HYDROGEOLOGY
ON AGRICULTURAL DEVELOPMENT

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OVERSEAS TECHNICAL COOPERATION AGENCY

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CHAPTER I. INTRODUCTION

§ 1. Where we can find water around and in the interior of the globe?

Water is found generally around and in the interior of the globe in several forms described in the following:

(1) Atmospheric water

Around our globe we see a layer of "atmosphere" of about 10km thickness and it contains in various forms water particles which were formed by evaporation from the ocean and the surface of the earth.

Cloud, fog and mist are gases which are composed of water-vapor. A part of them falls on the globe as rain in a form of liquid, or as snow, hail or sleet in a form of solid, as seen in a season or in countries of cold climate. This atmospheric water in various forms is called precipitation or rainfall when if falls on the globe.

(2) Surface water

The ocean represents "hydrosphere" and gathers water of precipitation which fell on the ocean itself and on the land surface. The surface land water is fresh water and comprises rivers, canals, glaciers, lakes, ponds, swamps, reservoirs and depressions on the land surface. It is distinguished in scale and in salinity from sea water in the ocean.

The river water carries dissolved salt brought from land by erosion and disintegration into the ocean and gives to the sea water-salinity. A part of the surface land water as well as the sea water is lost by evaporation and is carried to the atmosphere in form of water vapors.

(3) Subsurface or underground water

A part of precipitation on the land surface seeps under the ground and is kept there temporarily.

Forming the uppermost layer of the lithosphere, there is the soil zone and water due to infiltration from ground surface is contained in pore spaces among soil particles in a suspended state and moves downward vertically. This water gives moisture to soil and is called soil water or infiltrated water. A part of moisture is returned to the atmosphere directly by evaporation and the other part is absorbed by plants to return again to the atmosphere by transpiration.

An excess of soil water goes downward to be stored beneath, to form an underground reservoir, dammed up on an underlying impermeable strata or compact rock mass. This is the shallow groundwater and water moves mainly in a horizontal direction, which is called percolation. Its movement is controlled by the slope of surface of water body called water table, and obeys the laws of hydrodynamics. This shallow groundwater is called "free groundwater". Posepny (1894-1895) called it "Vadosum" and Weithofer (1933) "Fliesswasser (moving water)".

There is another type of groundwater or deep groundwater, which Posepny (1894-1895) called "Profondum" and Weithofer (1933) "Stagnante Wasser (stagnant water)". It situates underneath the shallow groundwater and the water moves very slowly, only by diffusion, and is not controlled by law of hydrodynamics. This groundwater can find its outlet only as spring. This deep groundwater is called "confined groundwater".

The stagnant water is considered to spread till the depth where rocks have no pores to contain water or porosity is zero. This depth is supposed as 10km by Van Hise (1904), 18km by Adams (1924), 30km by King (1897-1898) and 50km by Sieberg. We can estimate theoretically that the water exists till the depth of more than 30km, as deduced by Smith (1953), who calculated it from geothermal rate of increase of 1°C for every 33m and relation of pressure, volume and temperature of water with depth.

Soil water and groundwater are summarized as subsurface water or underground water.

Subsurface and surface land water as well as sea water are, in general, derived from atmospheric water. They are called meteoric water by their origin.

§ 2. Special occurrence of groundwater

There is water originated from magma, which is called magmatic water. Magma is composed of melted rock-forming mineral combined with water vapor and other gases. Magmatic water is driven out from magma when it intrudes into rocks and consolidates to crystalize minerals.

Volcanic water is included together with magmatic water in pore spaces in surface lava flow and amygdaloidal structure is resulted. This water is expelled only by destruction of pore spaces as seen in pumice. Volcanic water is produced by combination of oxygen and hydrogen in molten magma and it does not contain argon, which is characteristic of atmospheric origin.

Cosmic water is water which comes in from universal space outside of the globe.

Magmatic, volcanic and cosmic water is called juvenile water, which means new water.

When sedimentary deposits are consolidated beneath the ocean, salty water is originally included in pore spaces among their composing materials. Such water is called connate water. After upheaval of sedimentary rocks over the sea, salty water is slowly driven out by replacement with fresh water which infiltrates from the ground surface. But when the water is stagnant, confined in deep layer or prevented from some barriers geologically formed, and preserves original quality of connate water, it is sometimes called fossil water. We see such examples in some area of sedimentary formations and in oil fields.

When sediment is compressed by overlain strata, its porosity becomes smaller as seen in case of freshly deposited material decreasing from as high as 50% 59 3 ~ 4% in shales and water contained in interstices being squeezed out. Such water is called water of compaction. Farther compression in metamorphism causes dehydration and water, which was fixed in minerlas, is forced out. This water is called metamorphic water. Water of compaction and metamorphic water are so called rejuvenated water.

Special occurrence of water above mentioned makes a part of water supply in the lithosphere. While weathering, on the contrary, plays a roll of depriving of water to combine with minerals of rocks and fix it to form new minerals. The quantity of water thus consumed reverses again to water supply through erosion, disintegration and other geological processes.

§3. Quantity of water around and on the globe

The whole quantity of water on the globe is estimated by Halbfasse (1913) as 1,304,068,550 x $10^9 m^3$, which corresponds to a depth of 2,558m when spread uniformly on the whole surface of the earth. This estimation is very near to the value 2,684m which was estimated by Goldschmidt (1933).

The quantity of water contained in the atmosphere is estimated as 12,500 x $10^9 m^3$ or the depth of 25mm when spread on the whole surface of the globe by Humphreys (1931) and Merril (1942), 30mm by Goldschmidt (1933) and Fox (1952), and 24mm by Reichel (1949).

The sum of the volume of water in the ocean and that of surface land water is calculated by Halbfass (1913) as 1,300,000,000 x $10^9 m^3$, or it corresponds to the depth of 2,550m. It shows that the surface water occupies 99.9% of the total volume of water of the globe.

About the quantity of groundwater, the estimation varies according to authors. They deduced it from some assumption on the lower limit of depth where pores are included in the rock, that is the maximum depth of existence of groundwater, and the mean porosity for the whole depth.

Assuming the lower limit of depth with pores being 18,500km and the porosity of 12.5% being uniformly distributed for the whole depth, Delesse (1861-1862) obtained as the quantity of groundwater 1,175,089,000 x $10^9 m^3$ or 2,312m as the depth of ground water, when considered being spread uniformly on the whole surface area of the globe.

Slichter (1902) estimated the total quantity of groundwater being 430,000,000 x $10^9 m^3$, or 900 ~ 1,000m depth, when spread on the whole surface area of the globe, taking that the lower limit of depth for groundwater is at 9,656m below the bottom of the oceans, and assuming that the mean porosity for the whole depth is 10%.

Both of them assumed that the porosity is uniform for the whole depth and took its value too large, and accordingly their estimations seem too much exagerated.

While Van Hise (1904) considered that the porosity diminishes gradually as the depth from 6.89% near the surface to zero at 10km depth where exists the lower limit of groundwater. He obtained the mean porosity of 0.69%, which represents 69m depth of groundwater when spread on the surface of the continent. But he thinks the depth being still larger as two or three times of this value.

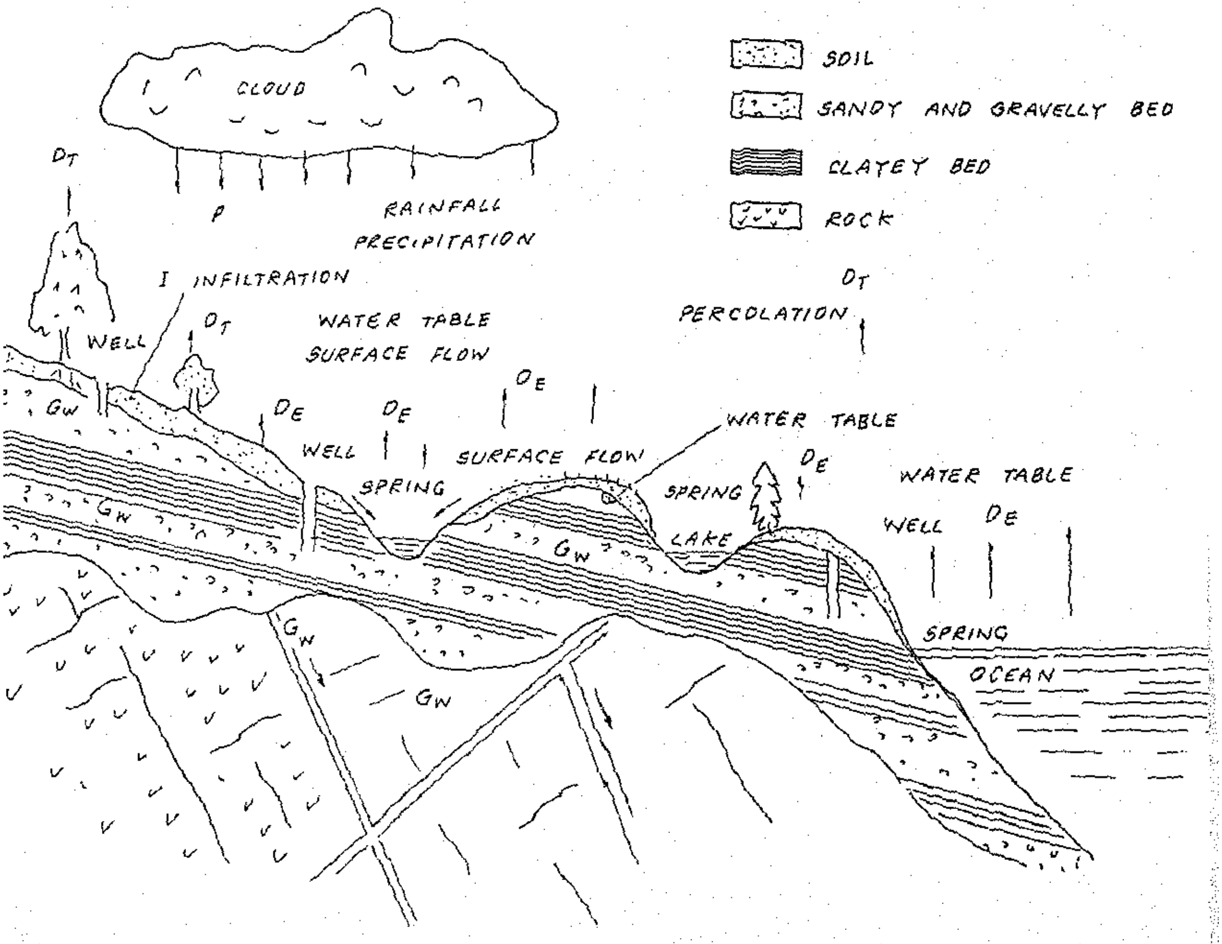

Fig. 1 Circulation of water

§ 4. Hydrologic cycle and definition of "Hydrogeology"

Water circulates in various forms around and near the surface of the globe.

Some part of sea water as well as of surface land water, which is kept temporarily in lakes, ponds, swamps, rivers, canals and depressions, is lost by evaporation to form water vapor in the atmosphere. A part of snow deposit and of glaciers is lost likewise by evaporation after melting, and some other part directly by sublimation.

A part of precipitation on the land surface is carried away by rivers and glaciers to the ocean and some part infiltrates under the ground. The remaining part is partly lost by evaporation and sublimation before or after arriving on the surface of the earth, and partly lost by evaporation and transpiration after infiltration. The groundwater moves toward a valley, a lake or sea and joins surface water to be carried to the ocean.

It joins directly to sea water in coastal regions.

All these processes are called hydrologic phenomena and such circulation of water is called hydrologic cycle. "Hydrology" is a science to study hydrologic phenomena. Groundwater and its movement is always observed under the ground in and through soils and rocks. The science of groundwater treats hydrologic relations under the ground and occurrence and movements of groundwater. The knowledge of hydrodynamics is necessary for the explication of its movement. Thus the terminology "Hydrogeology" (in English) or "Hydrogeologie" (in French) is now generally used for this science. It means "Groundwater geology". It is used since the time when the science of groundwater was approved as a branch of geology at the general meeting of the International Geological Congress held in Algiers in 1952. It is translated in every country as "Hydrogeologie" in Germany, ГИДРОгеогия in U.S.S.R., هيدرولوجيا مائية in Arabic countries, 水文地質学 (Suimon-chishitsugaku) or 水地質学 (Mizu-chishitsugaku) in Japan. But, on the other hand, "Groundwater hydrology" is used in U.S.A. and "Hydrologie souterraine" in France. It has the same connotation as "Geohydrology". In "Hydrogeology", geology is more emphasized than hydrology.

§5. Method of study on "Hydrogeology"

Hydrological phenomena are influenced by many factors – meteorological, climatological, topographical and biological – which are different in time as well as in space.

Kinds of soils and rocks and their distribution as well as geologic structure are different in area and thickness. Accordingly, characteristics of soils and rocks to permit groundwater to occur and to make it move vary geographically. Hydrogeological properties must be studied for every type of sediments and rocks.

Volume of groundwater and its yield, that is the quantity which is produced during some period, at a certain place, is what is resulted from hydrological relations combined with local hydrogeological properties of rocks.

Thus the groundwater problems must be solved by researches of various factors concerned. It is needful to collect data of observations on precipitation, evaporation, transpiration, rate of flow of river water, amount of water reserved in lakes, ponds and swamps and also that of spring water which flows out of groundwater reservoir. It is also important to know physical and chemical qualities of soils and rocks and also to investigate in detail on the geological characteristics at the region concerned.

Mutual relations of these factors are so much complicated that it belongs to a very difficult thing to explain briefly or to express by a simple mathematical formula the summarized results affected by these factors. But we must endeavor to find general rules by arrangement of these data. To prepare "hydrogeological maps" in the world by collaboration of hydrogeologiests in the world is one of the most important problems in hydrogeology requested by the International Association of Hydrogeologists (I.A.H.).

"Hydrogeology" is a new science which has developed recently and it may furnish plenty of knowledge in branches of engineering which treat soils, rocks and water. Knowledge of hydrogeology must be applied not only on engineering concerning water supply, reclamation of land, irrigation and drainage, selection of sites and construction of tunnels, dams and reservoirs, but also on technics of protection against soil abrasion, and slides, subsidence of land and other catastrophes. Answering to an urgent need of water for human life, as for drinking, agricultural and industrial uses, the development of "Hydrogeology" is very important and it is desirable to apply this science for finding more water resources.

In occurrence of petroleum, three fluids of gas, oil and water are observed. They are governed by the same hydrodynamic law, but the groundwater in petroleum reservoir has a particular condition and it is treated in "Petroleum geology".

§6. Historical review and development of the science of groundwater

"There may be water under the ground". This idea was held among people since before the age of Christ, and various methods were tried to find water from under the ground.

