piezometer, or in other words if z = 0, the comparative fresh-water head can be expressed as

$$h_f = h_s \frac{\rho_s}{\rho_f} = \frac{p}{\rho_s g} \frac{\rho_s}{\rho_f} \tag{31}$$

For example, if the hydraulic head is measured in salt water as being 30 m above the reference level which is assumed to coincide with the bottom end of the piezometer, and the mass density of the groundwater is found to be 1025 kg/m^3, then the length of a column of fresh water of the same weight is

$$h_f = 30 \frac{1025}{1000} = 30.75 \text{ m}$$

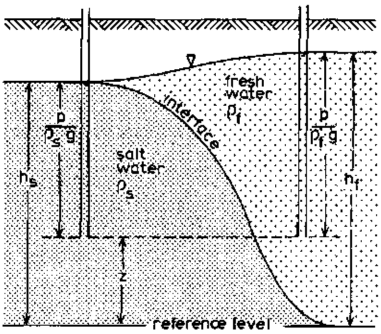

Fig.6. Hydraulic heads in fresh and salt water bodies.

6.4 SOME APPLICATIONS OF DARCY'S LAW

6.4.1 THE FALLING HEAD PERMEAMETER

Figure 7 shows the principle of a falling head permeameter. The head that causes the water to flow vertically downward through the sample decreases with time. According to Darcy

$$Q(t) = K \frac{h(t_1) - h(t_2)}{L} A$$

where Q(t) is the rate of flow through the sample as a function of time t, and $h(t_1)$ and $h(t_2)$ the hydraulic heads as function of time measured in respect to the constant outflow water level.

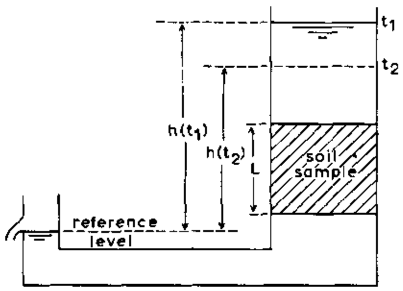

Fig.7. Falling head permeameter.

The rate of flow Q(t) is equal to the rate of lowering of the head dh/dt in the measuring tube, multiplied by the cross-sectional area. When the cross-sectional area of sample and measuring tube are equal, the following expression is obtained

$$-\frac{dh}{dt} = K\,\frac{h(t)}{L}$$

where the negative sign indicates that the head decreases with increasing time. Integration between the limits t_1 and t_2, during which h changes from $h(t_1)$ to $h(t_2)$ gives

$$\int_{h(t_1)}^{h(t_2)} -\frac{dh}{h} = \int_{t_1}^{t_2} \frac{K}{L}\,dt$$

or

$$-\ln h(t_2) + \ln h(t_1) = \frac{K}{L}(t_2 - t_1)$$

from which it follows that

$$K = \frac{L}{t_2 - t_1}\ln\frac{h(t_1)}{h(t_2)} \qquad (32)$$

For example, if the length of the sample L = 10 cm, $t_2 - t_1$ = 15 minutes, $h(t_1)$ = 45 cm and $h(t_2)$ = 35 cm, we find

$$K = \frac{10}{15}\ln\frac{45}{35} = \frac{10}{15}\,0.26 = 0.17 \text{ cm/min} = 24.5 \text{ m/day.}$$

6.4.2 FLOW THROUGH STRATIFIED SOILS

So far, soils have been considered homogeneous and isotropic. Isotropy with respect to hydraulic conductivity means that the hydraulic conductivity of the soil at a certain point has the same value for any direction of flow. Soils in situ, however, are rarely homogeneous and instead are made up of layers with different hydraulic conductivity.

Consider Fig.8 where water flows in a horizontal direction through three layers that have a different hydraulic conductivity K_1, K_2, and K_3, and a different thickness D_1, D_2, and D_3.

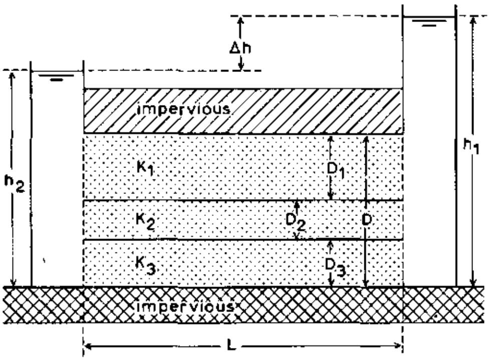

Fig.8. Horizontal flow through layered soil.

If we assume that there will be no flow across the boundaries between the individual layers, then the hydraulic gradient $i = (h_1 - h_2)/L = \Delta h/L$ applies to the flow through each layer.

The flow rate per unit width of each layer (q_1, q_2 and q_3) can be expressed by

$$q_1 = K_1 D_1 i$$
$$q_2 = K_2 D_2 i$$
$$q_3 = K_3 D_3 i$$

and the total flow is

$$q_1 + q_2 + q_3 = q = (K_1 D_1 + K_2 D_2 + K_3 D_3)i = \Sigma(KD)i \tag{33}$$

We may also write for the total flow through the three layers

$$q = \overline{K}(D_1 + D_2 + D_3)i = \overline{K}Di \qquad (34)$$

where $\overline{K}$ is the average hydraulic conductivity.
Equating the right hand sides of Eqs.33 and 34 yields the following expression for $\overline{K}$

$$\overline{K} = \frac{K_1D_1 + K_2D_2 + K_3D_3}{D_1 + D_2 + D_3} = \frac{\Sigma(KD)}{D} \qquad (35)$$

where $\Sigma(KD)$ is the transmissivity of the layered soil through which the water is moving horizontally.

Figure 9 shows a situation where water is flowing vertically downward through a soil profile made up of layers that have different thicknesses and different hydraulic conductivities.

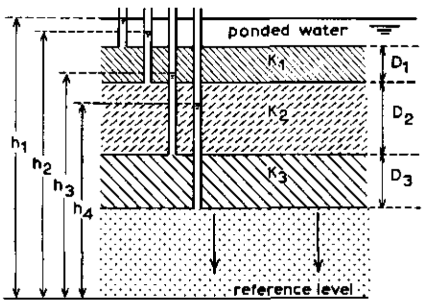

Fig.9. Vertical downward flow through layered soil.

The flow per unit cross-sectional area, i.e. the velocity of flow $v = Ki$, will be the same for each layer, assuming that the soil is saturated and no water can escape laterally. Hence

$$v = K_1\frac{h_1 - h_2}{D_1} \quad \text{or} \quad v\frac{D_1}{K_1} = h_1 - h_2$$

$$v = K_2\frac{h_2 - h_3}{D_2} \quad \text{or} \quad v\frac{D_2}{K_2} = h_2 - h_3$$

$$v = K_3\frac{h_3 - h_4}{D_3} \quad \text{or} \quad v\frac{D_3}{K_3} = h_3 - h_4$$

Since $(h_1 - h_2) + (h_2 - h_3) + (h_3 - h_4) = h_1 - h_4 = \Delta h$, adding these equations yields

$$v = \frac{\Delta h}{\frac{D_1}{K_1} + \frac{D_2}{K_2} + \frac{D_3}{K_3}} = \frac{\Delta h}{c_1 + c_2 + c_3} \tag{36}$$

where c_1, c_2, and c_3 are the respective hydraulic resistances of the three layers through which the flow passes vertically. The dimension of c is (T), for which usually days are used (D in m, K in m/day), see Chap.13, Vol.II. Its reciprocal value, $1/c = K/D$ is, in analogy with KD for horizontal flow, sometimes called transmissivity for vertical flow.

As an example, let us assume a situation as indicated in Fig.10, i.e. an upper clay layer in which the water table is assumed to remain stable (for instance by drainage or evaporation). The saturated thickness of this clay layer $D_1 = 9$ m; its hydraulic conductivity $K_1 = 1.0$ m/day.

Below this layer a clay bed is found; it is 1 m thick and its hydraulic conductivity $K_2 = 0.05$ m/day. This second clay bed lies on a sand layer that contains groundwater whose hydraulic head lies above the water table in the upper clay layer ($\Delta h = 0.05$ m). This head difference causes a vertical upward flow from the sand layer through the covering clay beds. According to Eq.36 the rate of this upward flow is

$$v = \frac{0.05}{9/1 + 1/0.05} = \frac{0.05}{9 + 20} = \frac{0.05}{29} = 0.0017 \text{ m/day}$$

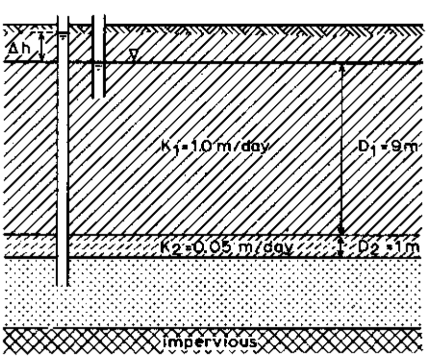

Fig.10.
Vertical upward flow through two clay beds with different hydraulic conductivities and different thicknesses.

6.5 BASIC EQUATIONS OF GROUNDWATER FLOW

In the previous sections we have briefly discussed the most elementary type of flow, namely linear flow, which we used to establish Darcy's law.

From a physical viewpoint, all fluid systems necessarily extend in three dimensions and their analysis then becomes much more complicated. In many groundwater flow problems, however, the flow is substantially the same in parallel planes and may then be treated as having a two-dimensional character. By this we mean that the velocity distribution vector in the fluid system varies only with two of the rectangular coördinates, and is independent of the third. For example, when land is drained by long and parallel open ditches or pipe drains, the flow pattern is the same in every vertical plane normal to the drains. Another case of two-dimensional flow is the flow to a pumped well that fully penetrates an aquifer. In this type of flow, the fluid motion is also independent of the vertical coördinate z. The often-used term "radial flow" means a two-dimensional flow symmetrical about an axis of symmetry.
For the solution of two- or three-dimensional flow problems, Darcy's law must be combined with the continuity equation discussed in Section 2.6. The resulting basic flow equation is a partial differential equation, called the Laplace equation.

6.5.1 THE LAPLACE EQUATION FOR STEADY FLOW

If we regard water as an incompressible fluid, the continuity equation for time-independent flow, as we have seen in Section 2.6, reads

$$\frac{\partial v_x}{\partial x} + \frac{\partial v_y}{\partial y} + \frac{\partial v_z}{\partial z} = 0$$

According to Darcy's law, and assuming a homogeneous and isotropic soil (hence $K_x = K_y = K_z = K$ where K is a constant), we may write

$$v_x = - K \frac{\partial h}{\partial x} \quad , \qquad v_y = - K \frac{\partial h}{\partial y} \quad \text{and} \qquad v_z = - K \frac{\partial h}{\partial z}$$

where v_x, v_y, and v_z are the velocity components in a rectangular coördinate system.
Substituting in the above continuity equation yields

$$\frac{\partial(-K\frac{\partial h}{\partial x})}{\partial x} + \frac{\partial(-K\frac{\partial h}{\partial y})}{\partial y} + \frac{\partial(-K\frac{\partial h}{\partial z})}{\partial z} = 0$$

or

$$\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} + \frac{\partial^2 h}{\partial z^2} = 0 \tag{37}$$

This is the Laplace equation for three-dimensional flow.
For two-dimensional flow it reduces to

$$\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} = 0 \tag{38}$$

Laplace's equation is also written as

$$\nabla^2 h = 0 \tag{39}$$

where the symbol ∇, called "del", is used to denote the differential operator

$$\frac{\partial}{\partial x} + \frac{\partial}{\partial y} + \frac{\partial}{\partial z}$$

and ∇^2, called "del squared", is used for

$$\frac{\partial^2}{\partial x^2} + \frac{\partial^2}{\partial y^2} + \frac{\partial^2}{\partial z^2}$$

which is called the Laplacean operator.

The above equations of flow and continuity are valid for a variety of saturated flows. Whenever a particular flow problem is investigated, its solution is uniquely determined only if it is known what happens at the boundaries of the flow region. Hence, for the solution of a particular flow problem, these so-called boundary conditions should be properly defined. They may include statements on the hydraulic head or the inflow and outflow conditions at the boundary, or that a boundary is a streamline etc. For further details on boundary conditions, see Section 7.

Finally it should be noted that, in drainage flow, complications arise from the fact that the flow region is usually bounded by the phreatic surface, whose shape is unknown. Simplifying assumptions have therefore been introduced, which lead to approximate solutions. The accuracy of such solutions is usually good enough for practical purposes.

For the solution of problems of nonsteady flow, the boundary conditions should be specified at all times, and the state of flow should be given at the time $t = 0$ at every point of the flow region. These specifications are called initial conditions.

6.5.2 THE DUPUIT-FORCHHEIMER ASSUMPTIONS

As was noted in the previous sections, in some studies of groundwater movement, including those of drainage flow, the water table is considered a free water surface. A free water surface is a surface in contact with and in equilibrium with the atmosphere. It is therefore a streamline along which the pressure is atmospheric.

Free-surface flow problems are difficult to solve because of the non-linear boundary conditions. An analysis of such problems based solely on the Darcy and Laplace equations leads to solutions of a rather complex form. A mathematically exact solution, however, is not always a desirable goal, when one considers the approximate nature of the differential equations themselves, the boundary conditions, and the assumptions of homogeneity, isotropy, and recharge from rainfall or irrigation water. This is the reason why scientists developed approximate methods of solution derived from hydraulics, which require less powerful mathematical tools.

By analogy with the flow in open channels, a hydraulic flow is defined as a mainly one-dimensional free-surface flow. It has the form of a stream-tube, whose transversal dimensions are much smaller than its length. The cross-section of the tube may vary only gradually with the distance along the main flow, so that transversal flow components may be neglected. Hence the streamlines are almost parallel to each other and the equipotential surfaces are almost planes perpendicular to the main flow, thus also almost parallel.

This method of solution was first developed by Dupuit in 1863 in the study of steady flow to wells and ditches (Fig.11). Dupuit assumed that

- for small inclinations of the free surface of a flow system, the streamlines can be taken as horizontal everywhere in a vertical section, and
- the velocity of flow is proportional to the slope of the free water table, but is independent of the flow depth.

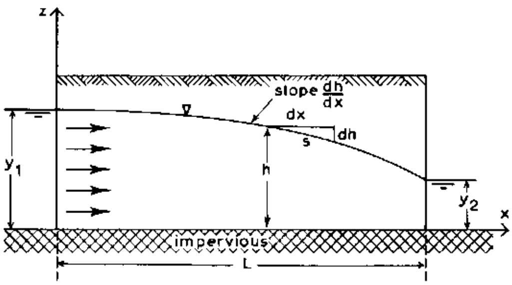

Fig.11. Steady flow in an unconfined aquifer, illustrating Dupuit's assumptions.

These assumptions imply a reduction of the dimensions of the flow (the two-dimensional flow becomes one of one dimension), and the flow velocity at the phreatic surface is proportional to the tangent of the hydraulic gradient instead of the sine (to dh/dx instead of dh/ds).

On the basis of these assumptions, Forchheimer (1886) developed a general equation for the free surface by applying the equation of continuity to the water in a vertical column in a flow region, bounded above by the phreatic surface and below by an impervious layer, the height of the fluid column being h, see Fig.12.

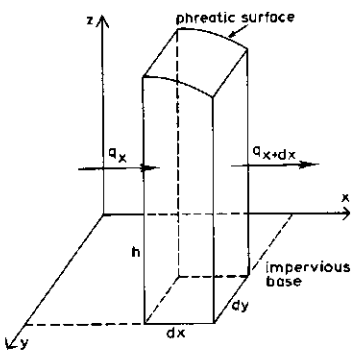

Fig.12. Approximate horizontal flow in space element of fluid as an assumption for deriving Forchheimer's linearized continuity equation.

Taking the surface of the impervious bed to be horizontal, i.e. coinciding with the plane through the horizontal coördinates x and y, the horizontal components of the flow velocity are given by

$$v_x = - K \frac{\partial h}{\partial x} \qquad \text{and} \qquad v_y = - K \frac{\partial h}{\partial y}$$

If q_x is the flow in the x-direction per unit width in the y-direction, then the flow entering through the left face of the column is the product of the area h dy and the velocity v_x

$$q_x\, dy = -K\left(h\frac{\partial h}{\partial x}\right)_x dy$$

Moving from the left to the right face of the column, the flow is changing at a rate $\partial q_x/\partial x$. When leaving the right-hand face of the column, q_x dy has changed to $q_{x+dx}dy$, i.e. to $(q_x + \frac{\partial q_x}{\partial x}dx)dy$.

The difference between outflow and inflow per unit time in the x-direction is

$$(q_{x+dx} - q_x)dy = \frac{\partial q_x}{\partial x}dx\, dy = -K\frac{\partial}{\partial x}\left(h\frac{\partial h}{\partial x}\right)dx\, dy$$

Similarly the change in flow in the y-direction is

$$\frac{\partial q_y}{\partial y}dx\, dy = -K\frac{\partial}{\partial y}\left(h\frac{\partial h}{\partial y}\right)dx\, dy$$

Assuming steady flow, the condition of continuity requires that the sum of the changes add up to zero. Hence

$$-K\left[\frac{\partial}{\partial x}\left(h\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(h\frac{\partial h}{\partial y}\right)\right] dx\, dy = 0 \qquad (40)$$

and

$$\frac{\partial}{\partial x}\left(h\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(h\frac{\partial h}{\partial y}\right) = 0 \qquad (41)$$

or

$$\frac{\partial^2 h^2}{\partial x^2} + \frac{\partial^2 h^2}{\partial y^2} = 0 \qquad (42)$$

which is Forchheimer's equation.

The Dupuit approximations have many advantages. To mention only a few:
- the mathematical difficulties encountered in solving many flow problems are reduced considerably,
- the three-dimensional water and impervious boundaries are replaced by fictitious vertical boundaries which may curve only in the horizontal plane,
- there are no boundaries with accretion from above or below, i.e. no water is added from above (rainfall, irrigation, recharge, etc.) or from below (inflow

of groundwater, artesian water),
- there is only one dependent variable, h, the elevation of the free water surface,
- the non-linear boundary conditions at the free surface vanish, so that the problem becomes linear,
- the principle of superposition can be applied, which enables the shape of the water table and the velocity at any point of the flow system to be determined.

See also BEAR et al. (1968).

6.5.3 NONSTEADY FLOW

When nonsteady flow conditions pertain, a state sometimes called transient flow, the sum of the changes of flow in the x- and y-direction must equal the change in the quantity of water stored in the column considered. This change in storage is reflected either in a drop or a rise of the phreatic surface when water is either released or taken up by the soil.
Quantitatively the change in storage is given by

$$\Delta S = \mu \Delta h \tag{43}$$

where
ΔS = change in water stored per unit surface area over the time considered,
μ = the effective porosity of the soil,
Δh = change in groundwater level over the time considered (L).

The effective porosity is thus defined as the fraction of the soil that releases or takes up water per unit change of water table height.
The principle of continuity requires that Eq.40 be now written as

$$- K \left[\frac{\partial}{\partial x} \left(h \frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y} \left(h \frac{\partial h}{\partial y}\right)\right] dx\, dy = - \mu \frac{\partial h}{\partial t} dx\, dy \tag{44}$$

or

$$\frac{\partial^2 h^2}{\partial x^2} + \frac{\partial^2 h^2}{\partial y^2} = \frac{2\mu}{K} \frac{\partial h}{\partial t} \tag{45}$$

Equation 44 may also be written as

$$K \left[h \frac{\partial^2 h}{\partial x^2} + \left(\frac{\partial h}{\partial x}\right)^2 + h \frac{\partial^2 h}{\partial y^2} + \left(\frac{\partial h}{\partial y}\right)^2 \right] = \mu \frac{\partial h}{\partial t} \tag{46}$$

If h is large compared with the changes in h, we may consider h a constant with the mean value D, and neglect the second order terms
$\left(\frac{\partial h}{\partial x}\right)^2$ and $\left(\frac{\partial h}{\partial y}\right)^2$, and we thus obtain

$$\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} = \frac{\mu}{KD} \frac{\partial h}{\partial t} \tag{47}$$

This equation is identical with that for two-dimensional heat conduction or that for two-dimensional flow of a compressible fluid through a porous medium.

6.5.4 REPLENISHMENT

It has been assumed so far that no accretion occurs, i.e. that the free surface receives no water from precipitation or irrigation. If there is accretion, for example by a constant rainfall at a rate R, the principle of continuity requires for steady flow (see Eq.44)

$$- K \left[\frac{\partial}{\partial x} \left(h \frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y} \left(h \frac{\partial h}{\partial y}\right) \right] dx\, dy = R\, dx\, dy \tag{48}$$

or

$$\frac{\partial^2 h^2}{\partial x^2} + \frac{\partial^2 h^2}{\partial y^2} = - \frac{2R}{K} \tag{49}$$

where R is the rate of recharge (LT^{-1}).
For nonsteady flow we have

$$- K \left[\frac{\partial}{\partial x} \left(h \frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y} \left(h \frac{\partial h}{\partial y}\right) \right] dx\, dy - R\, dx\, dy = - \mu \frac{\partial h}{\partial t} dx\, dy$$

or, by a procedure similar to that used in the derivation of Eq.47

$$KD \left(\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2}\right) = \mu \frac{\partial h}{\partial t} - R \tag{50}$$

It should be noted that the solutions to Eq.47 can be derived for quite arbitrary boundary and initial conditions for h. But the Dupuit assumptions underlying the original equation (44) are questionable and, in special cases, may lead to considerable errors in the results obtained. Therefore other methods, free from these assumptions, have been developed to solve two-dimensional groundwater flow problems. These methods are based on the technique of complex variables, see Section 6.3.

6.5.5 FLOW IN AN UNCONFINED LAYER BETWEEN TWO WATER BODIES

In this and the following sections some applications of the steady-state formula derived above will be presented.

Figure 11 shows the groundwater flow through a strip of land bounded by two water courses whose levels are at a height y_1 and y_2 above the impervious base. The groundwater table in the flow region is a free surface. If we take a rectangular coördinate system, the xz plane represents the plane of flow. It is assumed that the free surface is not recharged from rainfall or irrigation.

Forchheimer's equation (Eq.42) for one-dimensional flow reduces to

$$\frac{\partial^2 h^2}{\partial x^2} = 0$$

Integration gives

$$h^2 = Ax + B \tag{51}$$

where A and B are integration constants.

With the boundary conditions

$$\begin{aligned} &\text{for } x = 0 \qquad && h = y_1 \\ &\text{and } x = L \qquad && h = y_2 \end{aligned}$$

we obtain

$$A = \frac{y_2^2 - y_1^2}{L} \qquad \text{and} \qquad B = y_1^2$$

Substituting the values of A and B into Eq.51 yields

$$h^2 = \frac{y_2^2 - y_1^2}{L}\, x + y_1^2 \tag{52}$$

which shows that the groundwater table has a parabolic shape.

As to the discharge per unit width through a vertical cross-section, the solution is derived from the Darcy equation

$$q = - Kh \frac{dh}{dx}$$

which after integration and substitution of the boundary conditions yields

$$q = \frac{y_1^2 - y_2^2}{2L} K \tag{53}$$

As an example, let us assume that L = 25 m, y_1 = 6 m, y_2 = 5 m and K = 0.2 m/day. The discharge per unit width of water body is

$$q = 0.2 \frac{36 - 25}{2 \times 25} = 0.044 \text{ m}^3\text{/day}$$

If the water body is 400 m long, then the total discharge is

$$Q = 400 \times 0.044 = 17.6 \text{ m}^3\text{/day}$$

The head h at a distance x = 15 m is

$$h^2 = \frac{25 - 36}{25} 15 + 36 = 29.4 \text{ m}^2$$

$$h = 5.4 \text{ m}$$

It is noted that, due to the Dupuit assumptions, the thus computed position of the water table lies a little below the actual water table, in particular near the outflow.

6.5.6 STEADY FLOW TO PARALLEL DITCHES WITH A UNIFORM RECHARGE ON THE SOIL SURFACE

As another example of the application of the Dupuit assumptions, let us assume a homogeneous and isotropic soil, bounded below by an impervious layer and drained by a series of parallel ditches that penetrate the soil layer to the impervious base. The soil surface is recharged uniformly by rainfall at a rate R. The water levels in the ditches are at a height y_o, and the groundwater head is h. The ditches are a distance L apart, see Fig.13.

The problem is to find an expression for the height of the groundwater table midway between the ditches, H. Assume that the hydraulic gradient at any point is equal to the slope of the water table above that point (Dupuit-Forchheimer assumption). It can be seen from the figure that the assumption of horizontal flow is incorrect near the ditches, where the flow paths are curved. Where the slope of the groundwater table is relatively flat, the Dupuit-Forchheimer assumptions are nearly valid and only minor errors in the calculations will result.

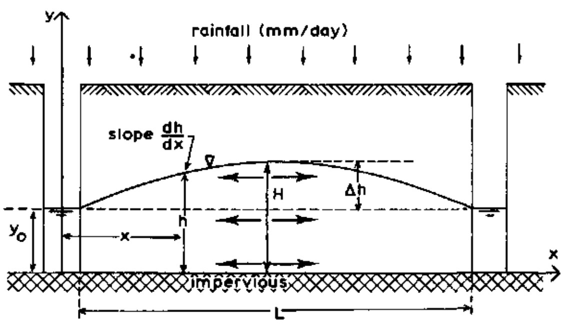

Fig.13. Flow to parallel ditches, which penetrate an unconfined aquifer to the impervious base. The water table is in equilibrium with the recharge from rainfall. The water level in the ditches is at equal height.

The solution of the problem can be found by taking a rectangular coördinate system whose origin lies on the impervious base below the centre of one of the ditches. From Fig.13 it can be seen that a vertical plane drawn midway between the ditches represents a division plane. All the water inflowing from the right of this plane flows into the right ditch and, similarly, all water entering to the left flows to the left ditch.

Let us consider the flow through a vertical plane at a distance x from the left ditch. All the water entering the soil to the right of this plane must pass through it on its way to the ditch. If R is the recharge per unit area of the soil surface per unit time, then the flow per unit time through the considered plane is

$$q_x = R\ (L/2 - x)$$

Obviously, under the present assumptions we may also apply Darcy's law to the flow through the plane, thus obtaining a second expression for q_x. For the hydraulic gradient we may write dh/dx, and at the plane the cross-sectional area of flow is equal to h. Hence

$$q_x = Kh \frac{dh}{dx}$$

Since the flow in the two cases must be equal, we may equate the right side of the two equations. Hence

$$Kh \frac{dh}{dx} = R(L/2 - x) \tag{54}$$

Multiplying both sides of this equation by dx gives

$$Kh\ dh = R(L/2 - x)dx$$

or

$$Kh\ dh = (RL/2)dx - Rx\ dx$$

which is an ordinary differential equation and can be integrated. The limits of integration are

$$\begin{array}{ll} \text{for } x = 0 & h = y_o \\ \quad\ x = 1/2L & h = H \end{array}$$

so that we may write

$$K \int_{h=y_o}^{h=H} h\ dh = R \int_{x=0}^{x=1/2L} (1/2L - x)dx$$

Substituting these limits yields

$$1/2K(H^2 - y_o^2) = R(1/2L)^2 - 1/2R(1/2L)^2 = 1/2R(1/2L)^2$$

$$L^2 = \frac{4K(H^2 - y_o^2)}{R} \tag{55}$$

which may be written as

$$L^2 = \frac{4K(H + y_o)\ (H - y_o)}{R}$$

The head difference $H - y_o = \Delta h$, or $H = y_o + \Delta h$ which, on substitution, yields

$$L^2 = \frac{4K(2y_o + \Delta h)\Delta h}{R}$$

or

$$L^2 = \frac{8Ky_o\Delta h}{R} + \frac{4K\Delta h^2}{R} \tag{56}$$

This formula is often used in solving drainage problems. It should be noted that in the drainage formulas elsewhere in this book the symbol h is used instead of Δh to denote the head difference.

6.5.7 STEADY FLOW TOWARDS A WELL

As a last example the flow towards a fully penetrating well will be analyzed (Fig.14). A homogeneous and isotropic aquifer is assumed, bounded below by a horizontal impervious layer and fully penetrated by a well. While being pumped, such a well receives water over the full thickness of the saturated aquifer since the length of the well screen is equal to the saturated thickness of the aquifer.

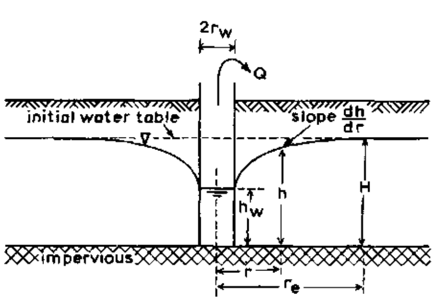

Fig.14. Horizontal radial flow towards a pumped well that fully penetrates an unconfined aquifer. No recharge from rainfall.

The initial groundwater table is horizontal, but attains a curved shape after pumping is started. Water is then flowing horizontally from all directions towards the well (radial flow).

It is further assumed that there is no recharge, and that the groundwater flow towards the well is in a steady state, i.e. the hydraulic heads along the perimeter of any circle concentric with the well are constant (radial symmetry).

The flow through any cylinder at a distance r from the centre of the well can be found by applying Darcy's law and assuming that the hydraulic gradient at this cylinder is equal to the slope of the groundwater table at the circle of this cylinder, dh/dr (Dupuit-Forchheimer assumption). Substituting this gradient, and the cross-sectional area of flow $A = 2\pi rh$ into the Darcy equation, yields

$$Q = 2\pi rhK \frac{dh}{dr} \tag{57}$$

where Q = the well discharge for steady radial flow to the well (L^3T^{-1}), and K = the hydraulic conductivity of the aquifer material (LT^{-1}).

On integration, we obtain

$$h^2 = \frac{Q}{\pi K} \ln r + C \tag{58}$$

Integration between the limits $h = h_w$ at $r = r_w$ and $h = H$ at $r = r_e$ yields

$$\int_{h=h_w}^{h=H} h^2 = \frac{Q}{\pi K} \ln \int_{r=r_w}^{r=r_e} r$$

or

$$H^2 - h_w^2 = \frac{Q}{\pi K} \ln \frac{r_e}{r_w}$$

and after rearranging

$$Q = \frac{\pi K(H^2 - h_w^2)}{\ln(r_e/r_w)} \tag{59}$$

which is Dupuit's formula.
A specific solution of this equation can be obtained by substituting a pair of values of h and r as observed in two observation wells at different distances from the centre of the well: for $r = r_1$, $h = h_1$ and for $r = r_2$, $h = h_2$.
The equation then reads

$$Q = \pi K \frac{h_2^2 - h_1^2}{\ln(r_2/r_1)} \tag{60}$$

If the drop in water table is small compared with the saturated thickness of the aquifer D then, by approximation, we may write $h_2 + h_1 = 2D$ and the equation becomes

$$Q = 2\pi KD \frac{h_2 - h_1}{\ln(r_2/r_1)} \tag{61}$$

It is obvious that this equation fails to describe accurately the drawdown curve near the well where the strong curvature of the water table contradicts the Dupuit-Forchheimer assumptions. For practical purposes and not too short distances from the well, the formula can be used without appreciable errors (see Chap.25, Vol.III).