In the first age, it was convinced that water of the ocean comes through some passages under the ground in the continent, and salty water has been converted afterwards into fresh water to come out as springs. And next was made a hypothesis that the water is produced by condensation of water vapor in air on walls in some caves, where air is cooler than outside, in mountainous regions. The former hypothesis is not reasonable, but some part is true concerning the latter one. Because the dissolved salts in the sea water become concentrated more and more by evaporation and by addition of various kinds of salts, dissolved in water which passes through soils and rocks, but salty water never converts into fresh water.

"Groundwater is originated from seepage of rain water or river water through pores in soils and fissures in rocks", it is an idea that is believed contemporaneously as the *infiltration theory*. Marcus Vitruvius was the first scholar who published this theory in the age of Christ.

Any new idea was not born till the end of the 17th century. The science of groundwater did not show any advancement nor development. It was in this period that the first survey on hydrology in the Seine basin was made, based on physical measurements, in France. Pierre Perrault (1608 ~ 1680) measured the pluviosity (amount of precipitation), Edmé Mariotté (1620 ~ 1684) made measurement on the discharge of the Seine and Edmund Halley (1656 ~ 1742) investigated on evaporation from water surface. The conclusion obtained by this survey proved that a good balance is kept between figures on evaporation, pluviosity and river discharge, including discharge of spring water.

In the last half period of the 18th century, the properties of rocks concerning water absorption was made clear.

In the beginning and middle period of the 19th century, many investigations were made in France on conditions of artesian water and its movement of flow under the ground. It was in 1856 when Henri Darcy (1803 ~ 1858) published his famous theory on the movement of groundwater flow through sands. Later in the 19 century, "hydraulics" of groundwater developed in Europe. J. Boussinesq, G.A. Daubrée, J. Dupuit, P. Forchheimer and A. Thiem are famous authors in this period. It was since near this period that the study of groundwater commenced in U.S.A. These early studies were mainly on the occurrence and the movement of the artesian water*, and the science of groundwater has advanced much, since then, with increase and accumulation of many fundamental data concerning not only the movement, but also the physical and chemical properties of the groundwater.

In the 20th Century, researches on groundwater have become very active and many books and papers have been published. We can enumerate the names of R. Dachler, E. Imbeaux, K. Keilhack, W. Koehne, J. Kozeny, E. Prinz and G. Thiem in Europe. A. Hazen, C. King, C.S. Slichter and O.E. Meinzer contributed to the development of "Groundwater hydrology" in U.S.A.

After the hydrogeological science was admitted as a branch of geology at the International Geological Congress in Algier in 1952, as already mentioned, the "International Association of Hydrogeologists" (Association internationale des hydrogeologues; A.I.H.) was established in June 1957 in Paris. At this first meeting P. Fourmarier, honorable Professor of the University of Liège (Belgium), was elected as the first President of the Association. Since then the international meeting of hydrogeologists in the world is held once every year in different countries, and the Association has contributed much in development of this science through many numbers of communications on hydrogeological problems.

There is another organization which plays an important role through its activity for the development of hydrogeological science. It is the "International Association of Scientific Hydrology (I.A.S.H.)" (or "Association internationale d'hydrologie scientifique (A.I.H.S.)"), which is a division of the "International Union of Geodesy and Geophysics", and the international congress is held every three years at different places. Communications presented at the congress are published with aid by the Unesco.

There is in the Unesco "the project of researches concerning the Arid Zone" and "the long-term programme of research in Scientific Hydrology". Hydrological and hydrogeological problems are treated respectively in these programmes.

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* Artesian water is the groundwater which comes out of the deep underground with pressure, as seen in Egypt and in Australia.

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CHAPTER II. WATER BALANCE

§ 1. Hydrologic circulation on the land

The amount of the atmospheric water including rain and snow-fall is called precipitation, a part of which is lost in the atmosphere by evaporation before it arrives on the land and the sea. We shall limit here the precipitation, P, being only the water which has actually fallen on the land. After the moment when it has fallen on the ground, this water is distributed in three different ways (Refer to the table on p.13).

(1) Water losses by evapo-sublimation (D-1) from surface – water returning to the atmosphere by sublimation (D_S) and evaporation ($D_{E\cdot 1}$).

(2) Direct surface water (R_W) – kept stagnant in ponds, lakes, reservoirs, or flows in streams of rivers, canals,

(3) Water which seeps under the ground by infiltration (I).

1. Precipitation on the surface water

Of the precipitation on the surface water, some part is lost by evaporation from the water surface, constituting a part of the evaporation losses ($D_{E\cdot 1}$). In a season or in regions of cold climate it disappears directly by sublimation from the surface of ice cover, snow deposit, névé or glaciers (D_S). While the main part of it flows as a part of river water ($R_{W\cdot 1}$) and it is sometimes reserved, for a while, in lakes or reservoirs along their courses. When it is kept in ice form, it joins the river water after melting, some elapse of time later. Some part of the surface water seeps under the ground through the bank and the bottom of rivers or lakes, forming a part of groundwater ($G_{W\cdot 1}$).

2. Precipitation on the land

The rain- or snow-fall is intercepted, or prevented from flowing away to the ground, kept on trees, grasses and other vegetations. This phenomena is called interception and some part of this precipitation returns again to the atmosphere, inform of water vapor or gas, by evaporation ($D_{E\cdot 1}$) or by sublimation (D_S). Dew is the water drops formed by condensation of water vapor on the surface of soils or rocks and it freezes to frost in cold weather. On branches of trees, the silver-thaw is formed sometimes in cold weather. They also disappear by evaporation ($D_{E\cdot 1}$) or by sublimation (D_S).

During and even after the rain-fall, or after melting of snow deposit, a film of water covers the ground surface (sheet flow). At some places where the land form is depressed, detention is seen and a small quantity of water is reserved there. Such water loses some part of it by evaporation ($D_{E\cdot 1}$). The ice cover or the snow deposit on the ground surface disappears by sublimation (D_S) or by evaporation ($D_{E\cdot 1}$) after melting. These are the water losses by evapo-sublimation from the surface of the land (D-1).

The surface runoff, or the sheet flow on the surface of the ground, goes down the slope as a direct flow to join the rill water in the valley ($R_{W\cdot 2}$).

And the remaining portion of the precipitation is absorbed into soils or fissures of rocks; this is the infiltrated water (I). The water absorbed by infiltration from the ground surface is kept in spaces between soil particles as the soil water and the subsoil water underneath. The soil water is consumed by roots of vegetations and carried in the atmosphere by transpiration (D_T). The soil water and the subsoil water is lost partly by evaporation ($D_{E\cdot 2}$). Some part of it is kept temporarily as retained water (G_r) which is lost only by evaporation when the soil dries up ($D_{E\cdot 2}$). The remaining part of the infiltrated water goes downward to contribute to the groundwater ($G_{W\cdot 2}$).

Thus we can express the relations of P in the first step by

$$P = D\text{-}1 + I + R_W \qquad \dots\dots\dots\dots\dots\dots \qquad (1)$$

where

$$D\text{-}1 = D_S + D_{E\cdot 1}$$

In the second step after infiltration of some part of precipitation, the infiltrated water I is divided into two types of water, I-1 and I-2. The former is the water which is lost by evapo-traspiration ($D\text{-}2 = D_{E\cdot 2} + D_T$) and the later is the water which is reserved under the ground as retained water (G_r) and groundwater (G_W).

The whole amount of water losses D may be expressed:

$$\begin{aligned} D &= D\text{-}1 + D\text{-}2 \\ &= (D_S + D_{E\cdot 1}) + (D_{E\cdot 2} + D_T)\,. \\ &= D_S + D_E + D_T \qquad \dots\dots\dots\dots\dots\dots \qquad (2) \end{aligned}$$

When considered for a long duration, the retained water (G_r) is neglected and the whole amount of water which is kept on and under the ground is carried as the river water Q to pour into the ocean.

And

$$\begin{aligned} Q &= G_{W\cdot 1} + G_{W\cdot 2} + R_{W\cdot 1} + R_{W\cdot 2} \\ &= G_W + R_W \qquad \dots\dots\dots\dots\dots\dots \qquad (3) \end{aligned}$$

Thus (1) becomes as

$$\begin{aligned} P &= D^{\text{-}1} + I^{\text{-}1} + I^{\text{-}2} + R_W \\ &= D^{\text{-}1} + D^{\text{-}2} + G_W + R_W \\ &= D + Q \qquad \dots\dots\dots\dots\dots\dots \qquad (4) \end{aligned}$$

This equation is for the case when the groundwater basin accords with the topographical basin of the river concerned. That is, no amount of groundwater or river water is lost from nor gained to the river basin concerned.

While there happens sometimes, however, that some quantity, q, of river water is taken and derived artificially for irrigation, industrial or drinking purposes to the adjacent basin. The hydrogeological condition due to the characteristics of geological structure or properties of rocks often permits to make a part of groundwater escape likewise. Also may be expected their reversal cases. There may be the case, for example that the water balance is broken due to the additional quantity of water by pumping up of deep groundwater which is originated from adjacent basins or by importing surface water from any other basins. In such case, then, the above described equation (3) must be modified as

$$Q = R_W + G_W \pm q \quad \dots\dots\dots\dots \quad (5)$$

the sign of q being taken as $-$ or $+$ according to the former (deprived of water) or the latter (water being added) case respectively.

The equation of the water balance (4) may thus be rewritten,

$$P = D + Q \pm q \quad \dots\dots\dots\dots \quad (6)$$

The value, q, due to the artificial operation may vary every year, while that caused by the hydrogeological conditions may not do so. A care must be taken in calculation not to neglect its value, even when it may be trifle amount.

These factors are denoted in volume (m^3, ℓ or cub.ft.) or in depth (mm or inch). They are influenced by geographical, topographical, climatological, meteorological, hydrogeological and biological elements of the basin concerned.

The above description concerns always to one year cycle, and it is reasonable to draw a boundary at the end of the month of the least yield. The one year cycle taken in such manner is called the hydrologic year, (or water year) which is distinguished from the so called calendar year.

Some quantity of D and Q, derived from P for a certain year, may be retained at the end of the year and transferred to the beginning of the next year. The input at the beginning and the output at the end of the year introduce errors in estimation of water balance, but these errors may be cancelled and become negligible when one takes the mean yearly value during a period taken enough longer, as is expected the better results in the mean yearly precipitation, P, for the period taken as long as possible.

§ 2. Mutual relations between hydrological factors

1. Relation of D to P and Q

We can measure P and Q, and D is obtained by calculation:

$$D = P - Q$$

<table>
<tr><td colspan="3" rowspan="6">FIRST STEP</td><td colspan="8">Precipitation P</td><td rowspan="6">To (–) or from (+) the adjacent basin</td></tr>
<tr><td colspan="2">Deficit ⇑
D-1</td><td colspan="4">⇑ Infiltration
$I^{-1} = D^{-2}$ $I^{-2} = G_W$ ⇓</td><td colspan="2">Surface flow
R_W</td></tr>
<tr><td colspan="2">On the ground</td><td colspan="4">Under the ground</td><td colspan="2">On the ground</td></tr>
<tr><td colspan="3">Evapo-sublimation</td><td>Transpiration</td><td colspan="2">Groundwater</td><td colspan="2">Surface flow</td></tr>
<tr><td>D_S</td><td colspan="2">D_E
D_E.1 D_E.2</td><td>D_T</td><td colspan="2">G_W</td><td colspan="2">R_W</td></tr>
<tr><td>–°C
Ice</td><td colspan="3">+°C
Water-vapor</td><td>–°C
Melting water</td><td>+°C
Water</td><td>–°C
Melting water</td><td>+°C
Water</td></tr>
<tr><td rowspan="2">Surface water</td><td rowspan="2">Ponds Lakes Reservoirs Rivers Canals</td><td>On the ground</td><td>Ice cover Snow-deposit Névé Glacier D_S</td><td>Water surface D_E.1</td><td colspan="2"></td><td colspan="2"></td><td colspan="2">Surface flow R_W.1</td><td rowspan="2"></td></tr>
<tr><td>Under the ground</td><td></td><td></td><td colspan="2"></td><td colspan="2">Ground-water G_W.1</td><td colspan="2"></td></tr>
<tr><td rowspan="5">Land Surface</td><td>Vegetation</td><td>Inter-ception</td><td>Snow Ice Silver thaw D_S</td><td>Water-drops D_E.1</td><td colspan="2"></td><td colspan="2"></td><td colspan="2">Surface flow R_W.2</td><td rowspan="5">Precipitation P</td></tr>
<tr><td rowspan="4">Soils and Rocks</td><td>On the ground</td><td>Frost D_S</td><td>Dew Detention D_E.1</td><td colspan="2"></td><td colspan="2"></td><td colspan="2"></td></tr>
<tr><td rowspan="3">Under the ground</td><td rowspan="3"></td><td rowspan="3"></td><td>Soil water D_E.2</td><td>Soil water D_T</td><td colspan="2" rowspan="2">Retained water Gr</td><td colspan="2" rowspan="3"></td></tr>
<tr><td>Sub-soil water D_E.2</td><td></td></tr>
<tr><td colspan="2"></td><td colspan="2">Ground-water G_W.2</td></tr>
<tr><td colspan="3">LAST STEP</td><td></td><td>←</td><td colspan="2">⇑ Percolation ←</td><td colspan="2">G_W ←</td><td colspan="2">R_W –
Spring +</td><td>q</td></tr>
</table>

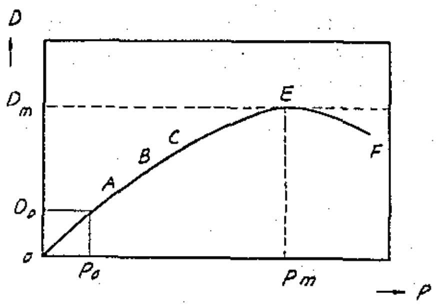
When these values are indicated by average height in the basin, they are called respectively pluviometric index for P, deficit of run-off for D and run-off for Q.

We have several data on these relations for different places. Concerning the relations between P mm and D mm, we obtain a linear relation as in the following;

For Sweden:	$D = 392 + 0.05\ P$ (by Wallen),
For Alps in Switzerland:	$D = 980 + 0.218\ P$ (by Lugeon)
For Carpatian in Poland:	$D = 594 + 0.14\ P$ (by Pomianovsky)
For Tropical country:	$D = 425 - 0.305\ P$ (by Keller)
For Western Europe:	$D = 405 + 0.058\ P$ (by Parde)

These relations are expressed by formulas somewhat different according to regions. Summarizing them, we can draw a general curve shown in Fig. 2.

Explication:



1°) Regions with precipitation relatively small, compared with temperature. All the precipitation is lost by evapo-transpiration; or

$D = P$ (Part OA of the curve)

2°) Regions with precipitation relatively low and the relation is expressed by a nearly straight line; or

$D = a + bP$
(Part AB or BC of the curve)

3°) Regions with moderate precipitation and the relation is expressed by an equation of the 2nd order; or

$D = a + bP + cP^2$ (Part AC of the curve, with a small variation of curvature)

4°) Regions with moderate precipitation, not so variable, and deficit is almost constant; or

$D = D_m$ (Part E of the curve)

5°) Regions with precipitation relatively high and the deficit decreases as it increases; or

$D = a - bP$ (Part EF of the curve)

2. Relation of D to T

In a limited area, there is a special relation between deficit D mm and temperature T°C.