6.6 SOME ASPECTS OF TWO-DIMENSIONAL FLOW

6.6.1 DISCONTINUOUS HYDRAULIC CONDUCTIVITY

In Section 4.2 it has been noted that soils are, in general, not homogeneous and isotropic, but are instead made up of different layers, having different hydraulic conductivities.

Let us assume a horizontal surface separating two regions of different hydraulic conductivities K_1 and K_2, see Fig.15. The directions normal and tangential to the common boundary of the two flow regions are denoted by n and t respectively. The components of the flow velocity in these directions are denoted by v_{n_1} and v_{t_1} in the upper region with hydraulic conductivity K_1, and by v_{n_2} and v_{t_2} in the lower region with hydraulic conductivity K_2.

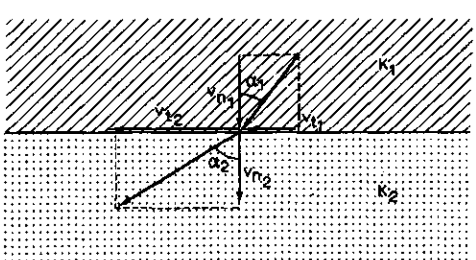

Fig.15. Refraction at the boundary of two homogeneous isotropic layers with hydraulic conductivities K_1 and K_2 respectively ($K_1 < K_2$).

Along the boundary the pressure, and hence the hydraulic head h, must necessarily have the same value on either side. The principle of continuity requires that all water leaving the upper region must enter the lower region. This condition implies that the components of the flow velocity vector normal to the common boundary must be the same on either side

$$v_{n_1} = v_{n_2} \tag{62}$$

Applying Darcy's law to the components along the boundary, v_{t_1} and v_{t_2}, we obtain

$$v_{t_1} = - K_1 \frac{\partial h}{\partial t} \qquad \text{and } v_{t_2} = - K_2 \frac{\partial h}{\partial t} \quad .$$

Since the hydraulic head h is the same on both sides of the boundary, and t is the coördinate along it, $\partial h/\partial t$ has the same value on both sides. Hence

$$v_{t_1}/K_1 = v_{t_2}/K_2 \tag{63}$$

If α_1 and α_2 represent the angles of the respective velocity vectors with the direction normal to the boundary, then it follows from Eqs.62 and 63 that

$$\frac{K_1}{K_2} = \frac{\tan \alpha_1}{\tan \alpha_2} \tag{64}$$

From this result it follows that there is an abrupt change in direction of the flow velocity vector along the boundary of two regions with different hydraulic conductivity. It also shows that if $K_2 >> K_1$, then $\tan \alpha_2$ is very large compared with $\tan \alpha_1$. Such a situation is found where a clay layer (K_1) covers a sand layer (K_2). In the clay layer the flow is almost vertical and yet in the sand layer it is nearly horizontal. This provides a justification for the assumptions generally made that in a semi-confined aquifer the groundwater flow in the sand can be regarded as horizontal, and in the covering clay layer as vertical.

6.6.2 POTENTIAL AND STREAM FUNCTIONS

Darcy's law for steady two-dimensional flow in a homogeneous soil states that

$$v_x = - K \frac{\partial h}{\partial x} \qquad \text{and} \qquad v_y = - K \frac{\partial h}{\partial y} \tag{65}$$

where v_x and v_y are the components of the flow velocity vector in the coördinate directions of x and y.
In this equation the hydraulic head h is a scalar quantity, i.e. at any point it can be expressed and defined solely by a number, in contrast to a vectorial quantity, which requires in addition a direction.
We have assumed that the soil is homogeneous - hence K is a constant - and that the water has a constant density and viscosity. Introducing a vector derivable from the scalar quantity Φ, namely gradient Φ or grad Φ, we can now express Darcy's law (Eq.20) in the following vectorial form

$$V = - \text{grad } \Phi \tag{66}$$

where

$$\Phi = Kh = K(z + p/\rho g) \tag{67}$$

In this equation Φ is called the velocity potential and has the dimension (L^2T^{-1}). Its derivatives with respect to the coördinates x and y constitute the components in the x- and y-direction of a vector, the flow velocity vector. As will be shown below, the velocity potential may be combined with a stream function Ψ.
As a direct consequence of its definition, the velocity potential Φ is a single-valued function in every point of the x, y-plane. Hence it is possible to draw lines of constant Φ in this plane (Fig.16). Such lines are called equipotential lines and are usually drawn at equal intervals, so that

$$\Phi_1 - \Phi_2 = \Phi_2 - \Phi_3 = ... = \Delta\Phi \tag{68}$$

In the case of nonsteady flow, lines drawn perpendicular to the equipotential lines are called streamlines and are indicated by the symbol Ψ. Streamlines give the instantaneous flow pattern and should be distinguished from so-called path lines, which show the lines along which fluid particles move in steady flow.
It can be demonstrated that a streamline, which is a directed line and in general curved, is at any point tangent to the velocity vector at that point or to - grad Φ. The components of this velocity vector, v_x and v_y, may be expressed as

$$v_x = - \frac{\partial\Phi}{\partial x} \quad \text{and} \quad v_y = - \frac{\partial\Phi}{\partial y} \tag{69}$$

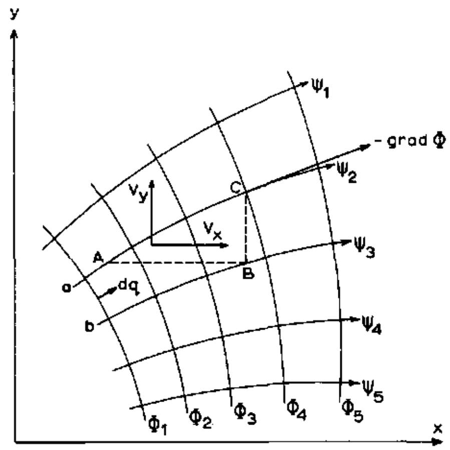

Fig.16.
Orthogonal network of equipotential lines and streamlines.

The tangent along any line Φ = constant has a slope $(\frac{dy}{dx})_{\Phi=C}$. The value of this slope may be found by taking the total differential of Φ = constant

$$d\Phi = \frac{\partial\Phi}{\partial x}\,dx + \frac{\partial\Phi}{\partial y}\,dy = 0$$

or

$$(\frac{dy}{dx})_{\Phi} = -\frac{\partial\Phi/\partial x}{\partial\Phi/\partial y} = -\frac{v_x}{v_y} \tag{70}$$

Let us now assume a flow rate dq in the stream tube bounded by two adjacent streamlines a and b of Fig.16. The principle of continuity requires that this volume per unit of time passes through the sections A-B and B-C, which are chosen in such a way that only the first has contributions due to v_y and the second only to v_x. As a first approximation, neglecting the contributions $(\partial v_x/\partial x)dx\,dy$ and $(\partial v_y/\partial y)dy\,dx$ as second-order terms, continuity requires that

$$dq = v_y dx = v_x dy$$

or

$$v_y dx - v_x dy = 0 \tag{71}$$

For two-dimensional flow the equation of continuity (Eq.16) reduces to

$$\frac{\partial v_x}{\partial x} + \frac{\partial v_y}{\partial y} = 0 \qquad (72)$$

from which it follows that Eq.71 is derived from a total differential.

A second function of great importance in the theory of groundwater flow is the stream function Ψ. The flow velocity vector must satisfy the equation of continuity (Eq.72). Hence the vector components v_x and v_y can be derived from a function Ψ by

$$v_x = -\frac{\partial \Psi}{\partial y} \qquad \text{and} \qquad v_y = +\frac{\partial \Psi}{\partial x} \qquad (73)$$

When these expressions are substituted into Eq.71 we obtain

$$\frac{\partial \Psi}{\partial x}\,dx + \frac{\partial \Psi}{\partial y}\,dy = d\Psi = 0$$

Hence

$$\Psi(x,y) = \text{constant} \qquad (74)$$

$\Psi(x,y)$ is called the stream function, and the lines of constant Ψ streamlines. Thus the streamlines may be considered in the same way as the equipotential lines. Equation 71 may now be written as

$$\left(\frac{dy}{dx}\right)_{\Psi} = \frac{v_y}{v_x} \qquad (75)$$

Comparing this equation with Eq.70, the orthogonality of streamlines and equipotential lines is proved, since from mathematics we know that two lines are orthogonal if the product of their slopes equals -1,

$$\frac{v_y}{v_x} \times -\frac{v_x}{v_y} = -1$$

Analogous to the hydraulic head h, the velocity potential Φ satisfies the Laplace equation (Eq.39). Hence

$$\nabla^2\Phi = \frac{\partial^2\Phi}{\partial x^2} + \frac{\partial^2\Phi}{\partial y^2} = 0 \tag{76}$$

A comparison of Eqs.76 and 73 and elimination of v_x and v_y leads to the relationships

$$\frac{\partial\Phi}{\partial x} = \frac{\partial\Psi}{\partial y} \quad \text{and} \quad \frac{\partial\Phi}{\partial y} = -\frac{\partial\Psi}{\partial x} \tag{77}$$

which are the Cauchy-Rieman (or d'Alembert-Euler) conditions.

Like equipotential lines, streamlines are drawn at equal intervals. By choosing $\Delta\Phi$ and $\Delta\Psi$ equal, and small enough, the system of streamlines and potential lines form elementary, curvilinear squares. This net of squares is called a flow net, Fig.16. For specified boundary conditions, the technique of sketching flow nets is often used as an approximate method to solve two-dimensional groundwater flow problems.

It should be noted that the velocity potential Φ and stream function Ψ are each sufficient to describe a groundwater flow problem completely. For both functions the basic differential equation is the Laplace equation. The components of the flow velocity can easily be obtained either from Φ or from Ψ, by differentiation with respect to one of the coördinates. Since Φ is directly related to the head h, which is measured in the field, a formulation of the flow problem in terms of Φ is to be preferred.

6.6.3 EXACT METHODS OF SOLUTION

As noted earlier, problems of free-surface flow are difficult to solve exactly because of the non-linear boundary conditions. Approximate methods of solution, based on the Dupuit-Forchheimer assumptions, have therefore been developed. In practice the results obtained with these methods are sufficiently accurate to be applied to drainage flow problems.

An exact solution of two-dimensional groundwater flow problems can be obtained by applying the so-called complex variable method and by formulating the problem in terms of velocity potential Φ and stream function Ψ as outlined above.

A complex variable is a quantity of the form $z = x + iy$, where x and y are real numbers and i denotes the imaginary unit $\sqrt{-1}$. If the Cauchy-Rieman conditions are satisfied, then the linear combination of the functions Φ and Ψ

$$\omega = \Phi + i\Psi \qquad (78)$$

is a function of the complex variable $z = x + iy$.
Any analytic function $\omega = f(z)$ corresponds to two real functions: $\Phi = \Phi(x,y)$ and $\Psi = \Psi(x,y)$, which both satisfy Laplace's equation and therefore may be considered potential functions corresponding to a groundwater flow in homogeneous and isotropic soil.
A flow problem can now be solved by finding a solution either for the potential, or for the stream function. Since both functions satisfy the same type of differential equation, the methods to be applied will be basically the same. A problem has been solved as soon as one of the functions is known. If this function is the potential for a two-dimensional flow problem, we have an expression for $\Phi(x,y)$. According to the first part of Eq.77, this expression has to be differentiated with respect to x to find $\partial\Psi/\partial y$. Integration of the latter gives an expression for $\Psi(x,y)$.
It is obvious that the same result will be obtained by differentiating $\Phi(x,y)$ with respect to y and integrating the result with respect to x. Thus the Cauchy-Rieman equations are used to find $\Psi(\Phi)$ when $\Phi(\Psi)$ is given, and inversely.
A pair of functions like Φ and Ψ, which both satisfy Laplace's equation, are called conjugate functions. These functions are such that indeed the curves $\Phi(x,y)$ = constant and $\Psi(x,y)$ = constant form a system of orthogonal trajectories.
It is beyond the scope of this chapter to describe the details of this complex variable technique. It should, however, be regarded as one of the most powerful tools in solving two-dimensional groundwater flow problems in an exact manner. The method facilitates the solution of flow problems for regions bounded by fixed potential lines and streamlines.

6.6.4 SOME OTHER APPROXIMATE METHODS OF SOLUTION

Finally two other approximate methods of solution should be mentioned.
. The relaxation method.
This method is a numerical method for the solution of Laplace's equation in two dimensions, and is based on the replacement of the differential quotients by finite difference expressions. Calculations are made by hand, though an electronic computer can be used. The flow region is divided into a grid or square net, as shown in Fig.17. Starting with the known values of the hydraulic head along the boundaries of the flow region, numerical values are arbitrarily assigned to the head at each grid point. According to the expression

$$h_o = \frac{1}{4}(h_1 + h_2 + h_3 + h_4) \quad (79)$$

the head at the grid point with subscript 0 must be the mean value of the values in the four surrounding grid points with subscripts 1 to 4. Equation 79 should be satisfied for all grid points. This can be done by adjusting the values empirically, retaining of course the given values along the boundaries.
The condition expressed by Eq.79 for each grid point is satisfied in consecutive steps. After assumed values of h have been assigned to each grid point, the error at each point is determined, using

$$\text{error} = 4h_o - (h_1 + h_2 + h_3 + h_4) \quad (80)$$

Then the point is taken where the absolute value of the error is largest and the value of h at that point is reduced by one fourth of the error value. Next the new value of the error at the surrounding points should be calculated. This procedure is repeated until the remaining errors are sufficiently small (Fig.18).

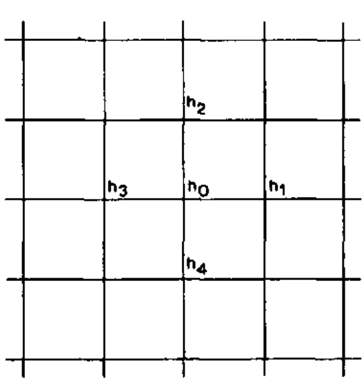

Fig.17. Mesh points for relaxation method.

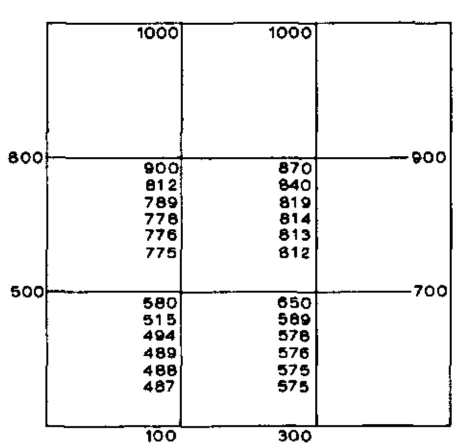

Fig.18. Example of relaxation method.

It will be clear that the smaller the size of the grid the better and more reliable are the results. The percentage error of the relaxation process as compared with an exact analysis of the same problem can be made as small as desired. It will be understood that, when working by hand, the calculations are laborious.

. <u>The method of squares</u>.

Instead of computing the values of Φ or Ψ at certain points of the flow region, or trying to find a potential or stream function, the equipotential lines and streamlines can be drawn by trial and error, making use of the property that they are perpendicular to each other. As noted earlier, these lines form elementary squares when the intervals between consecutive potential lines and streamlines are chosen equal in magnitude ($\Delta\Phi = \Delta\Psi$).

Starting from the boundary conditions, a first approximate system of potential lines or streamlines is drawn, whatever suits best in view of the data available. If, for example, the potential lines are drawn first, then the streamlines are drawn perpendicular to the potential lines. The thus obtained flow net is then adjusted in consecutive stages until streamlines and potential lines are everywhere orthogonal and form elementary squares, and the flow net complies with the boundary conditions. After the flow net has been adjusted, the total discharge (per unit thickness) flowing through the system under consideration can be calculated from

$$Q = |\Phi_1 - \Phi_2| \frac{m}{n} \tag{81}$$

where $|\Phi_1 - \Phi_2|/n$ is the potential drop over each square, n is the number of squares in a stream lane, and m is the number of stream lanes.

6.7 BOUNDARY CONDITIONS

From theory it is known that such partial differential equations as Laplace's equation have infinite numbers of solutions. The question arises as to how one may choose, among these infinities of solutions, which to apply to any particular problem. As already mentioned in Section 5.1, whenever a particular flow problem is investigated, its solution is uniquely determined only if it is known in detail what happens at the boundaries of the flow region. Boundary conditions in groundwater flow problems describe the detailed physical conditions that are to be imposed at the boundaries of the flow region. These boundaries are not necessarily impervious layers or walls confining the groundwater to a given region. Rather, they are geometrical surfaces at all points of which either the flow velocity of the groundwater, or the velocity potential, or a given function of both, may be considered as known. Some characteristic boundary conditions will now briefly be discussed.

6.7.1 IMPERVIOUS BOUNDARIES

Impervious layers are to be regarded as representing streamlines, because there is no flow across them. The flow velocity component normal to such boundaries vanishes. Hence we have Ψ = constant and $d\Psi/ds = 0$.
In practice a layer is considered impervious if its hydraulic conductivity is very small compared with the hydraulic conductivity of adjacent layers.

6.7.2 PLANES OF SYMMETRY

Planes of symmetry are shown in section in Fig.19 by lines such as A-B (vertically through the drain axis) and C-D (parallel to A-B, but midway between the drains). Because of the symmetry of the system, the pattern of equipotentials and streamlines on one side of such a "boundary" is the mirror image of that on the other side. Hence any flow velocity component immediately adjacent to the boundary which is perpendicular to that boundary must be matched by a component in the opposite direction immediately on the opposite side of the boundary. The net flow across the boundary must therefore be zero and the plane of symmetry is, like an impervious layer, a streamline of the system.

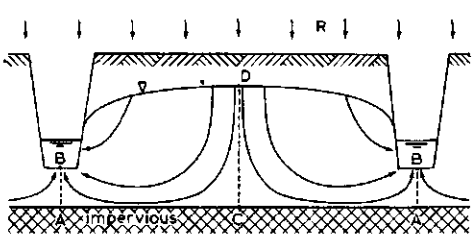

Fig.19. Boundary conditions for steady flow to drainage ditches.

6.7.3 FREE WATER SURFACE

The free water surface is defined as the surface where the pressure equals the atmospheric pressure. It is assumed that the free water surface limits the flow region, i.e. no flow occurs above this surface. This is untrue for most instances of flow through soils, but the assumption is useful in analyzing the flow through media having very small capillary fringes or wherever the flow region is very large compared with the capillary fringe.
For a free water surface the pressure component of the head, $p/\rho g$, is zero; hence

the total head is equal to the elevation component: $h = z$.
If there is no percolation of water towards the free water surface, the flow velocity component normal to that surface is zero and the free water surface then represents a streamline.
In the case of percolation, however, the intensity of vertical recharge R determines the value of the streamlines. In Fig.20 the rainfall intensity is R and there is a flow towards the ditch. The streamlines have a value Rx, where x is the distance from the ditch. The free water surface is neither an equipotential line nor a streamline, and the streamlines have their starting point at regular distances from each other.

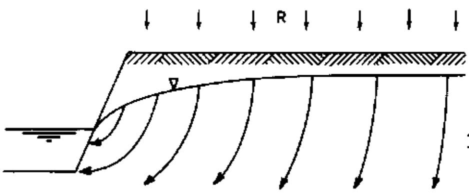

Fig.20. Boundary conditions free water surface.

6.7.4 BOUNDARIES WITH WATER AT REST OR SLOWLY MOVING WATER

Such boundaries are found along the walls of ditches and reservoirs and where, for instance, upward groundwater flow meets downward percolating water.
The hydrostatic pressure along the wall of the ditch, i.e. the pressure acting on the wall due to a certain height of water standing above it, is given by

$$p = \rho g(z_o - z) \qquad (82)$$

where z = the height of the point considered (Fig.21).

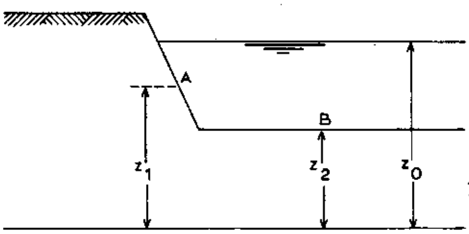

Fig.21. Boundary conditions water at rest or slowly moving.

It then follows that

$$z_o = \frac{p}{\rho g} + z \tag{83}$$

Now the right-hand member represents the potential or hydraulic head and thus the potential at each point along the ditch is equal to the height z_o of the water level in the ditch. In Fig.21 we have

Point	Elevation (z)	Pressure ($p/\rho g$)	Sum (z_o)
A	z_1	$z_o - z_1$	z_o
B	z_2	$z_o - z_2$	z_o

6.7.5 SEEPAGE SURFACE

At points in the soil above the groundwater table the pressure is negative, while at points below the groundwater table is it generally positive. An exception occurs if the groundwater table intersects the surface of the soil, as shown in Fig.22. In this case a surface of seepage occurs, defined as the boundary of the soil mass where water leaves the soil, and then continues its flow in a thin film along the outer boundary of the soil. Surfaces of seepage also occur on the downstream face of dams through which water is seeping.

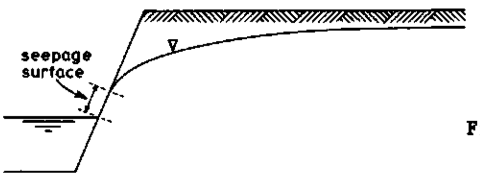

Fig.22. Boundary conditions for surface of seepage.

Along a seepage surface the pressure head p = 0 (atmospheric pressure). Hence the hydraulic head at any point on the seepage surface is equal to the elevation head at that point, or h = z.
A seepage surface is not a streamline for the groundwater movement because in the interior of the soil mass there may be a non-zero component of the flow velocity vector perpendicular to the boundary.

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INTRODUCTORY SUBJECTS

7. ELECTRICAL MODELS: CONDUCTIVE SHEET ANALOGUES

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PURPOSE AND SCOPE

The analogy between the flow of groundwater and that of electricity enables solutions for groundwater flow problems to be obtained from electrical models. The analogy is discussed and an example is given of the application of the conductive sheet type of electrical model to groundwater flow problems with simple boundary conditions.

CONTENTS

		page
7.1	Analogues	203
7.2	Analogy between the flow of groundwater and that of electricity	204
7.2.1	Flow of groundwater	204
7.2.2	Flow of electricity	206
7.2.3	The analogy	208
7.3	Two-dimensional models using Teledeltos paper	208
7.4	Streamlines, equipotential lines; boundary conditions	211
7.5	Simulation of boundary conditions in electrical models	213
7.6	Measuring in the model	217
7.6.1	Measurement of discharge	217
7.6.2	Measurement of equipotential lines	217
7.6.3	Measurement of streamlines, inverted model	217
7.7	Example: Determining the free water surface in a dam	218
Literature		221

7.1 ANALOGUES

The flow of groundwater is governed by Darcy's law and by the law of conservation of mass. Together, these two laws lead to differential equations like the Laplace equation for steady flow (Chap.6, Vol.I) and the "heat flow" equation for transient flow.

For steady-state conditions (to which our considerations will be limited), the basic Laplace equation reads

$$\nabla^2 h = 0 \tag{1}$$

with

$$\nabla^2 h = \frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} + \frac{\partial^2 h}{\partial z^2} \tag{2}$$

where h = hydraulic head (m), ∇^2 is the Laplace operator, and x, y, z are cartesian coördinates (m). The solution of such equations depends on the boundary conditions of the problem. Analytical solutions are available for simple boundary conditions, but for more complicated conditions solutions are either unknown or they become very complex.

Solutions may then be found through

a. approximations
b. numerical methods
c. analogue model studies.

a) Approximate solutions are based on simplifying assumptions (e.g. the Dupuit-Forchheimer assumptions). Their accuracy is often good enough for practical purposes.

b) Numerical solutions can be highly accurate but as the calculations are lengthy, they often require the use of an electronic computer.

c) Analogue solutions show varying accuracy. The most simple ones are about 10% in error, but with slightly more sophistication, the error level may be reduced to 1%. Even uncertainties of 10% are often acceptable in practice because the parameters employed (transmissivity KD, hydraulic resistance c, effective porosity μ, etc.) usually have errors of this order of magnitude, while an error level of 1% is more than accurate enough for most engineering purposes.

Two systems are said to be analogous if there is a one-to-one correspondence between each element in the two systems, as well as between the excitation and response functions of these elements and the systems as a whole (KARPLUS, 1958). Analogues are very instructive and they have the advantage that the engineer keeps in close touch with his problem. During the investigation he remains able to introduce variations and improvements in his design as soon as they seem appropriate.
Analogues are based on the similarities between laws which apply to different physical systems. The following group of processes obey similar laws
- flow of fluids through porous media,
- laminar flow of fluids (e.g. between parallel plates),
- flow of heat through a conductor,
- deflection of a stretched membrane by the action of a load, and
- flow of electricity through a conductor.

In each of these fields problems can be studied by constructing a model. This model may belong to the same field (e.g. on a smaller scale), but it may also belong to one of the other groups. Thus, a groundwater flow problem may be studied on a reduced scale (e.g. in a sand tank), but it may also be transformed into a heat flow or an electrical analogue. Electricity has several advantages: it is clean, easy to handle, and it can be conveniently and accurately measured.

7.2 ANALOGY BETWEEN THE FLOW OF GROUNDWATER AND THAT OF ELECTRICITY

7.2.1 FLOW OF GROUNDWATER

The one-dimensional steady flow of groundwater through a porous medium with rectangular cross-section is described by Darcy's law

$$Q = K \frac{\Delta h}{L} BD \tag{3}$$

where
Q = discharge ($m^3\ day^{-1}$)
K = hydraulic conductivity ($m\ day^{-1}$)
Δh = difference in hydraulic head (m)
B, D = width and thickness of the cross-section (m)
L = distance along the average direction of flow (m)

The quantity Q/BD, the discharge per unit cross-sectional area, is called specific discharge and denoted by v. If we write $\Delta h/L = -\nabla h$, Eq.3 takes the form

$$v = -K\nabla h \tag{4}$$

The minus sign in this equation merely indicates that the direction of flow is opposite to the direction in which h increases. For two or three dimensional flow in isotropic media, Eq.4 remains valid, if we define the operator ∇ (nabla or del) as taking derivatives with respect to the coördinates x, y, and z, and adding the results

$$\nabla = \frac{\partial}{\partial x} + \frac{\partial}{\partial y} + \frac{\partial}{\partial z}$$

To solve problems of groundwater flow Darcy's law alone is not enough. At a point in a three-dimensional space the specific discharge, which is a vectorial quantity, can be described by its three components, v_x, v_y, and v_z (Fig.1). Darcy's law gives only three relations between four unknown quantities: the three vector components and the head. A fourth relation may be obtained by recalling that the flow has to satisfy the fundamental physical principle of conservation of mass, which states that in a closed system no groundwater can be generated or destroyed.

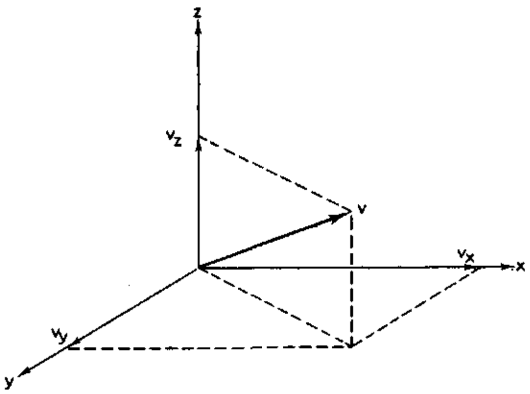

Fig.1. Specific discharge vector v, and its three components v_x, v_y and v_z.

Assuming that the density of water ρ is constant, this principle can be expressed mathematically by the equation of continuity

$$\frac{\partial v_x}{\partial x} + \frac{\partial v_y}{\partial y} + \frac{\partial v_z}{\partial x} = 0 \tag{5}$$

The combination of Eqs.4 and 5 leads to the Laplace equation

$$\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} + \frac{\partial^2 h}{\partial z^2} = 0 \tag{6}$$

which is often written in its abbreviated form as

$$\nabla^2 h = 0 \tag{7}$$

The operator ∇^2 (nabla squared, or del squared) is called Laplace's operator. It denotes the operation of taking the second derivatives with respect to the coördinates x, y, and z, and adding the results

$$\nabla^2 = \frac{\partial^2}{\partial x^2} + \frac{\partial^2}{\partial y^2} + \frac{\partial^2}{\partial z^2}$$

If the flow problem to be solved is a two-dimensional one, the term in z of Eq.6 vanishes.

7.2.2 FLOW OF ELECTRICITY

The one-dimensional flow of electricity through a conducting rod (Fig.2) is described by Ohm's law

$$\Delta U = RI \tag{8}$$

where
ΔU = potential difference (volts)
R = resistance of the rod (ohms)
I = current (ampères)

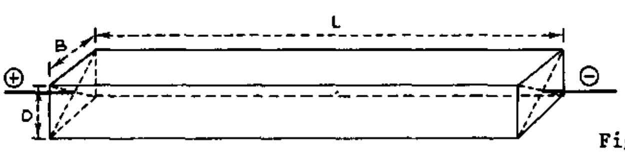

Fig.2. Resistance of a rod.

A rod with a rectangular cross-section has a resistance

$$R = \frac{1}{\sigma}\frac{L}{BD} \tag{9}$$

where

σ = specific conductivity ($\text{ohms}^{-1}\text{m}^{-1}$)
L, B, D = length, width and thickness of rod (m)

Combining Eqs.8 and 9 leads to the equation for the (one-dimensional) longitudinal flow

$$I = \sigma \frac{\Delta U}{L} BD \tag{10}$$

or

$$J = -\sigma \nabla U \tag{11}$$

where
J = I/BD is *current density* (amp m^{-2})
∇U = $-(\Delta U/L)$ is potential gradient (volts m^{-1})

The minus sign in Eq.11 again indicates that the direction of flow is opposite to the direction in which U increases.
For two- or three-dimensional flow Eq.11 remains valid. In the case of three-dimensional flow the current density J is a vectorial quantity which, analogous to the specific discharge in groundwater flow, can be described by its three components J_x, J_y, and J_z.