Coutagne (1943) found that D shows larger value where T is higher, thus

$$D = 210 + 30T \quad \text{for Europe}$$

and

$$D = 180 + 40T \quad \text{for the United States of America.}$$

According to Turc (1954), D is a function of P and T, and is represented as

$$D = \frac{P}{\sqrt{0.9 + P^2/L^2}}$$

where $L = 300 + 25T + 0.05T^3$, and T means the average yearly temperature in °C in the basin concerned.

Data on Deficit of Run-off (L. Turc, 1954-6)

	T°	P	D	Q
Java	19.5	3792.5	1129	2663.5
Madagascar	17.8	1772	852	920
Viet Nam	18.5	1760	880	880
A.O.F. and A.E.F.	24.6	1581	1139.5	441.5
Ethiopia	21.7	1013	866	157
Algeria and Tunisia (Tributaries of the sea)	15.5	701	458.5	142.5
Italy:				
Basin of Po	9.3	1266.5	589.5	677
Peninsular	11.8	1050.5	538	512.5
Sicily	14.3	721	512	209
Sardinia	14.2	860	549	311
France:				
Seine and tributaries	10.3	682.4	481.9	200.5
Between Seine and Loire	10.4	806	464.5	341.5
Tributaries of Loire	10.4	776	478	298
Tributaries of Garonne	8.5	1176	471	705
South of Garonne	9.7	827	460.5	366.5
Rhone and tributaries	4	1510	443	1067
Lower Provence	13	904	599	305
Middle France	8.3	1028	470	558
Britain Islands	8.5	1171	438.5	732.5
South of West Germany to Switzerland	6.9	1257	462	795
North and center of Germany, Poland, Czechoslovakia (west)	7.4	824	414	410

	T°	P	D	Q
Sweden and Denmark	5.8	712	350	362
Finland	1.9	568	264	304
European U.S.S.R. (excluding tributaries of Black Sea)	3.6	540	321	219
Tributaries of Black Sea	6.5	517	414	103
U.S.A.				
Tributaries of Atlantic Ocean and Gulf of Mexico	12.3	963.5	668.5	295
Tributaries of Pacific Ocean (excluding Colorado)	10.7	670	446	224
Colorado basin	7.2	387.5	283.5	104

Question 1

Using these data, plot P and D in relation to T° and draw curves which represent the relation of P to D concerning T°.

Values of Q and D in function of T° and P

(Annual relation, Wundt, W., 1953)

	P cm											
°C	20	40	60	80	100	120	140	160	180	200	220	240
	Q cm											
0°	6	19	36	53	73	93	112	132	152	172	192	212
+5°	33	12	25	40	58	78	98	118	138	158	178	198
+10°	2	6	16	28	43	62	81	101	121	141	161	181
+15°	2	2	9	17	29	45	62	82	102	122	142	162
+20°	1	1	3	8	16	28	42	60	80	100	120	140
+25°	1	1	1	2	4	12	22	35	55	75	95	115
	D cm											
0°	14	21	24	27	27	27	28	28	28	28	28	28
+5°	17	28	35	40	42	42	42	42	42	42	42	42
+10°	18	34	44	52	57	58	59	59	59	59	59	59
+15°	18	38	51	63	71	75	78	78	78	78	78	78
+20°	19	39	57	72	84	92	98	100	100	100	100	100
+25°	19	39	59	78	96	108	118	125	125	125	125	125

Question 2

Using these relations summarized by Wundt, in the same manner as in Question-1, represent in curves the relations of D to P concerning T°.

3. Moduls of discharge, q, and coefficient of run-off, or ratio of Q to P

The quantity of river water, Q, is expressed in volume (ton or cm^3/year) or in height (run-off, cm or mm/year). Sometimes the unit per second is used. That is, discharge (m^3/sec.) and moduls of discharge (ℓ/sec. km^2 or ft^3/sec. $mile^2$ = 10.9ℓ/sec. km^2). In consideration of one bissextile year for every 4 years, we can calculate run-off (in cm/year) multiplying 3.16 by moduls of discharge (ℓ/sec. km^2).

For example, the discharge of the Seine in Paris was measured as 287 m^3/sec. in 1955. The catchment area is 44,320 km^2. The moduls of discharge is $\frac{287 \times 1000}{44{,}320} = 6.47$ ℓ/sec. km^2 and the run-off is $6.47 \times 3.16 = 20.4$ cm/year.

Prof. Parde (1955) found that the moduls of discharge (ℓ/sec. km^2) shows higher value for smaller area, thus

30 ℓ/sec. km^2	for 20,000 km^2
35	10,000
45	1,000 ~ 2,000
60	250 ~ 500
70	50 ~ 100
80	10 ~ 50

The ratio of run-off Q to precipitation P is influenced by the geographical characteristics of the river basin.
This ratio changes considerably between 0 and 100%. In the desert and its vicinity, this ratio is very small as indicated in 3.5% for the Nile, 11% for the Colorado and less than 10% for the Missouri. It has a tendency to increase as the precipitation is abundant. For example, 23% for the Mississippi, 29% for the Seine, 41% for the Garonne, 56% for the Rhône and 65 ~ 68% for the P_0. In mountaneous regions with much rain-fall, it shows higher values as shown in 80 or 85% in rivers in the Pyrenees and the Alps.

Question 3

Find the coefficient of run-off from data given on page 16.

4. Relations among factors

Concerning the water balance of groundwater in a certain basin, it is especially important to make clear the proportion of the groundwater to the precipitation, $\frac{I\text{-}2}{P} = \frac{G_W}{P}$, or to the river discharge $\frac{G_W}{Q}$ (both in %), and accordingly the groundwater run-off (mm/year or ton/km^2, year). It is different for every basin due to various conditions in geology. The groundwater capacity, (G_W) (ton/km^2), which is the total quantity of groundwater contained in the whole thickness of aquifers, is desirable to be estimated. The aquifer means the water-bearing bed, which will be explained later.

§ 3. Water balance during a certain period

1. Natural state

We shall analyze here in detail the water balance for a certain period in a given basin. Precipitation P is divided into evapotranspiration D_1, infiltration I and direct surface flow or surface run-off R_W.

$$
\begin{aligned}
P &= D_1 + I + R_W \\
&= D_1 + (D_2 \pm \Delta G_r + G_W) + R_W \\
&= D_t \pm \Delta G_r + G_W + R_W
\end{aligned}
$$

The water moisture in the basin is then represented by

$$P - D_t = \pm \Delta G_r + G_W + R_W$$

and

$$I - D_2 = \pm \Delta G_r + G_W = \alpha_1 P,$$

α_1 being average coefficient of infiltration for the whole basin.

$\pm \Delta G_r$ is the difference of soil moisture or retained water in the zone of aeration at the end and the beginning of the period considered. Some quantity of groundwater comes into the basin from the adjacent basin of upper reach ($+g_1$) and some goes out to the next basin of lower reach ($-g_2$), their difference being represented as $\mp \Delta g$. A part of groundwater G_W percolates forward the river course and seeps out (effluent) as base flow ($-S_2$) to join the river water at a lower stage or, a part of the river water at a higher stage flows out (influent) in a reverse direction through walls or beds into the alluvial aquifer ($+S_1$), their difference being $\mp \Delta S$. The difference in groundwater storage during the period is

$$G_{nat} = G_W \mp \Delta g \mp \Delta S$$

Separating into two kinds of groundwater, as described later, free and confined, we obtain

$$G'_{nat} = G_W' \mp \Delta fc \mp \Delta g' \mp \Delta S'$$

$$G''_{nat} = G_W'' \pm \Delta fc \mp \Delta g'' \mp \Delta S''$$

where ′ represents free groundwater, ″ confined groundwater and Δfc the amount of water which interchanges between them. The difference in storage, G′, for free groundwater is represented by the difference in heights of the level of the water table at the end and the beginning of the period and that, G″, for confined groundwater is by the difference in statical pressure during the same period.

Representing in parentheses [ ], the amount of water which flows out from the basin, the outflow of groundwater at the lower boundary of the basin is

$$\pm [\Delta g] = \pm \Delta g' \pm \Delta g''$$

The discharged quantity of river water $[Q_{n\text{-}nat}]$ in the n-th separated basin along a river course is given by the difference of discharge at outlets for the n-th and (n-1)-th basins, or $[Q'_n] = Q_n - Q_{n-1}$. R_W and S_2 must also be calculated in the same manner.

So generally represented

$$\begin{aligned}[Q'_{nat}] &= R_W \pm \Delta S \\ &= R_W + S_2 - S_1 \\ &= (1 - \alpha_2)(R_W + S_2)\end{aligned}$$

We shall call effluent ratio for S_2/P, influent ratio for S_1/P and coefficient of influent flow for α_2.

Here is an example of water balance analysis for the Kumano-gawa basin which is shown diagrammatically in Fig. 3.
In this diagram the configuration of D-curve and the position of the straight line $\alpha_1 P$ are not correct, but they may be fixed if either of them would be confirmed by some measurement. The configuration of S_2-curve was drawn by the effluent groundwater flow analyses of hydrographs.

2. Artificially modified case

The water balance in natural state is modified artificially by utilization of water in the basin and by importation or exportation of water from or to other basins.

When the storage of lakes, ponds or reservoirs is controlled by dams, the river discharge, Q_{nat} or $R_W \pm \Delta S$, is subtracted by V when stored and is added by V when discharged. When some quantity of water of rivers, lakes or groundwater is expelled for usage in the basin, one must add $\pm\Delta G'_{(u)}$ to G'_{nat}, $\pm\Delta G''_{(u)}$ to G''_{nat} for the total utilized amount of water and accordingly $\pm\Delta r_{(u)}$ to $\pm\Delta G_r$ and $+D_{(u)}$ to D_t.

The amount of river water for the basin considered may be measured by

$$\begin{aligned}[Q'] &= R_W \pm \Delta S \mp V \pm \Delta R \\ &= (R_W \mp V \pm \Delta R + S_2) - S_1 \\ &= (1 - \alpha_2)(R_W \mp V \pm \Delta R + S_2) \\ &= (1 - \alpha_2)(R_W \pm V \pm \Delta R) + (1 - \alpha_2)S_2'\end{aligned}$$

If some quantity of water is imported artificially from other basins, derived by rivers or channels, or recharged into shallow or deep wells or, reversely, water of rivers, shallow or deep wells is expelled to be used in other basins, then $+W_{imp} - W_{exp} = \pm\Delta W$ must be added to the water supply source or P, and it is distributed to river water, free or confined groundwater respectively by $+R_{imp} - R_{exp} = \pm\Delta R$, $+q'_{imp} - q'_{exp} = \pm\Delta q'$ or $+q''_{imp} - q''_{exp} = \pm\Delta q''$. Thus every hydrological factor may be modified as in the following.

$$P \rightarrow P \mp V \pm (\Delta W)$$

$$D_t \rightarrow D = D_t + D_{(u)} + (D_{imp})$$

$$\pm\Delta G_r \rightarrow \pm\Delta_r = \pm\Delta G_r \pm \Delta_{r(u)}$$

$$G_W{}' \rightarrow G' = G_W{}' \mp \Delta fc \mp \Delta g' \mp \Delta S' \pm \Delta G'_{(u)} \pm (\Delta q')$$

$$G_W{}'' \rightarrow G'' = G_W{}'' \pm \Delta fc \pm \Delta g'' \mp \Delta S'' \pm \Delta G''_{(u)} \pm (\Delta q'')$$

$$R_W \rightarrow Q = R_W \pm \Delta S \mp V \pm \Delta R_{(u)} \pm (\Delta R)$$

These expressions are summarized in the following table.

Table
WATER BALANCE

	Natural State		Regulation		Utilization						Adjustment for	
Precipitation	Inter-flow between river and ground-water	Inflow and Outflow of ground-water	Lakes, ponds, veservoirs		River water	Shallow wells (free ground-water)	Deep wells (confined ground-water)	Imported water	Exported water	Total	Inflow or imported water from other	Outflow or exported water to basins
D_t				$+D_{(u.1)}$	$+D_{(u.2)}$	$+D_{(u.3)}$	$+D_{(u.4)}$	$+D_{imp}$		D		
				$+D_{(u)}$								
G_r				$\pm\ r(u.1)$	$\pm\ r(u.2)$	$\pm\ r(u.3)$	$\pm\ r(u.4)$	$\pm\ r.imp$		$\pm r$		
				$\pm\ r_{(u)}$								
G'_w	$-S_2'+S_1'$	$+g_n-1-g_n'$		$+G'_{(u.1)}$	$+G'_{(u.2)}$	$-G'_{(u.3)}$	$+G'_{(u.4)}$	$+q'_{imp}$	$-q'_{exp}$	$+G'$	$-g'_{n-1}$	$+\left[g'_n\right]$
	$=\mp\Delta S'$	$=\mp\ \Delta g'$		$\pm G'_{(u)}$				$\pm\ \Delta q'$			$-q'_{imp}$	$+\left[q'_{exp}\right]$
G_w''	$-S_2''+S_1''$	$+g_v''+g_n''$		$+G''_{(u.1)}$	$+G''_{(u.2)}$	$+G''_{(u.3)}$	$-G'_{(u.u)}$	$'q''_{imp}$		$+G''$	$-g''_r$	$+\left[g''_n\right]$
	$=\pm\Delta S''$	$=\mp\ \Delta g''$		$\pm G''_{(u)}$				$\pm\ \Delta q''$			$-q''_{imp}$	$+\left[q''_{exp}\right]$
R_w	$+S_2-S_1$	$+Q_{n-1}$	$\mp V$	$-V_{(u.1)}$ $(\pm V_2)$	$-R_{(u.2)}$	$+R_{(u.3)}$	$+R_{(c.4)}$	$+R_{imp}$	$-R_{exp}$	$+Q_n$	$-Q_{n-1}$	$+\left[R_{exp}\right]$
	$=\pm\ S$ $(\pm\Delta S'\pm S'')$			$\pm R_{(u)}$				$\pm\ \Delta R''$			$-R_{imp}$	
P			$\mp V$					$+W_{imp}$	$-W_{exp}$	$\pm V\pm\ W$	$-W_{imp}$	$+\left[W_{exp}\right]$

§4. Estimation of groundwater supply

Let us consider relations between groundwater storage G and river discharge Q' for precipitation P in natural state for a separated individual basin.