Continuity requires that

$$\frac{\partial J_x}{\partial x} + \frac{\partial J_y}{\partial y} + \frac{\partial J_z}{\partial z} = 0 \tag{12}$$

The combination of Eqs.11 and 12 leads to the Laplace equation

$$\frac{\partial^2 U}{\partial x^2} + \frac{\partial^2 U}{\partial y^2} + \frac{\partial^2 U}{\partial z^2} = 0 \tag{13}$$

which in its abbreviated form is written as

$$\nabla^2 U = 0 \tag{14}$$

where ∇^2 is the Laplace operator, see Section 7.2.1.

In the case of two-dimensional flow Eq.13 reduces to

$$\frac{\partial^2 U}{\partial x^2} + \frac{\partial^2 U}{\partial y^2} = 0 \qquad (15)$$

7.2.3 THE ANALOGY

Table 1 presents the analogous elements of groundwater flow and the flow of electricity.

Table 1. Corresponding elements of groundwater flow and that of electricity.

Groundwater		Electricity	
Hydraulic head difference	Δh(m)	Potential difference	ΔU(volts)
Hydraulic conductivity	K(m day^{-1})	Specific conductivity	σ($ohms^{-1}m^{-1}$)
Flow rate	Q(m^3 day^{-1})	Current	I(ampèrés)
Specific discharge	v(m day^{-1})	Current density	J(amp m^{-2})
Darcy's law	$v = - K\nabla h$	Ohms law	$J = - \sigma\nabla U$
Laplace equation	$\nabla^2 h = 0$	Laplace equation	$\nabla^2 U = 0$

7.3 TWO-DIMENSIONAL MODELS USING TELEDELTOS PAPER

Analogues may be discrete or continuous with respect to space variables. If an area under investigation is defined only at specific points of the analogue, the analogue is said to be discrete. If every point of the area is represented in the analogue, the analogue is continuous. Time, of course, is a continuous variable in both types.
The usual types of electrical models employed in studying problems of steady groundwater flow are listed in Table 2.

We shall confine our discussion to conductive sheet analogues.
Sheet conductors are very useful in the study of two-dimensional groundwater flow problems. The most commonly used conductive sheet is the Teledeltos paper, which is commercially available as electro-sensitive recording paper; it is easy to handle and relatively cheap. This paper is formed by adding carbon black, a con-

ductive material, to paper in the pulp-beating stage of the paper-manufacturing process.

Table 2. Types of electrical models

Type	Number of dimensions	Continuity	Error
Sheet analogues	2	yes	10%
Electrolytic tank	2 or 3	yes	2 - 5%
Resistance network	2 or 3	no	1%

The conductive paper is then coated on one side with a lacquer which acts as an electrical insulator and on the other side with a layer of aluminium paint (Fig.3). As a result of the manufacturing process the paper is not quite homogeneous, while the conductivity varies slightly in different directions (anisotropy). The resistance of any sheet is likely to show a 10 per cent variation and the resistance measured in the length direction of a roll is about 10 per cent lower than that across the roll. Nevertheless the paper is well suited for obtaining a first approximation in theoretical studies and for solving practical problems.

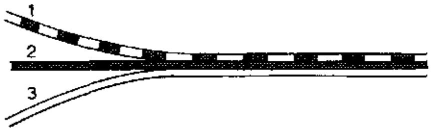

Fig.3. Teledeltos paper
1 = semi-conductive aluminium sheet
2 = conductive graphite paper
3 = lacquer coating.

In a sheet conductor, D is constant, so we may write Eq.9 as

$$R = \frac{1}{\sigma D}\frac{L}{B} = R_s\frac{L}{B} \tag{16}$$

If $L = B$, $R = R_s$; therefore R_s represents the resistance of a square, irrespective of its dimensions (Fig.4). It is a constant, characteristic of the material employed. Because

$$\frac{R}{R_s} = \frac{L}{B} \tag{17}$$

any arbitrary two-dimensional model with a resistance R is equivalent to a rectangular model having an L/B ratio equal to R/R_s (Fig.5).

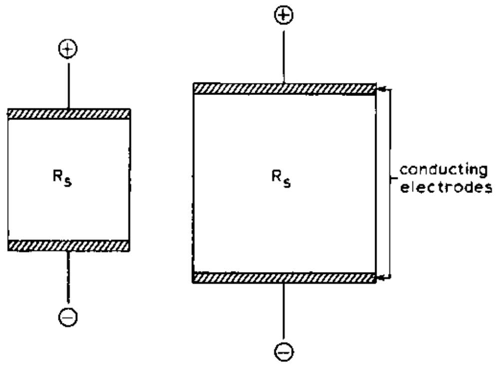

Fig.4. Two squares of a sheet conductor offer the same resistance to an electrical current.

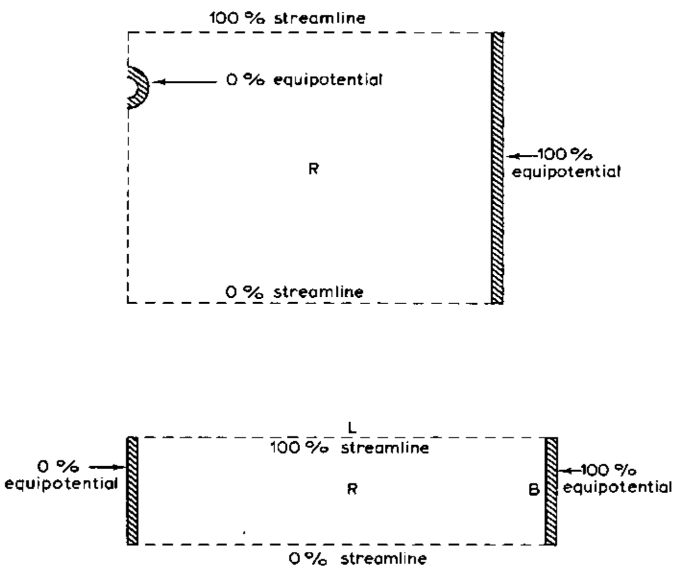

Fig.5. Transformation of an arbitrary region (A) into an equivalent rectangle (B).

This result is analogous to the transformations carried out in conformal mapping, in which an arbitrary two-dimensional region is mathematically transformed into an equivalent rectangle.

In an electrical model, the electrons automatically solve the problem for the given boundary conditions.
A two-dimensional groundwater flow in a rectangular region is described by

$$q = K \frac{\Delta h}{L} B$$

or

$$\frac{K\Delta h}{q} = \frac{L}{B} \tag{18}$$

where q is the two-dimensional discharge (m^2/day^{-1}).
Now, for given boundary conditions, L/B can be found from the electrical model.
Thus

$$\frac{K\Delta h}{q} = \frac{R}{R_s}$$

or, because $R = \Delta U/I$ (Eq.8), we may write

$$\frac{q}{K\Delta h} = \frac{R_s I}{\Delta U} = \frac{R_s}{S} \tag{19}$$

As both quotients are dimensionless, no scale factors enter into the considerations [*)]. The electrical model can therefore be constructed to any convenient scale; R can be measured and R_s is known.
Thus of three quantities, q, K, and Δh, each can be found if the other two are known.

7.4 STREAMLINES, EQUIPOTENTIAL LINES; BOUNDARY CONDITIONS

For two-dimensional flow in isotropic media, streamlines and equipotentials intersect at right angles: they are orthogonal. In a two-dimensional groundwater problem, the streamlines can be numbered, for instance from 0 for the first to 1.00 for the last. It is more convenient, however, to use percentages and to number the streamlines from 0 - 100 per cent.
The percentage Ψ^* thus obtained is a dimensionless number. It is related to the

[*)] For 3-dimensional cases, the use of a scale factor is necessary.

stream function Ψ of potential theory

$$\Psi^{*} = \frac{\Psi}{\Psi\ max} \times 100\ \% \tag{20}$$

Likewise, the equipotentials can be numbered from 0 - 100 per cent, taking 0% for the lowest potential and 100% for the highest. The percentage thus obtained is a dimensionless number. It is related to the potential function $\Phi = Kh$

$$\Phi^{*} = \frac{\Phi}{\Phi\ max} \times 100\ \% = \frac{Kh}{Kh_{max}} \times 100\ \% = \frac{h}{h_{max}} \times 100\ \% \tag{21}$$

Therefore, streamlines are defined by Ψ^{*} = constant, and equipotentials by Φ^{*} = constant.

For Ψ^{*} as well as for Φ^{*} the Laplace equation applies

$$\nabla^2\Psi^{*} = 0 \qquad \text{and} \qquad \nabla^2\Phi^{*} = 0 \tag{22}$$

Along the boundaries, either Φ^{*} or Ψ^{*} is defined in many cases, except along free groundwater surfaces and seepage surfaces which develop automatically in unconfined aquifers under the influence of gravity.
Whereas boundaries with given Φ^{*} or Ψ^{*} are easily simulated in electrical models (cf. Section 7.5.1), free surfaces are not, because there is no direct electrical equivalent for gravity.
The location of seepage surfaces is known beforehand, except for their upper limit, where they merge into a free surface. Of the latter, their location is entirely unknown and therefore the transition point and the free surface itself have to be found as a part of the solution of the flow problem.
In the saturated zone of the soil the hydraulic head h is defined by

$$h = z + p/\rho g$$

Along the free surfaces and seepage surfaces $p = 0$ and $h = z$. If the range for z is taken the same as for h (from 0 to h_{max}), the height may be expressed as a percentage of the total difference in head. The percentage z^{*} thus obtained is a dimensionless number. It is related to the actual height z

$$z^{*} = \frac{z}{h_{max}} \times 100\ \% \tag{23}$$

With $h = z$ and $\Phi^{*} = \frac{h}{h_{max}} \times 100$ %, it follows that

$$\Phi^{*} = z^{*} \qquad (24)$$

which is valid along free surfaces and seepage surfaces.
Sometimes the free surface is a streamline (e.g. in the case of steady unconfined flow between two ditches with different levels), but in other cases it is neither a streamline nor an equipotential (e.g. with rainfall between ditches).
Seepage surfaces are neither streamlines nor equipotential lines.
Table 3 gives a summary of the boundary conditions usually encountered.

Table 3. Boundary conditions for steady groundwater flow problems.

Kind of boundary	Stream function	Potential function
Streamline	Ψ^{*} = constant	-
Impervious boundary	Ψ^{*} = constant	-
Line of symmetry	Ψ^{*} = constant	-
Equipotential	-	Φ^{*} = constant
Free groundwater surface	sometimes Ψ^{*} = constant	$\Phi^{*} = z^{*}$
Seepage surface	-	$\Phi^{*} = z^{*}$

7.5 SIMULATION OF BOUNDARY CONDITIONS IN ELECTRICAL MODELS

Streamline boundaries are simulated by cutting off the conductive sheet at the desired locations with a pair of scissors or a razor blade (Fig.6).

Equipotential boundaries are created by putting conductive electrodes at the desired locations. Metal foil or wire pasted to the conductive sheet can be used for the construction of electrodes. A convenient method is to apply silver paint along the boundaries. In selecting the electrode material, care must be taken to ensure that the resistance of the electrode itself is negligible compared with the resistance of the conductive sheet. With the aid of a suitable voltage source (e.g. a simple dry cell), the electrodes are held at the required potential.

Recharge from an equipotential surface (e.g. ponded water can be simulated in the same way by feeding the electrode, corresponding with the equipotential surface line of the model.

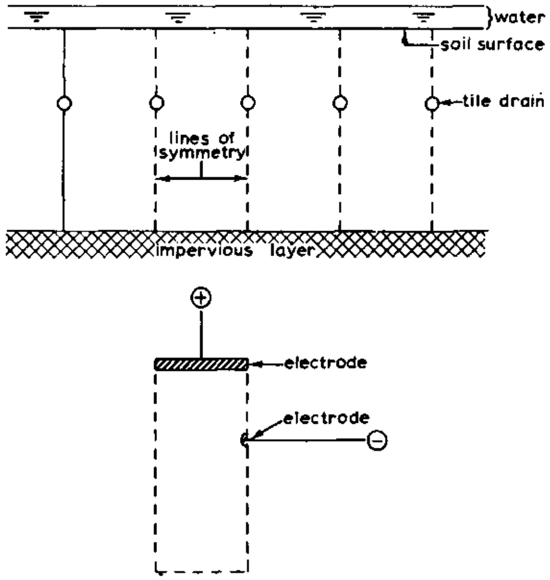

Fig.6.
Simulation of a groundwater flow problem with simple boundary conditions (ponded water case).

Recharge, evenly distributed along a surface (e.g. recharge from rainfall) is simulated by feeding the model at its upper side with a large number of equal currents at equal intervals. A convenient method is to feed the model from a high tension source (e.g. 100 volts) through a number of resistors. The resistors should have a resistance which is high compared to the resistance of the model. In this way each of the resistors will convey an approximately equal current, thus simulating an evenly distributed recharge (Fig.7).

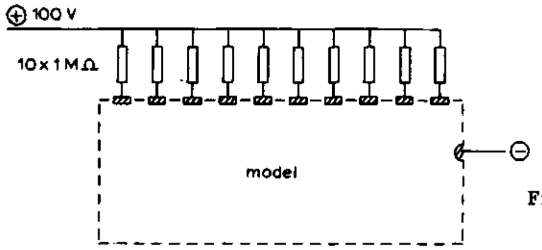

Fig.7. Simulation of evenly distributed recharge (e.g. caused by rainfall).

Free water surfaces cannot be obtained directly. In soils such surfaces form under the influence of gravity, to which electrons are not subjected. In nature free surfaces are governed by the relation

$$\Phi^{*} = z^{*}$$

In electrical models its equivalent is

$$\Phi^{*} = \zeta^{*} \tag{25}$$

where

$$\Phi^{*} = \frac{U}{U_{max}} \times 100\ \%$$

and ζ^{*} is the relative height (expressed as percentage) in the model.

In the model we represent
$\zeta^{*} = 0\%$ by a horizontal level corresponding with $h = 0$ in nature,
$\zeta^{*} = 100\%$ by a horizontal level corresponding with $h = h_{max}$ in nature.
If no recharge occurs at the water table, its surface is a streamline. By trial and error the upper boundary of the model is reshaped until the required condition is fulfilled. This is done by removing parts of the conductive sheet until the "potentials" Φ^{*} correspond with the "elevation" ζ^{*} (Fig.8).

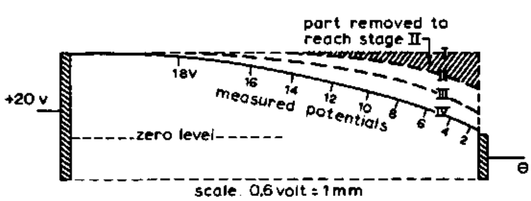

Fig.8. Location of a free water surface. I is initial model, II and III are intermediate stages, IV is final solution

If recharge occurs along the free water table (e.g. by rainfall) we can start with a model as shown in Fig.7. The free water surface is determined by the trial-and-error method of making successive incisions with a pair of scissors between the small electrodes, each time as far as the free water surface at that particular stage (Fig.9).

<u>Seepage surface</u> likewise obey condition (25). Their position is known beforehand, except for their upper limit, where they merge into a free water surface.

Water is flowing out along such surfaces so they are neither streamlines nor equipotentials.
Condition (25) is simulated by a large number of contacts along the expected seepage surface, each contact being held at the appropriate potential. A series of resistors (with a resistance considerably lower than the resistance of the model) can be used as a potential divider for this purpose. The contacts are most conveniently made by applying a strip of silver paint along the expected seepage surface and cutting this strip into separate units (Fig.10).

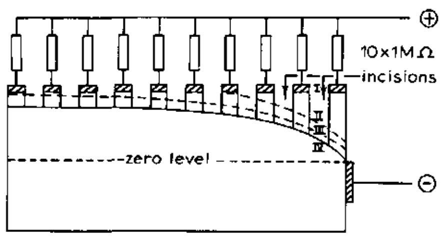

Fig.9. Location of a free groundwater surface formed by evenly distributed recharge.

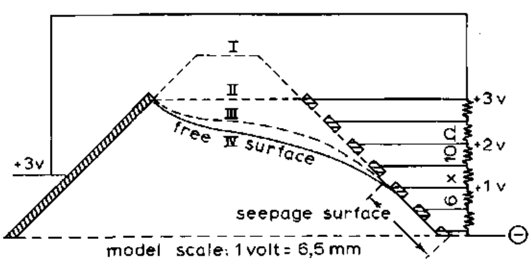

Fig.10. Seepage through a dam. Location of free groundwater surface and seepage surface.

7.6 MEASURING IN THE MODEL

7.6.1 MEASUREMENT OF DISCHARGE

For this purpose the total resistance of the model, R, must be measured with a Wheatstone bridge and R_s, K, and Δh must be known. From Eq.19, we know that the total resistance, R, determines the ratio $q/K\Delta h$ of the original groundwater flow.

7.6.2 MEASUREMENT OF EQUIPOTENTIAL LINES

Equipotential lines in the interior of the model are found by using a potential analyzer (a combination of a dry cell, a potentiometer, and a galvanometer) and a ballpoint as probe. In this way the equipotentials can be drawn directly on the conductive sheet. After the dry cell is connected to the electrodes of the model, the potentiometer is adjusted to the voltage corresponding with the equipotential line to be plotted, say the 60% line. The probe is then moved over the paper in such a way that the deflection of the galvanometer remains zero. The 60% equipotential line ($\Phi^{*} = 60$) is found as the locus of all points where the galvanometer reading is zero (Fig.11).

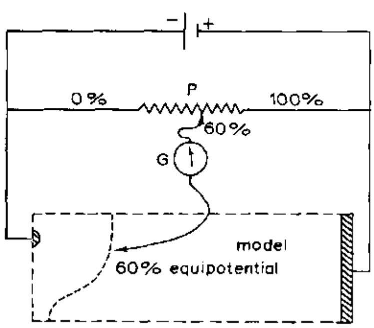

Fig.11. Location of equipotentials.

7.6.3 MEASUREMENT OF STREAMLINES, INVERTED MODEL

Streamlines cannot be measured directly because electrons can neither be marked nor coloured[1]. But since Eq.22 states that the stream function and the equipotential function obey the same laws, we can apply the concept of duality and construct an inverted model, interchanging Ψ and Φ along the boundaries. The equipotential lines of the inverted model correspond with the streamlines of the original model (Fig.12).

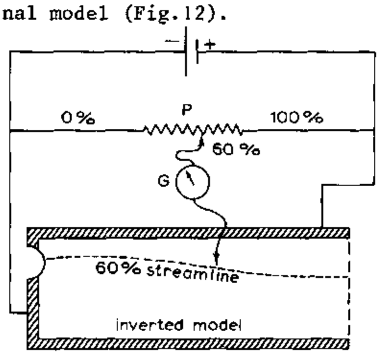

Fig.12. Location of streamlines by plotting the equipotentials in an inverted model.

[1] In electrolytic models, where the current is transported by ions, coloured ions are sometimes used for generating streamlines.

Table 4 gives a summary of the changes caused by the inversion and of the boundary conditions to be simulated in the inverted model.

Table 4. Conditions in inverted models.

Original model	Inverted model	In inverted model	
		stream function	potential function
Streamline	equipotential	-	Φ^{*} = constant
Impervious boundary	equipotential	-	Φ^{*} = constant
Line of symmetry	equipotential	-	Φ^{*} = constant
Equipotential	streamline	Ψ^{*} = constant	-
Free groundwater surface	same surface	$\Psi^{*} = \zeta^{*}$	sometimes Φ^{*} = constant
Seepage surface	same surface	$\Psi^{*} = \zeta^{*}$	-

In simple cases the inversion is easy, but with more complicated boundary conditions, variable resistors are employed, which are adjusted until the required condition ($\Psi^{*} = \zeta^{*}$) is fulfilled.

7.7. EXAMPLE: DETERMINING THE FREE WATER SURFACE IN A DAM

Figure 13A shows a cross-section of a dam whose faces have an inclination of 45 degrees. The base of the dam is chosen as a datum plane. The exact form of the free water surface BF and the length of the seepage surface FG are unknown and will be determined in the model.

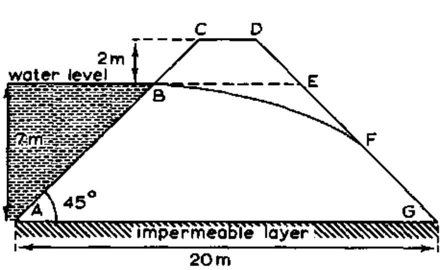

Fig.13A. Dam faces having an inclination of 45 degrees.

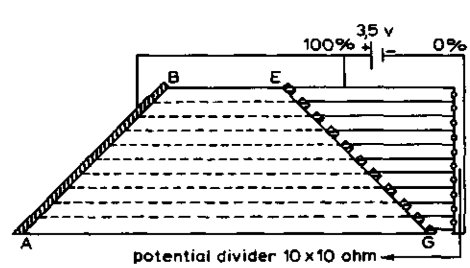

Fig.13B. Electrical model of the dam, scale 1 : 100.

The boundaries of the model can be simulated as follows (Fig.13B).
The base of the dam is a streamline, so that the boundary AG can be simulated by cutting the paper to shape. The inflow face is an equipotential line and the boundary AB is represented by a silver paint electrode. The electrode has a width of a few millimeters; the right-hand side of the electrode AB in Fig.13B is the boundary of the model. The boundary corresponding with the free water surface is unknown. It is assumed that BE is the first stage of the free surface. Since the free surface is a streamline, the boundary BE can be simulated by cutting the paper to shape. Along the outflow face, the electrical potential has to correspond with the height above the datum plane. At the outflow face, a continuous electrode of silver paint along the boundary EG is split up into 11 separate strips. For practical reasons, the width of a strip of electrode equals the distance between two strips. A slight error is introduced by representing a continuous seepage surface as a discontinuous series of electrodes in the model.
The procedure to be followed in constructing the model ABEG to scale 1 : 100 is as follows:

- take a piece of Teledeltos paper;
- apply silver paint along the boundary AB; long electrodes of silver paint should be provided with a strip of aluminium foil to improve their conductivity, so cut a strip of aluminium foil, 15 cm long and 0.5 cm wide;
- fix the strip of aluminium foil to the silver paint before the paint is dry;
- fold over the free end of the aluminium strip to make it sufficiently strong for a clip to be attached;
- apply silver paint along boundary EG;
- divide electrode EG into 11 separate strips;
- fix thin wires to these strips using silver paint;
- attach the model to a piece of cardboard using adhesive tape;
- attach a potential divider consisting of 10 resistors of 10 ohms each along EG on the cardboard;
- connect the thin wires from the strips along EG to the potential divider;
- with a ballpoint draw lines on the model parallel to AG (dotted lines in Fig. 13B), dividing the distance between BE and AG into 10 equal parts. These lines can be numbered as the 10%, 20% .... 90% lines for ζ^{*};
- allow the model to dry for several hours.

The resistance of the potential divider is much less than the resistance of the

model, so that the current in the divider exceeds the current in the model, and the variation of the potential along EG remains closely linear.
As the potential along the free surface should vary linearly with its vertical height above AG, the potential at the intersection of the free surface and, for instance, the 90% height line ($\zeta^{*} = 90$) should be 90% of the voltage difference (e.g. 3.5 Volt). The intersection of the free surface with the 100% height line is already known.

To determine the free surface:

- connect the electrode AB and the terminals of the potential divider to the contacts of the potential source in the potential analyzer;
- adjust the desired voltage difference;
- set potentiometer of the potential analyzer at 90%;
- move the measuring probe over the paper from left to right along the 90% height line till the meter reading is zero and mark this point;
- follow the same procedure for the 80% height line, and so on.
  In this way points are being found for the first approximation of the free water surface;
- draw a line through the located points;
- remove part of the paper at the upper side of this line and repeat the measuring process.

The electric current will traverse the whole of the conducting model. By cutting off the paper, the physical boundaries of the model are changed and the whole process has to be repeated. After this has been done three of four times (Fig. 13C), the final stage is reached and the free water surface in the dam and the length of the seepage surface are determined.

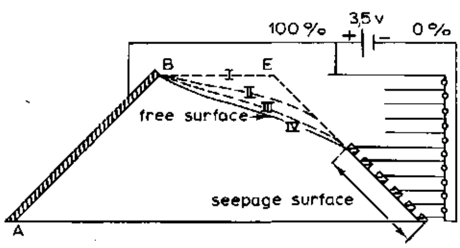

Fig.13C. Final solution of the free surface.

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PRINCIPAL SYMBOLS USED IN VOLUME I

SYMBOL	DESCRIPTION	DIMENSION
A	cross-sectional area; horizontal surface area	L^2
B	width;	L
	bulk density	ML^{-3}
CEC	cation exchange capacity (in meq/100 g soil)	
c	hydraulic resistance to vertical flow	T
D	thickness; saturated thickness of water bearing layer	L
D_w	soil water diffusivity	L^2T^{-1}
d	diameter	L
EC_e	electrical conductivity of saturation extract	$ohm^{-1}cm^{-1}$
ES	exchangeable sodium (in meq/100 g soil)	
ESP	exchangeable sodium percentage	dimensionless
e	void ratio	dimensionless
F	force	MLT^{-2}
F_k	specific force	LT^{-2}
g	acceleration due to gravity	LT^{-2}
H	height of water table midway between two drains	L
h	hydraulic head or potential head ($= p/\rho g + z$) (energy per unit weight)	L
h_g	elevation head (= z)	L
h_m	matric head	L
h_p	hydrostatic head	L
I	electrical current	amp
i	hydraulic gradient; imaginary unit ($i^2 = -1$)	dimensionless
J	current density	amp L^{-2}
K	hydraulic conductivity	LT^{-1}
K'	intrinsic permeability	L^2
KD	transmissivity for horizontal flow	L^2T^{-1}
k	conductivity coefficient	$M^{-1}L^3T$
k_h	capillary conductivity	LT^{-1}
k_w	capillary conductivity coefficient	$M^{-1}L^3T$
L	length; drain spacing	L
M	mass	M
mho	reciprocal ohm	ohm^{-1}
mmho	milli-mho (= mho $\times 10^{-3}$)	ohm^{-1}

SYMBOL	DESCRIPTION	DIMENSION
n	porosity	dimensionless
P	hydrostatic pressure	$ML^{-1}T^{-2}$
p	relative hydrostatic pressure	$ML^{-1}T^{-2}$
	(p = 0 at atmospheric pressure)	
Q	discharge; rate of fluid flow	L^3T^{-1}
q	discharge per unit length;	L^2T^{-1}
	discharge per unit horizontal surface area	LT^{-1}
R	radius;	L
	recharge to groundwater per unit horizontal surface area;	LT^{-1}
	electrical resistance	ohm
r	radius; radial distance	L
r_e	radius of influence of well	L
r_w	radius of well	L
SAR	sodium adsorption ratio	dimensionless
t	time	T
U	electrical potential	volt
$\overline{V}$	specific volume (= $1/\rho$)	$M^{-1}L^3$
V_s	volume of solids	L^3
V_v	volume of voids	L^3
v	effective flow velocity (= Q/A)	LT^{-1}
W	work of energy	ML^2T^{-2}
w	volumetric water content	dimensionless
x, y, z	Cartesian coördinates	L
y	height of water level in channel, ditch or drain	L
z	elevation head	L
z^*	elevation head expressed in percentage of z_{max}	dimensionless
α	contact angle between fluid and surface	dimensionless
γ	specific weight	$ML^{-2}T^{-2}$
η	dynamic viscosity	$ML^{-1}T^{-1}$
θ	mass fraction of water in the soil	dimensionless
μ	effective porosity or drainable pore space	dimensionless
ν	kinematic viscosity (= η/ρ)	L^2T^{-1}
Π	osmotic pressure	$ML^{-1}T^{-2}$
ρ_a	mass density of air	ML^{-3}

SYMBOL	DESCRIPTION	DIMENSION
ρ_w	mass density of water	ML^{-3}
σ	specific conductivity;	$ohm^{-1}\ L^{-1}$
	surface tension	MT^{-2}
Φ	potential function (= Kh)	L^2T^{-1}
Φ^*	potential function expressed in percentage of Φ_{max}	dimensionless
ϕ	potential of water or specific energy (energy per unit mass)	L^2T^{-2}
ϕ_g	gravitational potential	L^2T^{-2}
ϕ_m	matric potential	L^2T^{-2}
ϕ_o	osmotic potential	L^2T^{-2}
ϕ_p	hydrostatic potential	L^2T^{-2}
ϕ_h	hydraulic potential (= $\phi_g + \phi_m$ or $\phi_g + \phi_p$)	L^2T^{-2}
ϕ'	pressure potential of water (energy per unit volume)	$ML^{-1}T^{-2}$
$\phi_m(S_m)$	matric pressure (suction)	$ML^{-1}T^{-2}$
ϕ'_h	hydraulic pressure	$ML^{-1}T^{-2}$
ϕ'_p	hydrostatic pressure	$ML^{-1}T^{-2}$
Ψ	stream function	L^2T^{-1}
Ψ^*	stream function expressed in percentage of Ψ_{max}	dimensionless
ω	complex potential function (= $\Phi + i\Psi$)	L^2T^{-1}
Δ	small increment of	
∂	partial differential of	
∇	differential operator	
∇^2	Laplacean operator	

SUBJECT INDEX

Adhesion of water 2.3.3
  capillary rise 5.2.2

Adsorption complex of the soil 2.2.3

Adsorption of water 5.2.2

Aeration see also Diffusion, Gases, Porosity, Soil aeration

Aeration 2.5

Aeration pores 2.3.2

Aeration porosity 2.5.1

Aggregate structure 2.3.2

Aggregation
  influence of drainage 4.3.1
  influence of organic matter 2.2.4

Agrological soil profile 2.1.4

Air see Soil air

Alkalinity of soil 3.4
  effects on structure 4.6

Alumino-silicates 2.2.2

Alluvial fans 1.3.2

AM see Available moisture

Amendments
  structure improving 3.5

Angular blocky structure 2.3.2

Anion adsorption capacity 2.2.3

Apparent density 2.3.1

Apparent velocity 6.3.1

Aquatic plants 4.2.2

Aquifers
  classification 1.2

Arable crops see also Crops

Arable crops
  influence of drainage 4.8.2
  oxygen needs 2.5.4

Artesian water 1.2

Atmospheric air
  composition 2.5.1

Atterberg limits 2.3.3

Available moisture see also Soil moisture

Available moisture 2.4.2; 4.2.1

Backswamps 1.3.1

Bacteria 2.2.4
  nitrogen fixation 2.7.2
  respiration 2.5.1

Bahada 1.3.2

Base flow 1.3

Base level 1.3

Base saturation 2.2.3

Basins 1.3.1

Beach formations 1.4

Bernoulli's energy equation 6.2.5

Biological investigations of soil 2.1.5

Biopores 2.3.2

Black alkali 3.4

Blocky structure 2.3.2

Bog plants 4.2.2

Bottom-set beds 1.3.3

Boulder clay 1.6

Boundaries see also Streamlines

Boundaries
  free water surface 6.7.3
  impervious 6.7.1
  planes of symmetry 6.7.2
  seepage face 6.7.5
  simulation at the model 7.7
  with slowly moving water 6.7.4