	Condition	Groundwater storage G_{at}	River discharge Q'nat	
1.	$S_2 = \ {}_1P \mp \Delta g$ $S_1 = 0 \qquad {}_2 = 0$	${}_1P \mp \Delta g - S_2 = 0$	$R_w + \ {}_1P \mp \ g$ $= P - D \mp \ g$	Q' > 0
2.	S_2 $S_1 = 0 \qquad {}_2 = 0$	${}_1P \mp \Delta g - S_2$	$R_w + S_2$	
3.	S_2 $S_1 \qquad {}_2 < 1$	${}_1P \mp \Delta g \mp \Delta S$ $= \ {}_1P \mp \Delta g - S_2 + S_1$ $= \ {}_1P \mp \Delta g - S_2 +$ ${}_2 (R_w + S_2)$ $= \ {}_1P \mp \Delta g - (1 - \ {}_2) S_2$ $+ \ {}_2R_w$	$R_w + S_2 - S_1$ $= R_w \pm \ S$ $= (1 - \ {}_2)(R_w + S_2)$	
4.	$S_2 = 0$ $S_1 = R_w \qquad {}_2 = 1$	${}_1P \mp \Delta g + R_w$ $= P - D \mp \Delta g$	$R_w - S_1 = 0$	Q' = 0
5.	$S_2 = 0$ $S_1 \quad R_w \qquad {}_2 = 1$	${}_1P \mp \Delta g + R_w + S_1'$ $= P - D \mp \Delta g + S_1'$	$-S_1'$	Q' < 0 Extract from Q

These relations are diagramatically shown in Fig. 4.

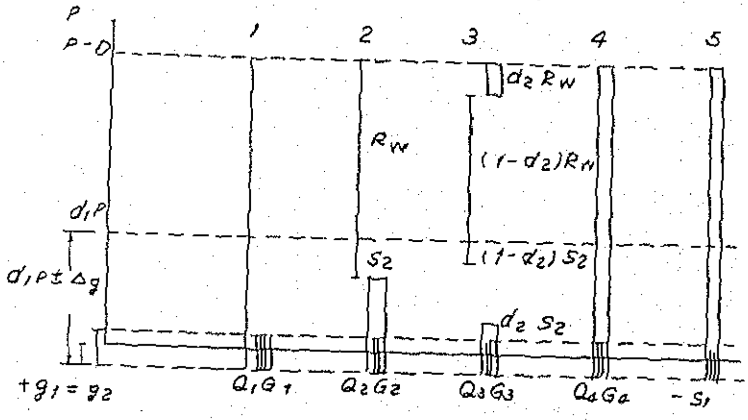


Fig. 4

Now we can consider as in the following. If there are abundant areas of easily infiltrating rocks, average coefficient of infiltration α_1 may be larger and R_W may become smaller. There may be probable for S_2 to decrease as rocks are more permeable. When river bed is composed of more previous detritus, then the coefficient of influent flow α_2 shows larger values or S_1 may increase. Accordingly $Q^1 = R_W + S_2 - S_1$ will take after all smaller values. In a basin composed of mainly non-previous rocks, infiltration is small and R_W is larger, S_2 being relatively larger especially in valleys of steeper slopes and S_1 being small when bed deposits are not thick and consequently $Q' = R_W + S_2 - S_1$ will assume larger values. It may be said thus that the groundwater supply is large when Q' is smaller, and it becomes smaller when Q' is larger.

When the water balance is modified artificially, every equation for groundwater storage and for river discharge must be modified according to explication above described so that every condition may be satisfied.

CHAPTER III. SOIL AND ITS CLASSIFICATION

§ 1. Definition of soil

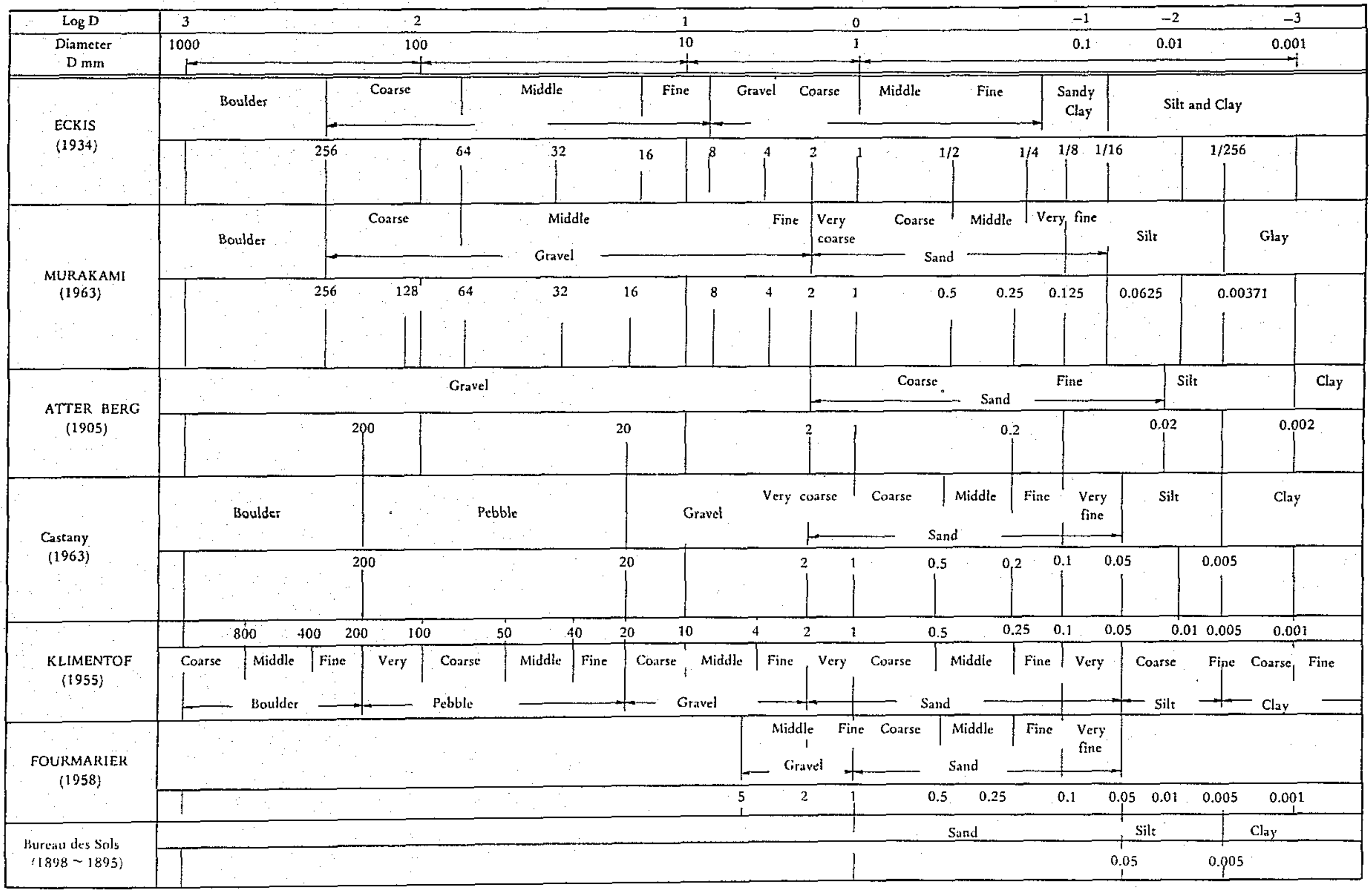
Soil is considered to cover at least 95 per cent of the land surface (Shaler, N.S.). The geologists in earlier times treated the soil as "merely the fine communited end product of disintegration without particular form or recognizable characteristics" as being pointed out by C.F. Tolman in his book "Ground water", and also they thought that "the soil often hinders geological research by covering the rocks in place" as L.E. de Beaumount described in the "Leson dé geologie pratique" (1845). Thus they did not show much interest in soil, but the soil is very important from the hydrogeological point of view. Because it is situated on the land surface between the atmosphere and hard rocks underneath, and it is through the soil that infiltrates a part of precipitation to contribute to the groundwater. Soil is defined as the layer or mantle of mixed mineral and organic material penetrated by roots of plants (Tolman). The roots are compactly compressed by soil, being precipitated around them by cementing material and the opening of root passages remains after the decay of organic material encased therein. Thus the vertical infiltration of water is made easy or a "vertical previousness" is produced, resulting formation of different soil zones perpendicular to the direction of roots penetration. Especially at or near the ground surface the soil is subject to repeated wetting, partial drying and leaching by organic and mineral solution, and thus deposition and cementation happen in the intermediate and lower soil horizons. C.F. Shaw (1928) called "eluviation" for leaching and "illuviation" for deposition.

The soil zones were grouped by a Russian pedologist K. Glinka in three horizons A, B, and C.

Horizon A is the surface soil and is characterized by mineral decomposition, transformation of the mineral constituents and organic debris into colloids and the solution and abstraction of the soluble compounds (leaching). The texture is granular or powdery. In clay soils, cracks are produced by shrinkage due to drying. Prismatic or columnar structure is often observed in them. It is previous and water is absorbed easily and seeps downward. The thickness of this horizon rarely exceeds 40cm.

Horizon B is the subsoil. In this zone the dissolved material in the horizon A deposits. Calcium carbonate or clayey material containing aluminium and ferric hydroxide accumulates to become hardpans. When this accumulation takes place, the subsoil is more or less firmly cemented and limits downward movement of roots and water. This horizon may be previous due to aggregate grain structure, flocculated colloidal matter and vertical parting. But it is usually less previous owing to the fact that clayey material and calcareous cements are richer than horizon A. The clayey material may give it a large absorptive capacity. Horizon B is usually less than 1 meter thick.

Horizon C is the substratum. This is the portion of decomposed rock in place or freshly deposited alluvium or other sedimentary material. This horizon is commonly not more than 2 meters in thickness but may extend to 20.


Log D
3
2
1
0
−1
−2
−3
Diameter D mm
1000
100
10
1
0.1
0.01
0.001
ECKIS (1934)
Boulder
Coarse
Middle
Fine
Gravel
Coarse
Middle
Fine
Sandy Clay
Silt and Clay
256
64
32
16
8
4
2
1
1/2
1/4
1/8
1/16
1/256
MURAKAMI (1963)
Boulder
Coarse
Middle
Fine
Gravel
Very coarse
Coarse
Middle
Very fine
Sand
Silt
Glay
256
128
64
32
16
8
4
2
1
0.5
0.25
0.125
0.0625
0.00371
ATTER BERG (1905)
Gravel
Coarse
Fine
Sand
Silt
Clay
200
20
2
0.2
0.02
0.002
Castany (1963)
Boulder
Pebble
Gravel
Very coarse
Coarse
Middle
Fine
Very fine
Sand
Silt
Clay
200
20
2
1
0.5
0.2
0.1
0.05
0.005
KLIMENTOF (1955)
800
400
200
100
50
40
20
10
4
2
1
0.5
0.25
0.1
0.05
0.01
0.005
0.001
Coarse
Middle
Fine
Very
Coarse
Middle
Fine
Coarse
Middle
Fine
Very
Coarse
Middle
Fine
Very
Coarse
Fine
Coarse
Fine
Boulder
Pebble
Gravel
Sand
Silt
Clay
FOURMARIER (1958)
Middle
Fine
Gravel
Coarse
Middle
Fine
Very fine
Sand
5
2
1
0.5
0.25
0.1
0.05
0.01
0.005
0.001
Bureau des Sols (1898 ~ 1895)
Sand
Silt
Clay
0.05
0.005

§ 2. Classification of soils in grain size

Before discussing on porosity (ratio of volume of space among grain particles to the whole volume) and permeability (degree of perviousness or easiness of percolation of water), we must know how the soil may be classified physically according to its grain size.

Classification of soils by geologist and pedologist is different. The geologist treats sedimentary formations which are usually more uniform in composition. He classifies the grains into gravel, sand, silt, particle clay and colloidal clay. When several grains are mixed in the soil, he names it according to the more predominant constituent of grains, and the component material in minor quantities is used as an adjective as sandy clay, silty sand, or sandy gravel etc. While the pedologist names the soil as the mixture of several size of grains: gravel, sand, loam, silt and clay.

Several classifications are used in different countries and by different authors, and their limits of grain size for classification do not always accord each other. They are shown for reference in the table (p.25). For example, the classification by Castany (1963) is as follows:

Diameter of grains (mm)		Sediments	
	–20	Pebble	
20	– 2	Gravel	
2	– 1	Very coarse sand	Coarse sand
1	– 0.5	Coarse sand	
0.5	– 0.2	Middle sand	
0.2	– 0.1	Fine sand	Fine sand
0.1	– 0.05	Very fine sand	
0.05	– 0.005 (5)	Silt	
0.0005	–	Clay	

When it is needful of more precise classification of particles of clay and colloidal clay, the decantation method is used and descending velocities of particles in the water vessel are measured.

§ 3. Representation of mechanical analysis

The soil sample is separated by sieves of different diameters and the weight of constituent grains for every sieve is measured and represented in percentage of the whole sample. This way of classification of soils is called the mechanical analysis or granulometry. There are three ways of diagramatical representation of the results.

1. Histogram

If the weight in per cent for every range of diameters is plotted in rectangular scale, taking diameters in abscissa and percentage in ordinate, we obtain a histogram which shows a step from diagram. We can assume a curve by joining middle points for every range of diameters of grains.

2. Cumulative curve

Compared with the above method, it is more convenient in practical use to plot them on a semi-logarithmic paper, taking diameters (mm) in abscissa and accumulated values of weight percentage in ordinate. Then a cumulative curve is obtained, covering from the maximum diameter to the minimum diameter. If we draw the curve taking a larger diameter in the left and a smaller diameter in the right hand side side, the total amount of grain constituents smaller than some diameter is represented by the right-hand side portion of the curve. The diameter d_x is the diameter, the grains of diameters smaller than which is contained by x% in weight in the whole sample. The weight percentage is easily obtained on the curve for range between any two arbitrary diameters.

The diameter d_{10}, the particles smaller than which is contained by 10% in the sample, is called the effective diameter, and this value is sometimes used as a criterion to distinguish several soils concerning their perviousness. The diameter d_{60} is obtained on the curve in the same manner.

The uniformity coefficient is the ratio of d_{60} to d_{10}, and it is sometimes called Hazen's coefficient U, using the name of the scholar who proposed it:

$$U = \frac{d_{60}}{d_{10}} \qquad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots \qquad (3.1)$$

When the value of U approaches as nearer as 1, the grains are of nearly the same size and have a larger porosity. The grain size distribution is considered uniform when $U < 2$, and non-uniform when $U > 2$.

K. Terzaghi and R.B. Peck propose d_{20} for the effective diameter and d_{70} in stead of d_{10} and d_{60} respectively in their recent study.

3. The U.S. Department of Agriculture uses the classification proposed by R.E. Davis and H.H. Bennett (1927). This classification comprises three components in weight percentage of sand, silt and clay (Table, p.29). They represented these classified soils on the equilateral triangle diagram which is shown in Fig. 5.

§4. Soil profiles and soil maps

In the soil survey, we take soil samples at different places as well as at different depths. We classify these samples by mechanical analyses and represent the results in soil profile and on a soil map. The former shows the vertical distribution of soil grains and the latter the horizontal distribution of surface soils (horizon A). Based on these representations, we can deduce the hydrologic properties or the infiltration capacity (Ref. to Chap. V) of the soils which enables to draw a hydrogeological map of the region concerned.

Soil classes		Sand (%)	Silt (%)	Clay (%)
1. Clay	Clay	0 ~ 50	0 ~ 50	30 ~ 100
2. Sandy clay		50 ~ 70	0 ~ 20	30 ~ 50
3. Silty clay		0 ~ 20	50 ~ 70	30 ~ 50
4. Sandy clay loam	Loam	50 ~ 80	0 ~ 30	20 ~ 30
5. Clay loam		20 ~ 50	20 ~ 50	20 ~ 30
6. Silty clay loam		0 ~ 30	50 ~ 80	20 ~ 30
7. Sand		80 ~ 100	0 ~ 20	0 ~ 20
8. Sandy loam		50 ~ 80	0 ~ 50	0 ~ 20
9. Loam		30 ~ 50	30 ~ 50	0 ~ 20
10. Silty loam		0 ~ 50	50 ~ 100	0 ~ 20
11. Silt		0 ~ 20	80 ~ 100	0 ~ 20

by Davis, R.E., and H.H. Bennett (1927)

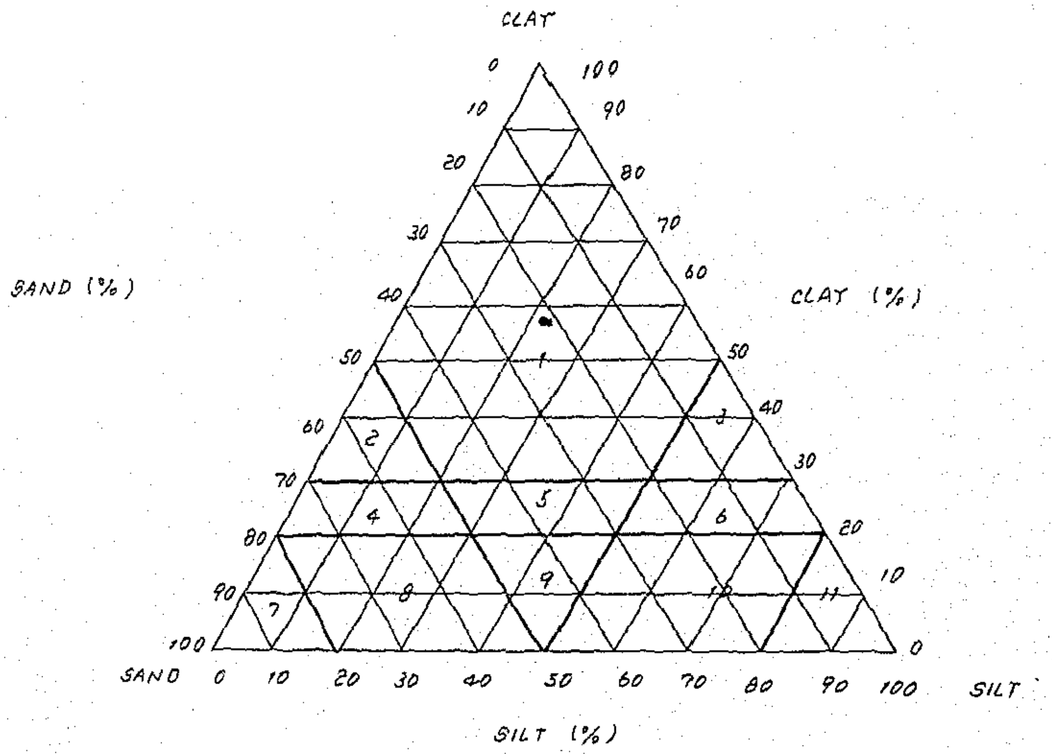


Fig. 5

CHAPTER IV. HYDROLOGIC PROPERTIES OF SEDIMENTS AND ROCKS

§ 1. Openings in sediments and rocks

The portion of sediments or rocks which is not occupied by solid minerl matter represents space and may be occupied by groundwater. This space is called opening, voids, interstices, pores or pore space. When the sediments or rocks are saturated with water, total space is filled with water, and if a part of water is expelled by some forces such as vapor pressure, gravity or pumping, the degree of saturation decreases and the space is occupied partly by water and partly by air. A proportion of total voids included in a unit volume of material is not the same for different sediments and rocks, and this proportion varies always according to the conditions, dry or wet, of the material. We must thus define "porosity" or the property of containing openings and "coefficient of saturation".

§ 2. Porosity

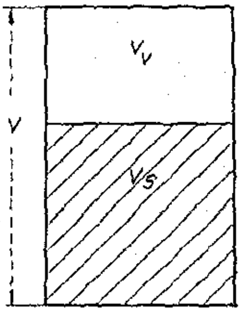


Fig. 6

There are two kinds of porosity: total porosity and effective porosity.

(1) Total porosity (m)

It is a ratio of the volume of space, V_V, to the total volume of the sediment or rock, V, and is denoted in percentage or

$$m = \frac{V_V}{V} \times 100 \qquad \text{.......... (4.1)}$$

The total porosity is called also true porosity, coefficient of porosity or modulus of space.

If we denote the volume of the solid part (mineral matter and organic matter) as V_S, then

$$V_V = V - V_S$$

and

$$m = \frac{V - V_S}{V} \times 100 \qquad \text{.......................... (4.2)}$$

Space index, e, is a ratio of the volume of space, V_V, to the volume of the solid part, V_S, or

$$e = \frac{V_V}{V_S} \qquad \text{.......................... (4.3)}$$

Accordingly the following relation is deduced between the total porosity and the space index:

$$m = \frac{e}{1 + e} \tag{4.4}$$

and

$$e = \frac{m}{1 - m} \tag{4.5}$$

Total porosity is considered according to its value as being

small	for	5% or less
medium	for	5 ~ 15%
large	for	over 15%

(2) Effective porosity (m_e)

A part of the water filled in the total space in the sediment or the rock percolates downward by gravity action and it is called gravity water, whose volume is denoted V_e (ref. to § 6 (iii)). The ratio of the volume of gravific water, V_e, to the total volume of the sediment or the rock, V, is called effective porosity and is denoted as m_e in percentage, or

$$m_e = \frac{V_e}{V} \times 100 \tag{4.6}$$

§ 3. Total porosity observed in sediments and rocks

We can summarize here the range of the total porosity for various kinds of sediments and rocks.

(1) Sediments

The highest porosity in natural sediments is reported by Meinzer as 80 to 90 per cent for freshly deposited alluvium (which is called "slurries") of the Mississippi. Such high porosity is very rare except in alluvial or mudflow deposits. The total porosity of sediments varies in general from 5 to 50 per cent with an average value of 15 per cent.

Gravels	25 – 40%
Gravels of 4mm	36
Sands and gravels	25 – 30
Sands	25.95 – 47.64
Recent alluvium	5 – 15
Alluvium of the Rhein (Fessenheim)	20
Arkose sand	44 – 50
Clay	44 – 50
Marl	47 – 50
Recent mud	80 – 90
Loam	34 – 50

Uncemented alluvial material such as clay, mud and loam may have porosity as high as 50 per cent. Uniform sand, especially quick sand, may have porosity nearly as 45 per cent. In gravel it varies according to variation of volume of interstitial material, showing from 25 to 40 per cent. A mixture of gravel, sand and silt assumes in general a small percentage, because fine particles fill a pore space between gravels and sands.

(2) Rocks

The total porosity of rock material depends on fissures and fractures contained in it, and they are produced either by primary or secondary process or in combination of both of them, about which will be described in the next paragraph.

Limestone	0.5 – 17%
Chalk in the North of France	22 – 37
Chalk	14 – 44
Oolitic limestone	3 – 20
Marble of Carrare*	0.10 – 0.20
Shale	1 – 10
Sand stone	4 – 26
Dolomite	2.9
Granite	0.02 – 1.5
Gypsum	4
Basalt	0.1 – 2.9

* Carrare (Toscane, Italy)

§4. Types of openings

There are two types of openings: primary and secondary openings. The opening which was created when the sedimentary or igneous rock was formed or which is originated from the geological processes is primary opening. While the opening which was developed secondarily after the rock was formed is called secondary opening.

Openings may be classed also by their sizes as supercapillary, capillary and subcapillary. Capillary openings are sufficiently small so that water is held by surface tension: supercapillary openings are those larger than capillary ones and subcapillary openings are so small that water is held mainly by adhesive forces. Openings are sometimes connected each other and sometimes isolated: they are called respectively as communicating or interconnected and isolated.

(1) Primary openings

(a) Sediments and sedimentary rocks

When the sediments deposit, many openings are produced primarily, and size, shape and manner of arrangement of grains determine the porosity of rocks.

(b) Igneous rocks

In crystalline rocks openings occur with subcapillary size between crystals and they are generally not interconnected. Water moves only when fractures are formed, by secondary forces, connecting these openings.

In volcanic rocks, many openings are observed in lava flow as in form of amygdaloidal or gas cavities. They are usually isolated enclosed in dense lava, and the water movements do not occur except when they are connected by fractures produced due to some secondary forces such as weathering or crusta movements. Certain amygdaloidal lavas are reported having 80 per cent pore space, but nevertheless water can not be obtained unless the rock is broken. Such lava sometimes produces abundant quantity of groundwater when these cavities are communicated by many fractures which were formed by latter processes.

(c) Metamorphic rocks

Gneiss and crystalline schists are subject to deformation by secondary crustal movement, and there produce many openings between their schistositie.

(2) *Secondary openings*

Joints, fractures, solution openings and openings formed by plants and animals are the secondary openings.

The fractures are produced by crustal movements. Larger fractures, less in number, often extend to great depth, but are observed many small fractures near the surface. Faults are fractures formed by the dislocation of two blocks of rock along the sliding plane, and it is very important when they are open and keep much quantity of groundwater. But most faults are incipient closed tightly and impervious to water.

Solution openings are commonly seen in limestone and chalk. They are connected by fracture, sometimes forming caverns, and the tubular openings are considered converging toward a discharge exit, often located in a deep valley. The groundwater flows turbulently forming an underground river and emerges as a large spring on the ground.

§ 5. Factors affecting openings in sediments

The total porosity of sediments is affected by "pore patterns" as named by Tolman for factors which control the pore space, that is (a) quality of materials, (b) form, (c) dimension of grains and degree of sorting and (d) arrangements or orientation of grains which compose them.

(a) The porosity is high when the rock fragments or pebbles composing sediments are porous like agglomerate or pumice.

(b) When the sediments are composed of cubes, the porosity is larger than those composed of spheres. Likewise aggregation of angular materials as seen in the deposit of colluvial fan shows higher porosity than the deposit of beach or dune sand. Such difference of porosity becomes negligible as the size of composing material becomes smaller.

(c) In well sorted sediments their composing grains are assemblage of equal size and they show a high porosity. While in unsorted or poorly sorted sediments, grain size covers a wide range of diameters and finer particles are filled in pore space between coarser materials and it reduces the porosity. A deposit of coarse materials seems having high porosity than fine materials, but it shows, on the contrary, low porosity as it is mostly at the same condition in nature as in a poorly sorted case.

Even if the sediments are well sorted, the pore spaces between them are diminished by cementation and they show a low porosity. And consequently the heterogeneity of size of materials shows a tendency to diminish porosity.

It is due to this reason that we define ten per cent size of grains, d_{10}, as an effective size and we attach importance to the uniformity coefficient, $\frac{d_{60}}{d_{10}}$, and the range of diameter represented by the steepest slope of the cumulative curve by the mechanical analysis.

(d) *In clayey and silty* sediments mica predominates very often. Silts which are freshly deposited with mica flakes oriented at random may have a high porosity and relatively large interstices, and the compacted silty deposits have mica flakes oriented nearly parallel and may have a low porosity and small interstices. Also in the gravelly sandy deposits, the pore patterns are always deformed due to the orientation of grains which are affected by compaction or any other secondary forces. The change of pore patterns disturbs the interconnection of pore spaces and affects, in turn, the perviousness of the formations.

Among above factors, (a) quality of materials and (b) form are observed by megascopically or microscopically in the laboratory, and (c) dimension of grains and degree of sorting are determined by the mechanical analysis. While concerning the arrangement or orientation of grains, it is not kept at the same condition when the sample of sediments is brought to the laboratory. It is because, by disturbance of arrangements, the results, obtained from the laboratory experiments, do not represent natural conditions. So it is very difficult to determine this character exactly.

§6. Retained water and free water

(1) *Adhesive water*

Adhesive water is the water which is absorbed on the surface of rock and it includes two types of water: hygroscopic water and pellicular water.

a) Hygroscopic water is absorbed on the surface of soil particles in form of isolated small cell. The quantity of this water depends on the porosity, humidity (degree of saturation), temperature and pressure of atmosphere, but the porosity is the main factor which controls it. Hygroscopic water varies in general from 0.2 ~ 0.5% for coarse sand to 15 ~ 18% for medium and fine sand. Hygroscopic water is detached only by calcination.

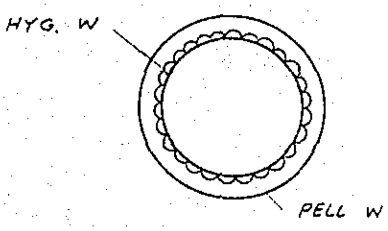


Fig. 7

Hygroscopic coefficient: – Maximum quantity of water, which is absorbed by dry soil in contact with air of 50% relative humidity at 25°C temperature is called hygroscopic coefficient.

b) Pellicular water covers the whole surface of particles of soil and also their surrounding hygroscopic water. The thickness of pellicular water has an order less than 0.1 micron. It moves by molecular *attraction* of the neighboring particles. A.F. Lebedev describes about this water as in the following.

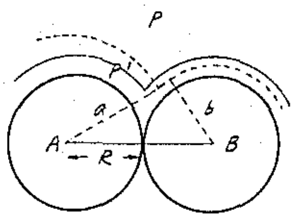


Fig. 8

We suppose the pellicular water which surrounds the two grains of sand which are represented by the spheres of equal diameter R, and the thickness of pellicular water for two spheres is different as shown in dotted line in Fig. 8. At the point P, the distance to the center of the sphere B is smaller than that to the center of the sphere A, and consequently the molecular attraction of B is larger than that of A. This difference of the *molecular attraction* of neighboring particles causes the movement of the water at P toward the sphere B and it continues till the thickness of the pellicular water surrounding A and B become the same. This movement is not due to gravity.

(2) Capillary water

Capillary water is the water which is kept in pore space by surface tension and is raised above the surface of groundwater by capillary force. Hydrostatic pressure is transmitted in it. There are two types of capillary water; isolated (or suspended) and continued.