Boundary conditions of the flow region 6.5.1; 6.7; 7.1
  electrical model 7.4
  simulation in inverted models 7.6.3
  simulation in models 7.5

Brackish transition layer 1.3.3

Braided rivers 1.3.1

Brak 3.4

Brown soils 2.3.4

Bulb crops see Arable crops

Bulk density 2.3.1

Capillary conductivity 5.4.3

Capillary flow
    function of pores 2.3.2

Capillary fringe 4.2.2
    definition 6.1

Capillary pores 4.2.1

Capillary rise 5.4.3
    water adhesion 5.2.2

Capillary water 6.1

Carbon dioxide 2.5.1; 2.5.3

Cation exchange capacity 2.2.3
    salty soils 3.4

Cauchy-Rieman conditions 6.6.2

CEC see Cation exchange capacity

Cementation 2.3.3

Cereal crops see Arable crops

Changes in water storage 6.5.3

Chemical fertility of soil 2.7.1

Chemical investigations of soil 2.1.5

Chroma 2.3.4

Classification
    salty soils 3.4
    soil colour 2.3.4
    soil structure 2.3.2
    soil texture 2.2.1
    water-bearing layers 1.2

Clay
    particle size limits 2.2.1
    physico-chemical characteristics 2.2.3
    soil composition 2.2.2
    textural classes 2.2.1

Clay fraction 2.2.2

Clay loam
    textural classes 2.2.1

Clay minerals 2.2.2
    cation exchange capacity 2.2.3

Clay soils
    cation composition 3.3
    salt concentration 3.3

Cliffs 1.5

Climatic conditions
    crop yields 2.7.1
    introduction of reclamation crops 4.6

Clods 2.3.2

Coastal plains 1.4

Coefficient of conductivity 5.4.2

Coherent structures
    soil 2.3.2

Cohesion
    soil 2.3.3
    water molecules 5.2.2

Colloids see Organic colloids

Colour
    soil 2.3.4

Compaction of soil 2.3.3
    unfavourable results 4.4.1

Complex variable 6.6.3

Conductive sheet analogues 7

Conductivity coefficient see also Hydraulic conductivity, Specific conductivity

Conductivity coefficient 5.4.2

Conductivity of soil 3.4

Confined aquifer 1.2

Confined water 1.2

Consequent streams 1.4

Conservation of energy
    law 6.2.5

Consistence limits 2.3.3

Consistence of soil 2.3.3

Content of water see Field capacity, Retention

Continental shelf 1.4

Continuity
    equation 5.4.1; 6.2.6; 7.2.1

Convection of soil air 2.5.3

Conversion factors for water potential 5.3.1

Crop drainage 2.5.6

Crop production
- effects of salinity 4.6
- groundwater level 4.8

Crops see also Plant development

Crops
- drainage 4.8
- oxygen needs 2.5.4; 4.2.1
- root penetration 4.2.2
- salt tolerance 4.6
- soil moisture conditions 4.2.2

Crumb
- soil structure 2.3.2

Crust
- formation 4.6
- soil aeration 2.5.3

Cuesta landscapes 1.4

Cultivation practices
- influence of drainage 4.4

Current density 7.2.2

d'Alembert-Euler conditions 6.6.2

Dams
- determination of the free water surface 7.7

Darcy K 6.3.2

Darcy's law 6.3; 7.2.1
- applications 6.4
- hydraulic conductivity 2.4.1; 6.3
- proportionality constant K 6.3.2

Decomposition
- soil organic matter 2.2.4

Deep ploughing
- non-saline sodic soils 3.5

Deflation 1.7

Deltas 1.3.3

Density of soil particles 2.3.1

Density of water 5.2.2; 6.2.1
- temperature influence 6.3.3

Depth of root penetration 4.2.2

Deposits see also Sediment

Deposits
- in coastal plains 1.4
- in deltas 1.3.3
- in lacustrine plains 1.5
- in river plains 1.3.1
- loess 1.7

Desert loess 1.7

Desert regions 1.7

Diffuse layer 2.3.3

Diffusion of soil air 2.5.3

Discharge
- Darcy's law 6.3.1
- of rivers 1.3
- of wells 6.5.7
- to ditches 6.5.6
- to drains 6.5.1
- through cross sections 6.5.2

Discontinuous hydraulic conductivity 6.6.1

Diseases of plants
- influence of drainage 4.7

Dissolution of limestone 1.8

Dissolved salts
- concentration 5.3.2; 5.4.1
- osmotic potential 5.3.2

Drainable pore space 6.5.3

Drainage
- arable crops 4.8.2
- cultivation practices 4.4
- definition 4.1
- fruit trees 4.8.3
- grassland 4.8.1
- physical soil conditions 4.3
- plant diseases 4.7
- yields 4.5

Drainage flow 6.5.1

Dry weight soil moisture fraction 5.1

Dune ridges 1.4

Dupuit-Forchheimer assumptions 6.5.2

Dynamic viscosity of water 6.2.2

Effective flow velocity 6.3.1

Electrical conductivity of soil 3.4

Electrical models
measurements 7.6
simulation of boundary conditions 7.5
two-dimensional 7.3.3
types 7.3.1

Electricity flow 7.2.2

Elevation head 6.2.4; 6.2.5

Energy units 5.3.1

Ephemerals 4.2.2

Equation of continuity 5.4.1; 6.2.6; 7.2.1

Equipotential lines 6.6.2
electrical model 7.4
flow net 6.6.2
measurement at the model 7.6.2
simulation in models 7.5.1

Erosion
coastal plains 1.4
glacial plains 1.6
streams 1.3

Evaporation rate 5.4.3

Excess water 4.1

Excessive soil moisture 4.1

Exchangeable bases 2.2.3

Exchangeable calcium
salty soils 3.2

Exchangeable cations
salty soils 3.2
thickness 3.4
tropical soils 2.2.3

Exchangeable sodium percentage
replacement 3.5
salty soils 3.4

Falling head permeameter 6.4.1

Fan see Alluvial fans

Faults 1.9

Fertility of soil
definition 2.7.1
supply of nutrients 2.7.2

Fertilizers
application by drainage 4.5
nitrogen supply 2.7.2

Field capacity see also Soil. moisture

Field capacity 2.4.2; 5.2.1
definition 5.2.3

Field investigations 2.1.5

Film water 2.3.2

Firm soils 2.3.3

Flood plain 1.3.1

Flow equations
general 5.4.1
groundwater 6.5

Flow net 6.6.2

Flow of electricity 7.2.2

Flow rate see Discharge

Flow velocity 5.4.1
effective 6.3.1

Flowering
retardation by soil salinity 4.6

Fluvioglacial deposits 1.6

Forces see Potentials

Forchheimer equation 6.5.2

Fore-set beds 1.3.3

Fossil coastal features 1.5

Fractures 1.9

Fragments 2.3.2

Free-surface flow 6.5.2
solution 6.5.1

Free water surface 6.7.3
determination 7.7
inverted model 7.6.3
model 7.4
simulation in models 7.5.2

Free water table see Groundwater table

Fresh-water head 6.3.4

Friability of soil 2.3.3

Friction see Viscosity

Frost
effects on soil 4.3.4

Frozen soils 1.6

Fruit trees
influence of drainage 4.8.3

Fully penetrating well 6.5.7

Fungous diseases see Diseases

Galvanometer
    measurement of equipotential lines 7.6.2

Gases
    interchange 2.5.3
    soil 2.5.1

Glacial lakes 1.6

Glacial loess 1.6; 1.7

Glacial plains 1.6

Glacial till 1.6

Glaciers 1.6

Gley 2.3.4

Gouy-Chapman Diffuse Double Layer Theory 2.2.3; 3.3

Graben 1.9

Gradient see also Hydraulic gradient, Potential gradient, Velocity gradient

Gradient
    hydraulic head 6.3.1
    streams 1.3

Grain structure 2.3.2

Granular structure 2.3.2

Grassland
    influence of drainage 4.8.1

Gravitational potential of water 5.3.2

Gravity see Specific gravity

Grey soils 2.3.4

Groundwater
    basic laws 6.2
    definition 6.1
    physical properties 6.2
    potential 6.2.4
    temperature influence on viscosity 6.3.3

Groundwater conditions
    alluvial fans 1.3.2
    areas affected by dissolution 1.8
    coastal plains 1.4
    deltas 1.3.3
    flood plains 1.3.1
    glacial plains 1.6
    lacustrine plains 1.5
    loess plains 1.7
    valleys 1.3.1

Groundwater flow 6.2; 6.3.1; 7.2.1
    analogy to electricity 7.2.3
    basic equations 6.5
    Darcy's law 6.3; 7.2.1
    electrical models 7.3
    linear 6.4.2
    through stratified soils 6.4.2
    two-dimensional 6.6; 7.2.1

Groundwater hydraulics 6

Groundwater level 1.3.1

Groundwater mound 1.3.1

Groundwater surface 6.7.3
    simulation at the model 7.4

Groundwater table
    definition 6.1

Groundwatertable depth
    in relation to crop production 4.8
    in relation to temperature effects 4.3.4
    in relation to yields 4.5

Gypsum
    application in saline soils 3.5
    requirements 3.5

Habitat of plants 4.2.2

Hagen-Poisseuille equation 6.3.2

Heat capacity of soils
    influence of drainage 2.6.2

Heavy fraction 2.2.2

Heavy soils 2.2.1

Horst 1.9

Hue 2.3.4

Humus see also Soil organic matter

Humus
    cation exchange capacity 2.2.3
    organic matter 2.2.4

Hydraulic
  conductivity 2.4.1; 6.3; 7.2.1
    discontinuous 6.6.1
    isotropy 6.4.2
    temperature influence 6.3.3

Hydraulic flow 6.5.2

Hydraulic gradient 6.3.1

Hydraulic head
    definition 6.2.4
    flow through stratified
      soils 6.4.2
    measurement 6.3.4
    seepage surface 6.7.5

Hydraulic potential 5.3.2
    gradient 5.4.1

Hydraulic resistance 6.4.2

Hydraulics
    groundwater 6

Hydrodynamics
    basic laws 6.2
    conservation of energy 6.2.5
    conservation of mass 6.2.6
    Darcy's law 6.3

Hydrogeology
    in relation to drainage 1.1

Hydrological soil profile 2.1.4

Hydrophytes 4.2.2

Hydrostatic pressure 5.2.2
    groundwater 6.2.4

Hygroscopic water 2.3.2

Ice
    erosion 1.6

Illite
    cation exchange capacity 2.2.3

Impervious boundaries 6.7.1
    inverted model 7.6.3
    model 7.4
    simulation in models 7.5

Impervious layers see Aquifers

Improving amendments
    saline soils 3.5

Infiltration 5.4.2
    reduction by soil salinity 4.6

Infiltration rate 2.4.1

Influent seepage see also
  Seepage surface

Influent seepage 1.3.1

Initial conditions of the flow
  region 6.5.1

Intake rate 2.4.1

Intercalation
    impervious 1.7

Interglacial period 1.6

Internal drainage
    restauration by tillage 4.4.1
    rootzone 2.4.1

Intrinsic permeability 6.3.2

Inverted models see also
  Electrical models

Inverted models
    measurement of stream-
      lines 7.6.3

Investigations of soil 2.1.5

Ionic double layer 2.2.3

Iron in tropical soils 2.2.2

Isotropic soils
    hydraulic conductivity 6.4.2

Kaolinite
    cation exchange capacity 2.2.3

Karst topography 1.8

Kinematic viscosity of water 6.2.2

Laboratory investigations 2.1.5

Lacustrine clays 2.2.1

Lacustrine plains 1.5

Lagoons 1.4

Lake plains 1.5

Lakes
    classification 1.5
    floors 1.5
    sediments 1.5

Laminar flow 6.3.1

Land form
- factor of soil formation 2.1.3
- glacial 1.6

Land-ice sheets 1.6

Laplacean operator 6.5.1; 7.2.1

Laplace's equation 6.5.1; 7.2
- analogy to electricity 7.1
- relaxation method of solution 6.6.4

Lattice see Structural units

Laws
- conservation of energy 6.2.5
- conservation of mass 6.2.6
- Darcy 6.3; 7.2.1
- linear resistance 6.3.1

Layers see Aquifers

Leaching of salts
- in arid regions 1.5
- reclamation problems 3.5

Leaky aquifer 1.2

Leguminous crops
- influence of drainage 4.8.2

Levee
- coastal plains 1.4
- deltas 1.3.3
- flood plain 1.3.1

Light fraction 2.2.2

Light soils 2.2.1

Limestone
- dissolution 1.8
- karst 1.8
- loess plains 1.7

Limon see Silt

Linear flow of groundwater
- through stratified soils 6.4.2

Linear resistance
- law 6.3.1

Liquid limit 2.3.3

Loam
- textural classes 2.2.1

Local base level
- lakes 1.3

Loess
- deposits 1.7
- desert 1.7
- glacial 1.6
- plains 1.7

Loess-like deposits 1.7

Macro-aggregates 2.3.2

Macrofauna in soil 2.2.4

Macropores 2.3.2

Macrostructures 2.3.2

Maize
- nutrient losses 4.5
- yields 4.8.2

Manure see Fertilizers

Mass density see Density

Mass flow of soil air 2.5.3

Massive structure 2.3.2

Matric potential of water 5.3.2

Matric suction 5.2.3

Meandering rivers 1.3.1

Meanders 1.3.1

Meltwater
- glacial 1.6

Mesophytes 4.2.2

Mesopores 2.3.2

Micro-aggregates 2.3.2

Microorganisms see also Bacteria

Microorganisms in soil 2.2.4

Micropores 2.3.2

Mineral soils
- physical characteristics 2.3
- porosity 2.3.1

Mineralogical composition of soil 2.2.2

Mineralogical investigations of soil 2.1.5

Models see Electrical models

Moisture see Soil moisture

Moisture content 2.3.3

Montmorillonite
- cation exchange capacity 2.2.3
- clay fraction 2.2.2

Moraines 1.6

Mottling 2.3.4

Moulding
capacity of the soil 2.3.3

Movement of water see Hydrodynamics, Transport

Munsell colour system 2.3.4

Negative pressure
measurement 5.2.3

Nitrates
effects on aeration 4.5

Nitrification see also Nitrogen fixation

Nitrification 2.7.2

Nitrogen fertilizer
crop yields 4.5

Nitrogen fixation
influence of aeration 2.5.5
influence of drainage 4.5
supply of nutrients 2.7.2

Non-capillary pores 2.3.2

Non-saline sodic soils 3.4
reclamation 3.5

Nonsteady flow 6.5.3

Nutrient losses 4.5

Nutrient status
improving by tillage 4.4.1
soil 2.7.1

Nutrient supply 2.7.2
influence of drainage 4.5

Off-shore sediments 1.4

Ohm's law 6.3.1; 7.2.2

One-dimensional flow
electricity 7.2.2
groundwater 6.5.2; 7.2.1

Operator (Laplace) 6.5.1; 7.2.1

Organic colloids
cation exchange capacity 2.2.3

Organic matter see Soil organic matter

Organic residues 2.2.4

Organic soils
drainage 4.3.3

Osmotic potential of water 5.3.2

Out-wash deposits 1.6

Ox-bow lakes 1.3.1

Oxygen
diffusion 2.5.3
plant requirements 2.5.4
soil air 2.5.1

Parent material
soil 2.1.3

Partially penetrating well 6.5

Particle density see also Soil particles

Particle density 2.3.1

Particle size limits
american 2.2.1
french 2.2.1

Pasture grasses
influence of drainage 4.8.1

Path lines 6.6.2

Peak discharge 1.3

Peak flow
braided rivers 1.3.1

Pedological soil profile 2.1.4

Peds 2.3.2

Perched water table 2.4.1

Percolating water
dissolution of limestone 1.8

Percolation see Recharge

Percolation capacity
reduction by soil alkalinity 4.6

Percolation rate 2.4.1

Periglacial climatic conditions 1.6; 1.7

Permafrost 1.6

Permanent wilting point see Wilting point

Permeability see also Hydraulic conductivity

Permeability
- clay 2.2.3
- definition 6.3.2
- loess deposits 1.7
- soil 2.2.1; 2.4.1

Permeameter 6.4.1

Pervious layers see Aquifers

Pests see Diseases

pH
- saline soils 3.4
- soils 2.2.3

Phreatic aquifer 1.2

Phreatic surface see Groundwater table

Phreatic water 1.2

Phreatophytes 4.2.2

Physical agricultural qualities of soil
- definitions 2.1.6

Physical investigations of soil 2.1.5

Physical soil properties
- definitions 2.1.6
- influence of drainage 4.3

Physical units 5.2.2

Physics of soil moisture 5

Physiological drought 2.5.4

Piedmont plain 1.3.2

Piezometer 5.2.3; 6.3.4

Piezometric head 1.2
- groundwater 6.2.4; 6.3.4

Plant aeration requirements 2.5.4

Plant growth requirements 2.5.4
- reclamation 3.5
- salty soils 3.4
- soil temperature 2.6.1

Plant development
- effects of salinity 4.6
- reduced conditions 4.2.2
- salt content 3.4
- soil-water relations 4.2.1

Plastic limit 2.3.3

Plastic number 2.3.3

Plasticity index 2.3.3

Plasticity of soil 2.3.3

Platy structure 2.3.2

Playas 1.5

Pleistocene 1.3.1

Ploughing
- in relation to texture 2.2.1
- influence of soil consistence 2.3.3
- soil aeration 2.5.6

Polders 4.8

Pore size diameter 2.5.2

Pore size distribution 2.3.2

Pore space see also Capillary pores

Pore space 2.3.1
- influence of drainage 4.2.1; 4.3.1

Porosity see also Particle density

Porosity 2.3.1
- aeration 2.5.1

Potential analyzer 7.6.2

Potential fertility of soil 2.7.1

Potential function 6.6.3; 6.6.2; 7.4

Potential gradient
- electricity 7.2.2

Potential lines
- method of squares 6.6.4

Potential of groundwater 6.2.4
- measurement 6.3.4

Potential of water flow
- analysis of water flow 5.4.1
- definition 5.3.1

Potentiometer
- measurement of equipotential lines 7.6.2

Pressure head 6.2.4; 6.2.5

Pressure units 5.3.1

Primary soil particles 2.3.2

Prismatic structure 2.3.2

Productive capacity of soil 2.7.1

Profile
    soil 2.1.4
    streams 1.3
    structural 2.3.2

Proportionality constant K 6.3.2

Proportionality factor see Darcy's law, Hydraulic conductivity

Quartz 2.2.2
    cation exchange capacity 2.2.3

Quaternary loess deposits 1.7

Radial flow see Two-dimensional groundwater flow

Radial symmetry 6.5.7

Rainfall
    recharge from 6.5.6

Recharge by rainfall 6.5.6
    simulation in models 7.5.2

Reclamation
    by drainage 4.6
    saline soils 3.5

Reclamation crops 4.6

Red soils 2.3.4

Reh 3.4

Relaxation method
    solution of Laplace's equation 6.6.4

Resalinization
    prevention 3.5

Residues
    organic 2.2.4

Resistance to penetration 2.3.3

Resting water
    boundaries 6.7.4

Retention of water see also Field capacity

Retention of water 2.2.1; 5.2
    curves 5.2.3
    mechanisms 5.2.2

River beds see also Braided rivers, Meandering rivers, Streams

River beds 1.3.1

River ridge 1.3.1

River terrace 1.3

Root crops
    influence of drainage 4.8.2

Root development
    in relation to soil colour 2.3.4
    reduced conditions 4.2.2
    soil-water relations 4.2.1
    supply of nutrients 2.7.2

Rootzone
    drainage 4.1
    internal drainage 2.4.1
    salt content 3.4
    soil colour 2.3.4
    total available moisture 2.4.2
    total readily available moisture 2.4.2
    total water potential 5.4.1

Runoff see Base flow, Groundwater flow, Surface runoff

Sabbagh 3.4

Saline soils 3.4
    reclamation 3.5

Saline-sodic soils 3.4
    reclamation 3.5

Salinity of soil
    effects on crops 3.3; 4.6
    influence of drainage 4.6
    plant growth 3.4

Salinization
    origin 3.1
    prevention 3.5

Salt tolerance of crops and plants 3.4; 4.6

Salts in soil
    occurrence 3.1
    osmotic potential 5.3.2
    types 3.2

Salt-water head 6.3.4

Salty soils
classification 3.4
origin 3.1

Salty groundwater 6.3.4

Sand
particle size limits 2.2.1
soil composition 2.2.2
textural classes 2.2.1

Sand bars
coastal plains 1.4
lacustrine plains 1.5

Sand fraction 2.2.2

Sandy loam
textural classes 2.2.1

Saturated soils
flow of water 5.4.1

Saturation 6.1

Saturation extract 3.4
salt tolerance of plants 4.6

Sea level 1.3

Secondary soil particles 2.3.2

Sediment
alluvial fans 1.3.2
braided rivers 1.3.1
coastal plains 1.4
lacustrine plains 1.5
meandering rivers 1.3.1
stratification 1.3
streams 1.3

Sedimentation in deltas 1.3.3

Seepage surface see also Free water surface

Seepage surface 6.7.5
inverted models 7.6.3
model 7.4
simulation in models 7.5.2

Semi-confined aquifer 1.2

Semi-confined water 1.2

Semi-pervious layers see Aquifers

SES see Sum exceedance value

SEW see Sum exceedance value

Sheet conductors
groundwater flow studies 7.3.2

Sheet floods 1.3.2

Sheets see Structural units

Sieves
U.S. standard 2.2.1

Silicates
clay fraction 2.2.2

Silt
particle size limits 2.2.1
soil composition 2.2.2
textural classes 2.2.1

Silt fraction 2.2.2

Silt loam
textural classes 2.2.1

Silty clay loam
textural classes 2.2.1

Simple structures 2.3.2

Slaking 2.3.2

Slowly moving water
boundaries 6.7.4

Soil
air 2.5
colour 2.3.4
compaction 2.3.3
composition 2.2.2
consistence 2.3.3
determination 2.1.1
development 2.1.3
engineering classification 2.3.3
factors A, B, C 2.1.6
factors of formation 2.1.3
fertility 2.7
moisture 2.4
permeability 2.4.1
plant growth 2.1.2
plasticity 2.3.3
porosity 2.3.1
salinity 3.4
structure 2.3.2
temperature 2.6
texture 2.2.1
water retention 5.2

Soil drainage 2.5.6

Soil aeration 2.5
influence of drainage 2.5.6; 4.3.2
plant requirements 2.5.4
pore space 4.2.1
soil processes 2.5.5

Soil air see also Oxygen

Soil air 2.5
composition 2.5.1
volume 2.5.2

Soil characteristics 2.2
definitions 2.1.6

Soil drainage 2.5.6

Soil fauna 2.2.4

Soil fertility 2.7
definition 2.7.1

Soil investigations 2.1.5

Soil moisture 2.4
adaptation of plants 4.2.2
by tillage 4.4.1
physics 5

Soil organic matter 2.2.4
influence of drainage 4.3.3
in relation to soil colour 2.3.4
in relation to structural stability 2.3.2

Soil particles
adsorption of water 5.2.2
aggregation 4.3.1
arrangement 2.3.2; 4.3.1
size 2.2.1

Soil porosity 2.3.1

Soil productivity 2.7.1

Soil profile 2.1.4

Soil solution
cation concentration 2.2.3
saline soils 3.2

Soil structure 2.3.2

Soil temperature 2.6
influence of drainage 2.6.2; 4.3.4
plant growth 2.6.1

Soil texture 2.2.1

Soil water content 5.1
matric suction 5.2.3

Soil water ratio 5.1

Soil water relations
plant development 4.2

Soil water retention 5.2.2
potentials 5.3
retention curves 5.2.3

Solonchak 3.4

Solonetz 3.4

Sorting sediments 1.3

Specific discharge 6.3.1; 7.2.1
measurement at the model 7.6.1

Specific conductivity see also Hydraulic conductivity

Specific conductivity 7.1
analogy to groundwater flow 7.2.3

Specific gravity 2.3.1

Specific heat of soil 2.6.2

Specific volume of soil 2.3.1

Specific weight of water 6.2.3

Spits 1.5

Squares
solution of groundwater flow 6.6.4

Stability of the soil structure 2.3.2

Standards for sieves 2.2.1

Steady flow see also Two-dimensional groundwater flow

Steady flow
Forchheimer equation 6.5.2
in an unconfined layer 6.5.5
Laplace equation 6.5; 7.1; 7.2
to parallel ditches 6.5.6
towards a well 6.5.7
two-dimensional 6.6; 7.2.1

Stickiness 2.3.3

Sticky soils 2.3.3

Storage of water
changes 6.5.3

Stratification
deposits 1.3

Stratified soils
infiltration into 5.4.2
water flow 6.4.2

Stream floods 1.3.2

Stream function 6.6.2

Stream losses 1.3.1

Stream-formed plains 1.3

Streamlines see also Equipotential lines

Streamlines
definition 6.6.2
electrical model 7.4
flow net 6.6.2
measurement at the model 7.6.3
method of squares 6.6.4
simulation in models 7.5.1

Streams see also River

Streams
    alluvial fans 1.3.2
    energy 1.2
    erosion 1.3
    profile 1.3
    sedimentation 1.3
    valleys 1.3.1

Structural profile 2.3.2

Structural stability 2.3.2

Structural units
    types 2.2.2

Structure see also Macro-structures, Simple structures

Structure of soil 2.3.2
    classification 2.3.2
    influence of drainage 4.3.1
    saline soils 3.5

Structureless soils see Simple structures

Subsoil
    effective soil depth 2.1.6
    internal drainage 2.1.6; 2.4.1

Subsoiling
    non-saline sodic soils 3.5

Sub-surface drainage 2.4.1
    soil aeration 2.5.6

Suction 5.2.2
    measurement 5.2.3

Sugarbeet see Arable crops

Sum exceedance value 4.8

Surface see also Seepage surface

Surface drainage 4.1
    maize 4.8.2

Surface intake rate 2.4.1

Surface runoff see also Peak discharge

Surface runoff
    glacial plains 1.6

Surface sealing 2.3.2

Surface soil
    textural grouping 2.2.1

Surface tension
    water retention 5.2.2

Swamp plants 4.2.2

Swamps
    coastal plains 1.4
    flood plains 1.3.1

Szik 3.4

TAM see Total available moisture

Tectonic valleys 1.9

Teledeltos paper 7.3.2

Temperature see also Soil temperature

Temperature
    in the soil 2.6; 4.3.4
    influence on soil aeration 2.5.3
    influence on water viscosity 6.3.3

Temporary waterlogged conditions 2.3.4

Tensiometer 5.2.3

Terraces
    coastal plains 1.4
    lacustrine plains 1.5
    rivers 1.3

Textural classes 2.2.1

Textural classification 2.2.1

Texture of soil 2.2.1

Thermal capacity of the soil 2.6.2

Tidal movement 1.3.3

Tidal rivers 1.4

Till see Glacial till

Tillage see also Weed control

Tillage
    influence of drainage 4.4.1

Tir 3.4

Topography see Land form

Top-set beds 1.3.3

Topsoil
    structural stability 2.3.2
    water intake rate 2.1.6; 2.4.1
    workability 2.1.6; 2.2.1

Total average moisture 2.4.2

Total potential of water 5.3.2

Total readily available moisture 2.4.2

Trace elements
absorption by plant roots 2.7.2

Trafficability of soil 2.5.6

TRAM see Total readily available moisture

Transient flow see Nonsteady flow

Transition layer 1.3.3

Transmission zone see Infiltration

Transport capacity
streams 1.3

Transport of water see also Capillary rise, Infiltration

Transport of water 5.4
equations 5.4.1

Tropical soils 2.3.4
clay fraction 2.2.2

Two-dimensional electricity flow 7.2.2

Two-dimensional groundwater flow 6.6; 7.2.1
analogy to electricity 7.2.2
basic equations 6.5
electrical models 7.3
exact methods of solution 6.6.3
method of squares 6.6.4
potential functions 6.6.2
relaxation method 6.6.4
stream function 6.6.2

Ultimate base level see Base level

Unconfined aquifers 1.2
groundwater flow in 6.5.5

Undulating landscape 1.6

Units
energy 5.3.1
physical 5.2.2
pressure 5.3.1
water potential 5.3.1

Unsaturated conductivity coefficient 5.4.2

Usar 3.4

Valleys 1.3.1

Value
soil colour 2.3.4

Varves 1.6

Velocity gradient of water 6.2.2

Velocity head see also Apparent velocity

Velocity head 6.2.5

Velocity potential 6.6.2

Vermiculite
cation exchange capacity 2.2.3
clay fraction 2.2.2

Virus diseases see Diseases

Viscosity of water 6.2.2
temperature influence 6.3.3

Void ratio of soil 2.3.1

Volcanic lakes 1.5

Volume of soil air 2.5.2

Volume weight of soil 2.3.1

Water see also Groundwater

Water
infiltration 5.4.2
potentials 5.3
presence in the soil 5.1
retention 5.2
transport 5.4

Water absorption see Intake rate

Water-bearing layers see Aquifers

Water conduction
function of pores 2.3.2

Water flow 5.3.1

Water infiltration 5.4.2

Water intake rate 2.4.1

Water potentials 5.3

Water-retaining properties of soil 2.4.2

Water retention see also Retention

Water retention 2.2.1; 5.2
function of pores 2.3.2

Watertable see Groundwater table

Water transmitting properties
in relation to pore size 2.5.3
soil 2.4.1

Water transport 5.4

Water vapor in the soil. 2.5.1

Waterlogging see also Temporary waterlogged conditions

Waterlogging 4.2.1
soil colour 2.3.4

Watertable aquifer 1.2

Weathertable minerals 2.2.2

Weed control see also Tillage

Weed control
influence of drainage 4.4.2

Well
discharge 6.5.7
fully penetrating 6.5.7
partially penetrating 6.5
steady flow to 6.5.2; 6.5.7

Wet soils 4.3.4
plasticity 2.3.3

Wheatstone bridge
measurement of discharge 7.6.1

White alkali 3.4

Wilting point 2.4.2; 5.2.1

Workability of the soil 2.2.1
soil-drainage 2.5.6

Xerophytes 4.2.2

Yellow soils 2.3.4

Yields
arable crops 4.8.2
drainage influence 4.5
fruit trees 4.8.3
grassland 4.8.1
groundwater level 4.8
soil salinity 3.4
subsoiling 3.5