Isolated (or suspended) capillary water is kept in suspension in a part of pore space between two particles of soil mixed with water vapor or air. It may not be removed by gravity and belongs to the retained water. It may be removed by centrifugal force. While the continued capillary water is just above the surface of groundwater body occupying the whole of pore spaces and capillary openings. It moves by gravity and it belongs to free water.

(3) Gravity water

Gravity water fills the openings other than the portions occupied by hygroscopic, pellicular and capillary water and freely moves by pressure difference. It infiltrates vertically through openings till it reaches to the surface of the groundwater body and percolate horizontally as groundwater flow. It is this free water that is utilized as groundwater. Every type of water under the ground before forming the groundwater is called subsurface water. Type of water is summarized as in the following table.

Type of water		Category	Extracted by
Adhesive water	Hygroscopic water	Retained water	Calcination
	Pellicular water		Centrifugal force
Capillary water	Isolated cap. w.		
	Continued cap. w.	Free water	Gravity
Gravity water			

§ 7. Degree of humidity and degree of saturation

(1) Degree of humidity (ω in weight percentage)

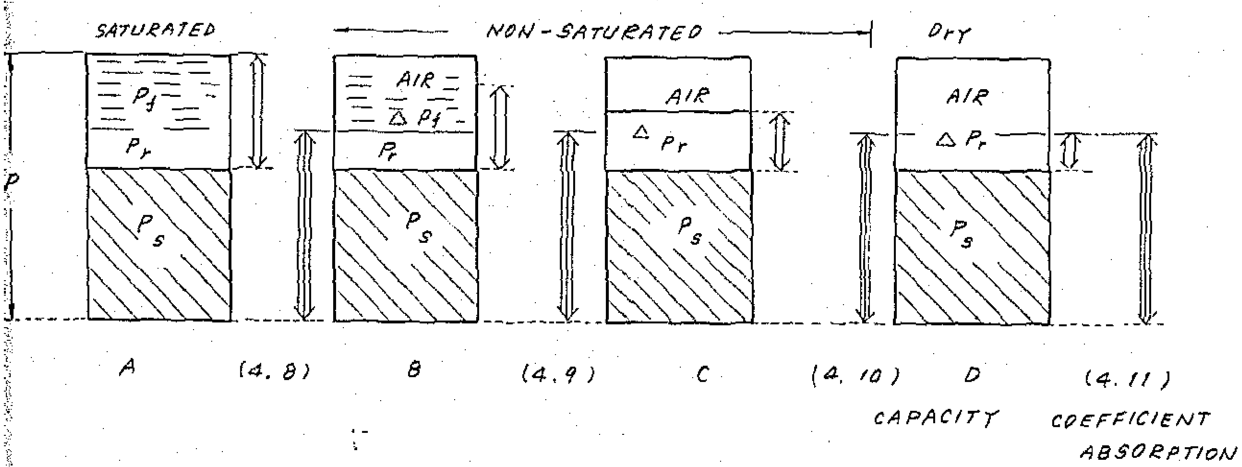


Fig. 9

When the sediment or the rock is saturated, the pore spaces are occupied partly by adhesive water (whose weight: P_r) and partly by free water (whose weight: P_f) ..... $P_r + P_f$ (A in Fig. 9). When it is non-saturated a part of the pore spaces is occupied by air instead of P_f, ... $P_r + \Delta P_f$ + air (B in Fig. 9). Even after the sample is dried at temperature 105°C, a small portion of P_r still remains, ... ΔP_r (D in Fig. 9). The weight of the dried sample is thus represented by

$$P_{dry} = P_s + \Delta P_r \quad \ldots\ldots\ldots\ldots \quad (4.7)$$

The ratio of the absorbed quantity of water to the dried sample is called degree of humidity; which is denoted by

Degree of humidity:

$$\text{when saturated,} \quad \omega A = \frac{P_r + P_f}{P_s + \Delta P_r} \times 100 \quad \ldots\ldots\ldots\ldots \quad (4.8)$$

$$\text{when non-saturated,} \quad \omega B = \frac{P_r + \Delta P_f}{P_s + \Delta P_r} \times 100 \quad \ldots\ldots\ldots\ldots \quad (4.9)$$

After all of the free water is abstracted by gravity or by pumping, the retained water only remains and its ratio is

$$\omega C = \frac{P_r}{P_s + \Delta P_r} \times 100 \quad \ldots\ldots\ldots\ldots \quad (4.10)$$

This is called field capacity.
When the sample is dried at 105°C, a part of adhesive water (hygroscopic water) still remains and

$$\omega D = \frac{\Delta P_r}{P_s + \Delta P_r} \times 100 \quad \ldots\ldots\ldots\ldots \quad (4.11)$$

This is called coefficient of absorption.

(2) Degree of saturation (s_r in volume percentage)

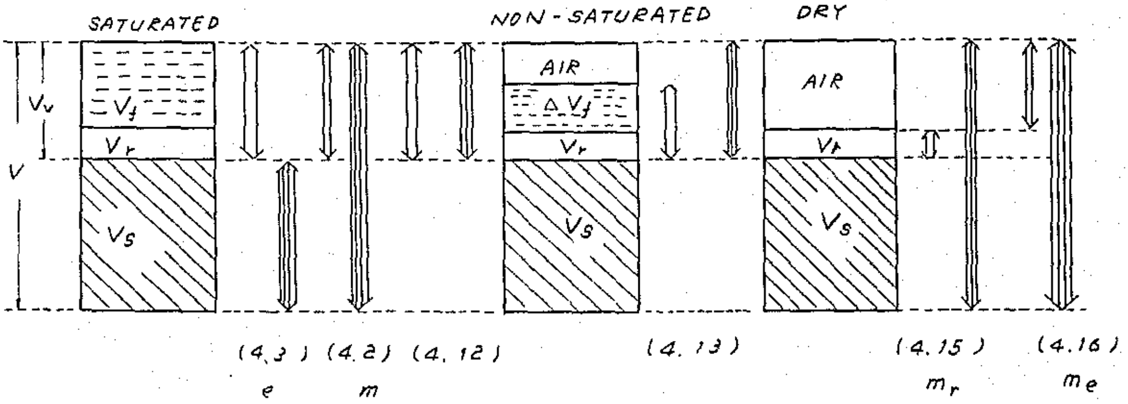


Fig. 10

When the quantity of water contained in the water bearing material is denoted by a ratio of the volume of water contained to the total volume of void, V_V, expressed in percentage, it is called the coefficient of saturation. When the sample is saturated with water,

$$s_r = \frac{V_r + V_f}{V_V} \times 100 = 100 \qquad \ldots\ldots\ldots\ldots \quad (4.12)$$

When it is non-saturated,

.... (4.13)

When it is dried in temperature 105°C, assuming ΔV_r being nearly negligible,

$$s_r = \frac{\Delta V_r}{V_V} \times 100 = 0 \qquad \ldots\ldots\ldots\ldots \quad (4.14)$$

According to K. Terzaghi and R.B. Peck, the coefficient of saturation is classed as follows.

Degree of saturation		Coefficient of saturation
dry	for	0%
a' little humid		1 ~ 25
humid		25 ~ 50
wet		50 ~ 75
very wet		75 ~ 99
saturated		100

§8. Specific retention and specific yield

Retained water remains after free water is abstracted by gravity. The ratio of the volume of retained water, V_r, to the whole volume of the sample, V, expressed in percentage, is called capacity of specific retention, coefficient of specific retention or simply specific retention, m_r, which corresponds to field capacity expressed in weight percentage, or

$$m_r = \frac{V_r}{V} \times 100 \qquad \text{(4.15)}$$

The volume of the abstracted water is the difference between the total pore space and the volume of the retained water, and this ratio expressed in percentage is called effective porosity or coefficient of specific yield or simply specific yield, which is the same as effective porosity already mentioned (2), or

$$m_e = \frac{V_f}{V} \times 100 \qquad \text{(4.16)}$$

Thus

$$m = m_r + m_e \qquad \text{(4.17)}$$

As the dimension of grains becomes larger and they are well sorted, the specific retention has a tendency to decrease, as shown in the next table by Hazen.

Effective diameter d_{10} mm	Uniformity coefficient U	Total porosity (%)	Specific retention (%)	Specific yield (%)
0.02	9	36	–	–
0.03	2.3	44	19	25
0.06	2.3	42	16	26
0.17	2	42	11	31
0.35	7.8	32.5	9.5	23
0.48	2.4	40	8	32
1.40	2.4	–	7.5	–
5	1.8	44	7	37

Accordingly as the specific yield or effective porosity increases, it permits to except more abundant quantity of groundwater.

The clay has a large total porosity. But it has a large coefficient of specific retention and accordingly has a small effective porosity. Thus the quantity of water which may be expelled from clayey layer is very small. The effective porosity of various kinds of material is summarized in the following table.

Gravel	25%
Sand, sand and gravel	20
Fine sand, sandstone	10
Clay and gravel, cemented gravel	5
Clay, lava	3
Alluvials	10 ~ 20
Alluvials of the Rhine	2 ~ 3
Chalk	2 ~ 5

The general tendency of change of total porosity according to grain size and the relation between the specific retention and the specific yield are shown diagramatically in Fig. 11.

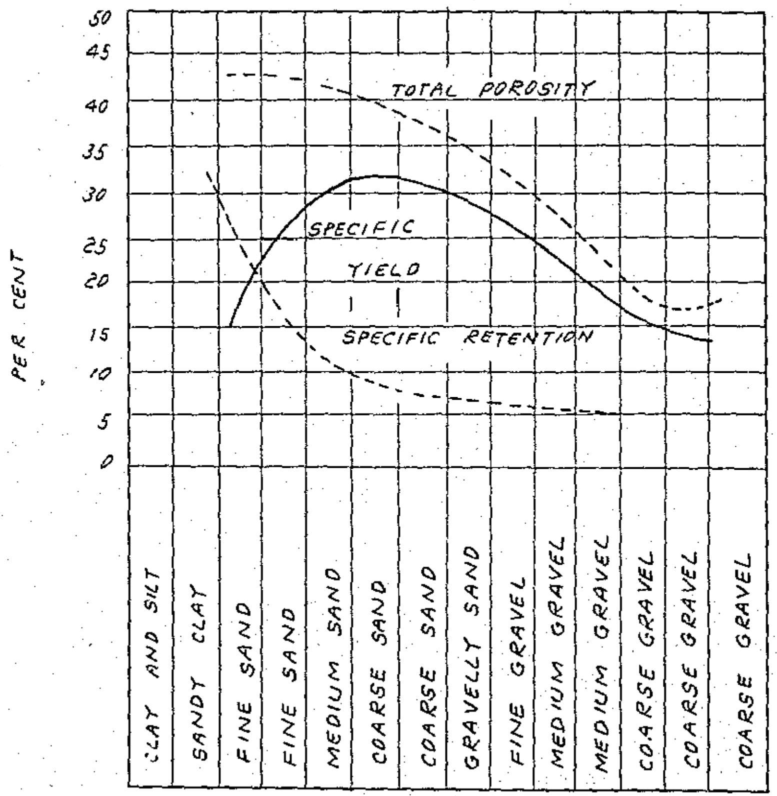


Fig. 11

§ 9. Permeability

The portion of pore space in which free water moves by gravity was defined as effective porosity. Even when the effective porosity is the same for different kinds of water-bearing material, their property for percolation of water or the rate of water movement in horizontal direction through them, that is, the quantity of water which moves through them during a certain period of time is different by material. This property is defined as permeability and its detailed description shall be given later.

CHAPTER V. QUALITY OF GROUNDWATER

§ 1. Quality

Groundwater contains various substances chemically dissolved in it. Its quality is originally of atmosphere, and carbon dioxide, CO_2, other gasses and minute particles in it are seized, in the first place, by precipitation. Precipitation water and surface water infiltrates and percolates in the ground and contacts with soils and rocks, in course of its passage through their interstices, when their constituent mineral substances are dissolved. Dissolved matter in subsurface water becomes more and more concentrated as it travels for longer distance and as the duration of contact is elongated. On the other hand, the quality of groundwater varies at different places and changes as it percolates along a certain path. Thus original quality of groundwater is modified by geology of the aquifer, and the distribution of quality and concentration of salts make it possible to suggest the direction of flow.

The salt concentration is also modified by various means: excess irrigation water adds substantial quantity of salts to groundwater. In coastal areas sea water intrudes into the aquifer. In the vicinity of some mines or industrially developed areas, groundwater is often polluted by waste water.

Groundwater is sometimes turbid with suspended matters, or colorful with organic colloidal matter. Turbidity or coloring is due to contact of water with clayey beds or peat bog.

§ 2. Chemical analyses

The groundwater sample taken from a well is submitted to the laboratory for complete chemical analyses. The concentration of all of the inorganic constituents must be determined and also the measurement of pH and electrical conductivity is made.

(1) Concentration of inorganic constituents

The mineral matters are dissolved in groundwater as dissociated ions and several other minor constituents coexist in it. The common cations and anions, their equivalent weights and minor constituents are shown in Table 1 in the next page.

The equivalent weight of an ion is the ratio of its atomic weight to its valence, expressed in grams, and is also called the gram equivalent weight. That is the weight in grams of an ion which combines with one gram of hydrogen. Dividing the concentration of an ion in ppm by gram equivalent weight of the ion, we get the number of gram equivalents per million (epm) or the milligram equivalent per litre (meq/ℓ). This means the reaching value and is indicated by "r" before the element symbol; e.g., r·Ca for Ca. The sum of the reaching values for cations and the sum for anions is nearly the same and both is equal to half of the total reacting values or the total concentration in meq/ℓ. Total concentrations is designated by C. (Refer to Table 2 below.)

Table 1

Major constituents			
Common cathions	Equivalent weight	Common anions	Equivalent weight
Calcium (Ca)	20.04 gr	Carbonate (CO_3)	30.00 gr
Magnesium (Mg)	12.16	Bicarbonate (HCO_3)	61.01
Sodium (Na)	23.00	Sulfate (SO_2)	48.03
Potassium (K)	39.10	Chloride (Cl)	35.46

Minor constituents
Iron (Fe)
Aluminium (Al)
Silica (SiO_2)
Boron (B)
Fluoride (F)
Selenium (Se)

Table 2

Chemical Analyses of Selected Groundwaters in California*
(After Doneon)

Well number	Major constituents in meq/ℓ					
	rCa	rMg	rNa	rCO_3 + $rHCO_3$	rSO_4	rCl
1	1.41	0.44	0.89	1.88	0.33	0.34
2	0.21	0.05	2.42	1.20	0.67	0.68
3	0.24	0.02	7.28	2.39	2.48	2.47

* T odd, D.K.: Ground Water Hydrology. John Wiley, New York. 1959. p.182

For well No.1 in the table, we have

r.Ca	1.41	$r.CO_3$ + $r.HCO_3$	1.88
r.Mg	0.44	$r.SO_4$	0.33
r.Na	0.89	r.Cl	0.34
	2.74 meq/ℓ		2.55 meq/ℓ

Total concentration: $C = 2.74 + 2.55 = 5.29$

Now we calculate the reacting value of every ion in percentage of concentration.

r % Ca	26.6	r % CO_3 + r % HCO_3	35.6
r % Mg	8.3	r % SO_4	6.24
r % Na	16.8	r % Cl	6.43
	51.7%		48.3%

The difference between the sum for cations and that for anions must be within 6% of total*. But if there is much difference, one must conclude either that there are other undetermined constituents present or that errors exist in the analysis.

(2) Electrical conductivity

The electrical conductivity is the specific electrical conductance and denoted by EC or K (mho/cm). This value is the reciprocal of the electrical resistivity (ohm-cm) and increases in proportion to the concentration of ions. Conductance is a function of water temperature and we must specify the value of electrical conductivity at a standard temperature of 25°C.

The value of electrical conductivity is approximately proportional to the concentration of total dissolved solids for most natural waters. Consequently the measurement of electrical conductivity or electrical resistivity may be substituted for the measurement of total dissolved solids.

§3. Graphical representation

We explain here two methods for representing the results of chemical analyses.

(1) *Collins' diagram*

This method of representation is a bar diagram system suggested by Collins. Two columns are drawn side by side, one for cations and one for anions, as shown in Fig. 12. Reacting values of ions are represented in meq/ℓ (I) or in % (II).

The left column is divided into three sections taking values for cations, from below to above, in order of r.Ca, r.Mg, r.Na + r.K. The right column is divided in the same manner into three sections, taking values for anions r.CO_3 + r.HCO_3, r.SO_4, r.Cl from below to above. Thus the heights of columns for cations and for anions nearly coincide.

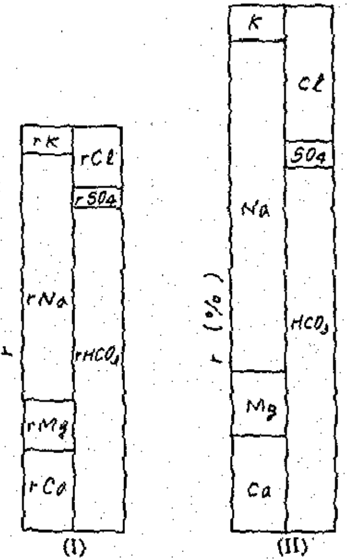


Fig. 12 Collins' diagram

Reacting value in
(I) meq/ℓ
(II) %

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* Castany, G: Traité Pratique des Eaux Souterraines. Dunod, Paris. 1963. p.589.

(2) Triangular diagram

An equilateral triangle is drawn and three laterals are taken as coordinate axes. Each axis is divided into 50 parts and the triangle is segmented into sections by lines parallel to each axis. The meq/ℓ is reduced to percentages. Cations and anions are separately grouped into three components. The summit of triangle represents 50% of each component against the opposit lateral, which represents 0%.

Plotting percentages of r.Ca, r.Mg, r.Na + r.K on coordinate axes and drawing three lines through these plots paralleling each axis, we get one point of intersection inside the triangle. This point represents the concentrations of component cations. In the same manner, by plotting r.CO_3 + r.HCO_3, r.SO_4 and r.Cl, we get one point inside the triangle, which represents the concentrations of component anions.

In these two diagrams coordinate axes for r.Ca, r.Mg and r.Na + r.K must correspond to those for r.CO_3 + r.HCO_3, r.SO_4 and r.Cl respectively. If we plot these two points of intersection for cations and anions on one triangular diagram and connect them, we can specify the groundwater type in quality according to the position of line. We can compare groundwater types for several wells by drawing such lines on the same triangular diagram.

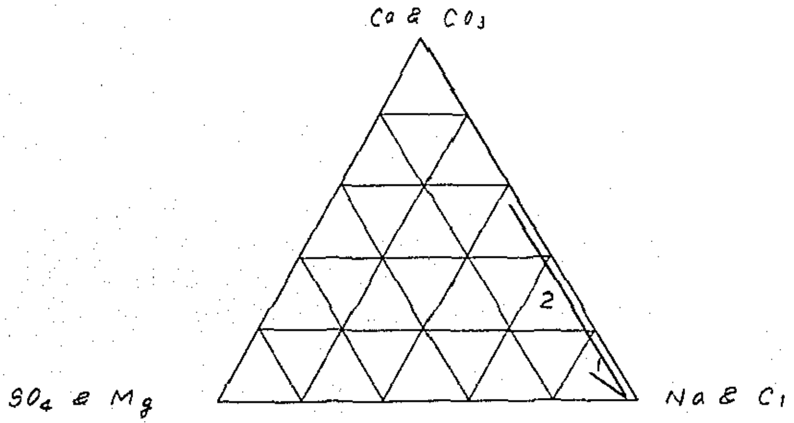

Fig. 13 Triangular diagram * anions (2) cations

CHAPTER VI. VERTICAL MOVEMENT OF WATER IN SOIL

§ 1. Movement of water in soil

The movement of water in soil is considered theoretically in two manners: the one in vertical direction and the other in horizontal direction and the other in horizontal direction.

§ 2. Vertical movement of water in soil and classification of soil zones

We shall treat in this chapter the movement in vertical direction and the distribution of humidity in soil, which is permeable and underlain by an impervious substratum like clay.

Suppose that water is poured in drop for a certain period of time uniformly on the surface of soil and the whole quantity of water infiltrates, not being lost by evaporation, into the soil. *This is the vertical movement of water. The degree of saturation of soil becomes larger and larger downward and it becomes* maximum or 100% when saturated.

Thus two zones may be distinguished in soil layers concerning *humidity* or *degree* of saturation:

1. zone of aeration
2. zone of saturation

(1) Zone of saturation

The soil near the bottom is saturated of water forming there a water forming there a water body, which is so called groundwater. Its upper limit is a free surface with the same hydrostatic pressure as the atmospheric pressure on the ground and is called water table. The part of soil layers occupied by groundwater is zone of saturation and degree of saturation here is 100%. When the pervious soil is homogeneous composed of sand grains of uniform size, the total porosity is represented as 40% of *the total volume.*

(2) Zone of aeration

While in the zone of aeration there is a capillary zone *called* capillary fringe immediately above the water table and water moves upward by capillary action to a certain height depending on assortment and arrangement of grains; in the homogeneous soil degree of saturation of soil changes from 100% immediately above the water table to 75% at the upper limit of the zone, that is the volume of water content changes from 40% to 30% of the whole mass.

The thickness of capillary fringe changes inversely proportional to grain size. It is about several decimeters in gravel, 30 ~ 60cm in sand, 120cm in sandy silt, 300cm in sandy silt, 300cm in silt and 3 ~ 4m in loam.

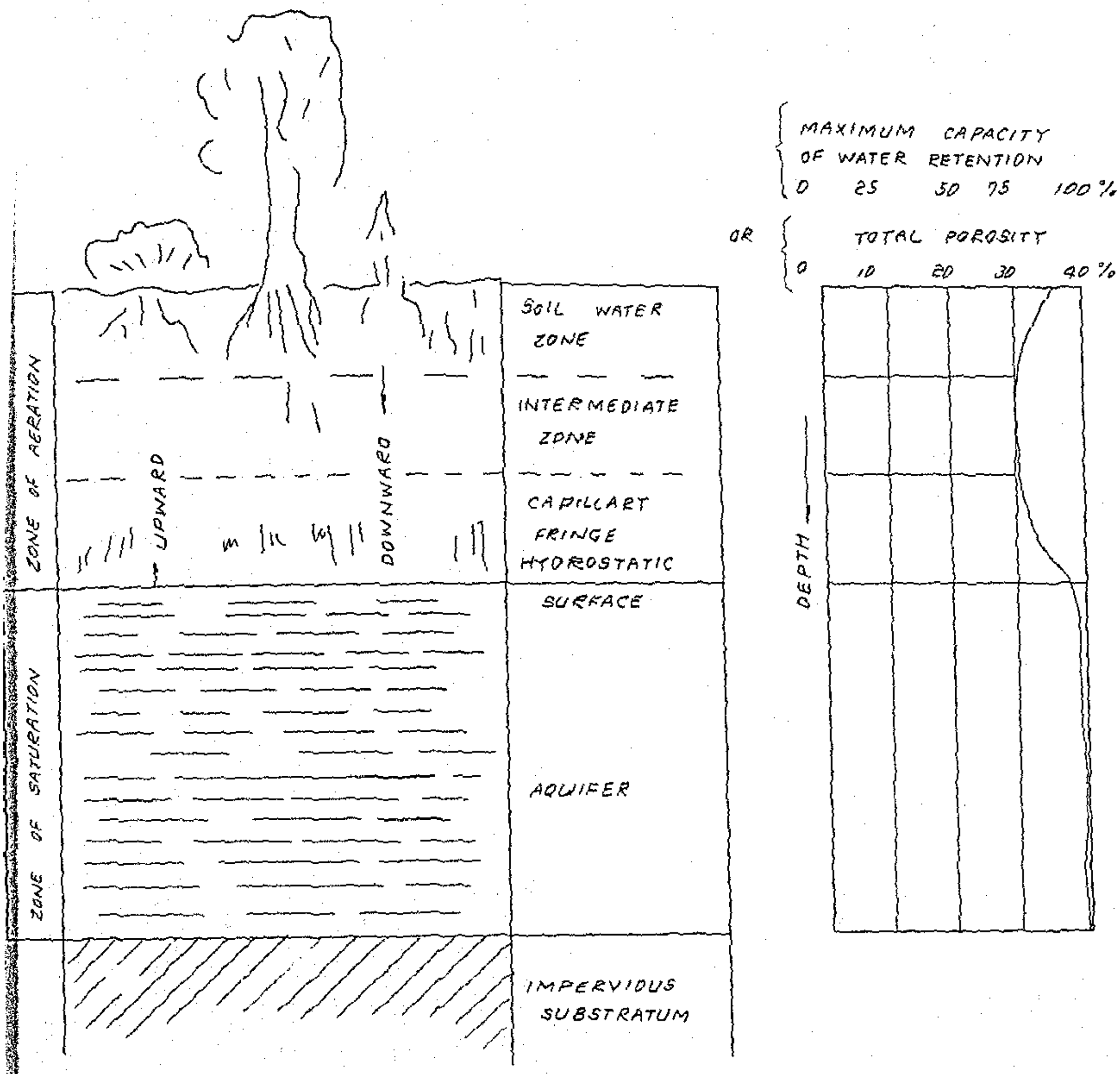

Fig. 14

There is an intermediate zone successively above the capillary fringe; degree of saturation in homogenous sandy soil is 75% in minimum or the volume of water occupies at least 30% of the whole mass. The void is occupied by water, air and water vapor. The water is kept retaining among soil grains as suspended water such as adhesive water and isolated capillary water. So this zone is also called zone of retention.

The thickness of the layer is from 0.6 to 2m, but it reaches sometimes to 10 ~ 20m or more.

Above the intermediate zone generally continues the soil water zone forming the uppermost layer of the zone of aeration and limited by ground surface. The soil shows in nature alternative layers and is penetrated by roots of plants. The water kept in this zone is partly consumed by transpiration through roots and partly lost by evaporation from the surface. Both of these movements of water are made upward. Thus it is the zone of evapotranspiration. The humidity may decrease below the coefficient of specific retention of soil. When it decreases to a limit of humidity for plant growth, the wilting of plant occurs and this limit is called coefficient of wilting of plants. In the desert region plant growth is only seen at the places where the humidity is kept over the wilting point.

Water is supplied, on the other hand, in this zone due to infiltration of rain and snow or by seepage of surface water. When water supply is excess in quantity compared with loss by evapotranspiration and makes the soil over-saturated, a swamp is produced there.

The thickness of this soil water zone is variable depending on climate and type of vegetation.

In some cases the capillary fringe continues directly to the soil water zone lacking the intermediate one.

§3. Infiltration formula in function of precipitation

Quantity of infiltration is estimated by measurements by lysimeter. The lysimeter is a receptrice of at least 150cm depth, inserted in the ground, filled with materials to be examined, and covered with some kinds of vegetation or naked. Rainfall received in the lysimeter infiltrates, moves downward and flows out from a funnel-shaped outlet, after saturating whole pore spaces of materials.

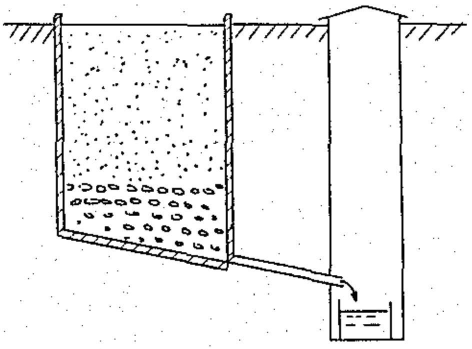

Fig. 15 Lysimeter

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* Bureau de Recherches Géologiques et minieres: Méthodes d'études et de recherches des nappes aquiferes. 1. publida, Paris. 1962.

There is a linear relation between rainfall, p mm, and quantity of infiltration, I mm:

$$I = a\,(p - b)$$

where a and b are constants:

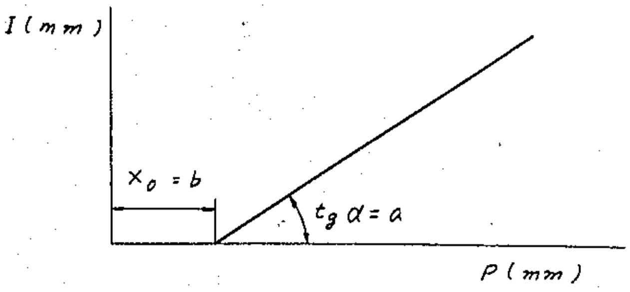

Fig. 16 Relation between infiltration and precipitation

For example, here are results obtained by lysimeters during a period 1943 ~ 1954 at Castricum, Holland, by lysimeters with $1m^2$ of surface and 1m depth.

Table 3

Observations by Lysimeters at Castricum, Holland

Year	Precipitation (mm)	Infiltration (mm)		
		I	II	III
1943	445.8	318.47	278.68	226.72
1944	579.8	464.61	411.78	376.72
1945	502.7	355.58	294.89	266.14
1946	531.3	404.85	360.60	325.95
1947	408.4	287.01	200.37	187.89
1948	390.9	275.05	178.32	154.71
1949	487	373.20	252.86	221.11
1950	671.2	518.75	363.55	344.94
1951	609.1	451.05	291.34	277.45
1952	590	475.18	313.74	299.55
1953	382.7	256.66	117.14	100.67
1954	561.2	434.18	249.04	232.23

Condition on the surface: I, naked soil; II, broad leave vegetation;
III, natural vegetation of dune

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[*] Castany, G.: Op. cit. p.166.

I) For the naked soil

Taking precipitation in abscissa and infiltration in ordinate on a rectangular scale, we draw a line which represents a general trend of arrangements of many plots. This line is shown by a line I in Fig. 17.

$b = x_0 = 172\text{mm}$

$a = \text{tg}\ \alpha = \text{tg}\ 43°31' = 0.95$

and $I = 0.95\ (p - 172)\ \text{mm}$

II) For the soil covered with broad-leaved trees

In the same manner, we get for this soil, a line II which is expressed by

$I = 0.9\ (p - 220)\ \text{mm}$

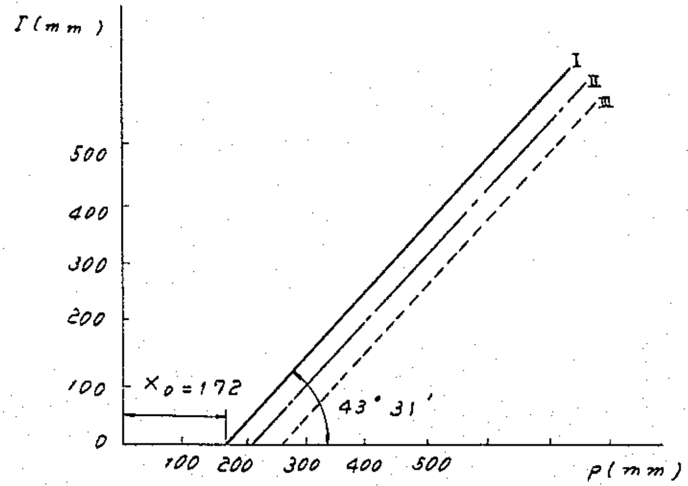


Fig. 17 Relation between infiltration and precipitation for lysimeters at Castricum, Holland. I, naked soil; II, soil with broad-leaved trees; III, soil with natural vegetation of dune.

III) For the soil covered with natural vegetation of dune

We obtain a line III which has a formula

$I = 0.95\ (P - 270)$

These results obtained by lysimeter experiments give general value of infiltration through the thickness of materials of 1m depth.

§ 4. Evapotranspiration

Evaporation from soil surface is not the same as that from water surface. The former amount is estimated by measurements by lysimeter. It is very difficult to measure separately amount of transpiration only, while it may be possible to estimate it by comparison of the measurements made in parallel by lysimeter for naked soil and for the soil with vegetation.

Another way to estimate the evapotranspiration is to use the balance formula. In the experiment by the lysimeter, run-off is null or $R = 0$. In neglecting the stock of water of soil we can write:

$$P = D + I \quad \text{.............................} \quad (5.6)$$

Applying the infiltration formula obtained by measurements by lysimeter on

$$I = a\ (P - b) \quad \text{.............................} \quad (5.1)$$

we get $D = P - I = (1 - a)\ P + ab$ .... (5.7)

In the example of Rothamsted, the mean annual precipitation is 707mm and $I = 360.5\text{mm}$

$$D = 346.5$$

CHAPTER VII. HORIZONTAL MOVEMENT OF WATER IN SOIL

§ 1. Percolation

Free groundwater is formed after the precipitation water infiltrated under the ground or the surface water seeped into the soil through beds of rivers or lakes etc. This water flows in a horizontal direction toward a river valley or some depression which is situated at a lower level. This movement is called "percolation". The velocity of this movement is controlled by the slope of the water table and affected by grain size and structure of media.

§ 2. Permanent and turburent flow

Sediments such as gravels, sand and clays are composed of different size of grains and show difference in effective porosity affected by their manner of sorting and arrangement. Some rocks are impervious of water, but have many fisures, cracks or other pores to permit easily entering of water in them, as seen very often in limestones and volcanic rocks.

Groundwater percolates in most cases as a laminar flow with a very slow velocity forming filtration tube through space among grains of the medium.

The laminar flow is a kind of permanent flow, which is defined as a current of water when the curve of the flow line is fixed in space and the water particle which comes successively always occupies the same position and has the same density; whose velocity may be constant or variable along the trajectory, and, as the discharge or the quantity of water per second is always constant, the sectional area of the flux of flow changes when the velocity changes.

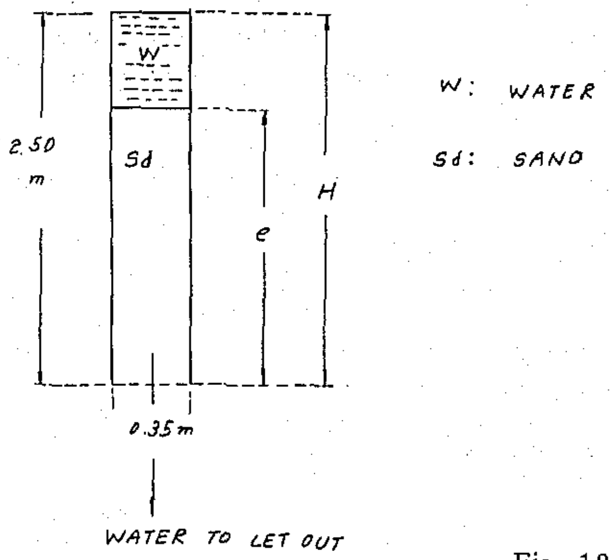
It is called a "uniform flow" when the permanent flow has a constant velocity with a non-changeable direction or the molecule of water does not change its movement in time and in space.

When the groundwater flows through cavities or cracks large enough in space, the flow is not already permanent, but turbulent, as it is said, and the movement of molecules of water always changes its velocity and direction.

§ 3. Darcy's law and permeability

A Frenchscholar Darcy found the following law concerning the flow of water through sandy column by his experiment in 1856.

He used a cylinder of 2.50m height and 0.35m of interior diameter. He packed in this cylinder gravels and sands as shown in Fig. 18; their granulometric composition in weight were:



Gravels in 17%
Gravels of 2mm in 12%
Sands of 1.1mm in 13%
Sands of 0.77mm in 58%

Fig. 18

He measured the water quantity which flowed out through the bottom during an unit duration of time and expressed this result by a formula:

$$Q = kS \frac{H}{e} \qquad \text{(6.1)}$$

where Q: discharge of water in m^3/s
S: sectional area in m^2
H: height of head in m
e: thickness of the column of material in m
H/e: loss of head for every unit length of the material or hydraulic gradient, i

The above formula is thus rewritten as

$$Q = kSi \qquad \text{(6.2)}$$

The rate of flow of water during a certain duration of time or velocity, v, through an unit sectional area is expressed by

$$v = \frac{Q}{S} = ki \qquad \text{(6.3)}$$

This is called Darcy's formula, where v is an apparent velocity, because S is the total area of the cross section, which comprises partly solid materials and partly spaces among them through which the water percolates. k is a proportional constant, which is called permeability and which differs for every material.

Strictly saying, it concerns with temperature accordingly specific gravity, δ, and viscosity, μ, of fluid and must be expressed as

$$k = ki \frac{\delta}{\mu} \qquad \text{(6.4)}$$

where k_i is true permeability.

Permeability is represented by the value at temperature 20°C and k is assumed being the same as k_i for groundwater, because the deviation of values of δ and μ due to temperature variation is very small and negligible. But we can not neglect such variation for a study of great depth as in the case of petroleum.

Hydraulic gradient, i, which means tan i, has no dimension of unit and i = 1 when the slop is 45°, then

$$v = k \quad \dots\dots\dots\dots\dots\dots\dots\dots \quad (6.5)$$

Accordingly k has the same dimension of unit as v $[\frac{L}{T}]$ or cm/s. The values of k for i = 1 are used when we compare their values for different materials.

We enumerate here data of k by some authors.

Dune sand in Holland	2×10^{-2} cm/s
Very fine sand	$10^{-2} - 2 \times 10^{-2}$
Sand with small amount of clay	8×10^{-2}
Gravelly sand	5×10^{-1}
Fine gravel (2 ~ 4mm)	3
Middle gravel (4 ~ 7mm)	3.5
Alluvium of the Rhine River	4.2×10^{-2}
Alluvium of the Durance (Serre Poncon)	5×10^{-2}

Generally speaking, values of k are given as in the following:
Permeability, k:

Gravel, coarse sand	$10 - 10^{-3}$ cm/s
Fine sand	$10^{-3} - 10^{-4}$
Silt	$10^{-4} - 10^{-6}$
Clay	$10^{-6} - 10^{-9}$

We classify the water bearing quality of rocks into several categories according to their permeability.

Permeability	Expression	Category of rocks (Description)
$k > 10^0$ cm/s	much permeable	Aquifer (Water circulates easily)
$10^0 > k > 10^{-4}$	permeable	
$10^{-4} > k > 10^{-7}$	little permeable	Aquiclude (Circulation of water is very small)
$10^{-7} > k$	impermeable	Aquifuge (Water does not circulate)

We can expect enough quantity of water from aquifer, but not from aquiclude and aquifuge.

§ 4. Determination of permeability in natural condition

Value of permeability which was determined experimentally in laboratory using sample of soils transferred in a vessel does not represent strictly that in natural condition.

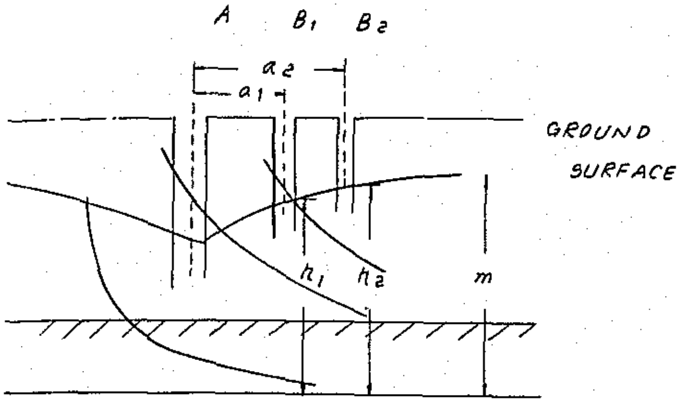


Fig. 19

We can determine permeability in the field by using two test wells (B_1 and B_2 in Fig. 19) which situate in the area of influence of a pumping well (A in Fig. 19).

When water is pumped up at the main well, A, flows of water around this well concentrate into it and deformed water table takes form of an inverted cone.

Permeability, K, is then represented by the following formula:

$$K = \frac{2.3Q\log_{10}\frac{a2}{a1}}{2m(h_2 - h_1)} \quad \cdots\cdots\cdots\cdots\cdots\cdots \quad (6.6)$$

where Q: volume of water pumped up in m^3/day,

a_1 and a_2: distance of the test well, B_1 and B_2, to the main well, A,

m: mean thickness of aquifer,

and h_1 and h_2: Height of water table at wells, B_1 and B_2, when pumping in the well, A, is under operation.

CHAPTER VIII. GROUNDWATER STORAGE AND ESTIMATION OF GROUNDWATER SUPPLY

§ 1. Quantity of groundwater storage

Natural groundwater storage is groundwater stored in an aquifer, unconfined or confined, as a result of recharge by infiltration and subsurface inflow and discharge by subsurface outflow.

A part of precipitation, surface water or water in cultivated land etc. infiltrates into the ground and this subsurface water is lost partly by evapotranspiration from the aeration zone before arriving to the saturation zone below. The subsurface inflow is groundwater derived by percolation from adjacent area of upper reach. Water of these two categories consists groundwater supply. The subsurface outflow is water flowing out from springs or fissures in a valley or on the slope of a hill.

Infiltration is affected by hydrometeorological conditions, that is quantity, duration and intensity of precipitation and its quality such as rain, snow and hail, and also by permeability and effective porosity of soils and rocks. Inflow and outflow is a function of permeability and hydraulic gradient of water table or piezometric surface. There are many factors influencing permeability and effective porosity; physical, chemical, geological, botanical and climatical etc.

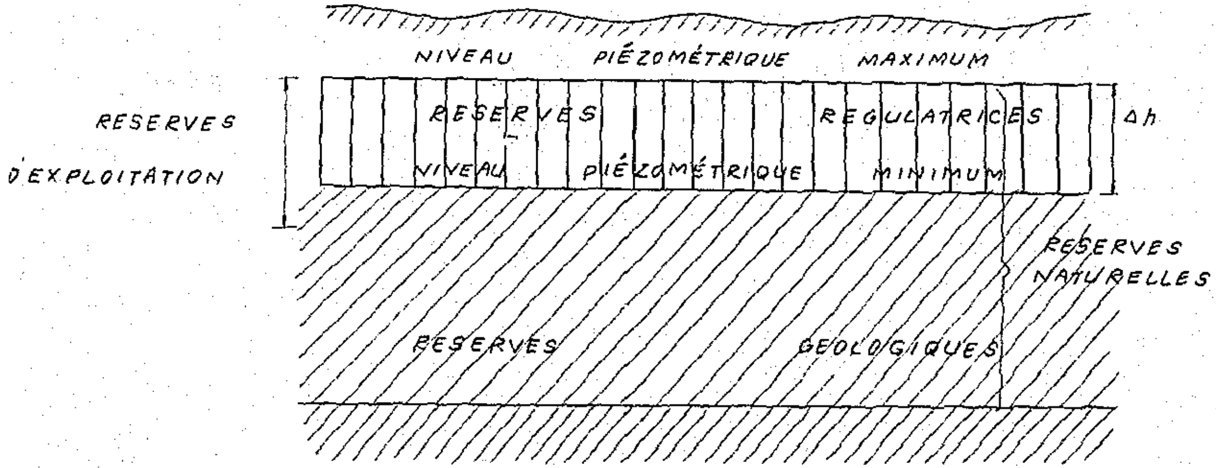


Fig. 20 Groundwater Storage in Unconfimed Aquifer

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* Schoeller, H.: Les eaux souterraines. Masson, Paris. 1962. p.319.

Quantity of groundwater storage is not kept constant, like ores in mines, but changeable according to a hydraulic equilibrium between groundwater supply and loss due to subsurface outflow, and consequently the level of the water table or piezometric surface always fluctuates.

This fluctuation is made between the highest and the lowest levels during a year. The water in this part is a regulated storage. It is this quantity that is allowable to expoitate safely during a year. The groundwater stored below this lowest level to the base of the aquifer, which is underlain by an impervious layer, is a geological storage. These two kinds of storage are unified as a natural storage.* The quantity of natural storage is determined by thickness and storage coefficient or effective porosity of the aquifer.

For estimation of change in groundwater storage it is necessary to make some measurements, as described in the following, according to the hydrogeological condition of the area concerned.

§ 2. Measurement and estimation of groundwater supply

1. Measurement of infiltration by lysimeter

The method of measurement by lysimeter was already described in Chapter VII.

2. Measurement of yield at a spring issued from the basin of one hydrogeological unit

If a spring represents a single outlet from a basin of one hydrogeological unit, there is no subsurface inflow from other basins, and yield of spring or subsurface outflow is derived only from infiltration of precipitation in the basin.

Annual infiltration I, is a function of annual precipitation, P, and is expressed by

$I = xP$, where x is the coefficient of infiltration.

If we denote U for annual infiltration depth and S for the mean effective porosity or storage coefficient in the aquifer, then

$$I = U \cdot S$$

Annual yield of spring, Y, divided by area of the basin, A, or $\frac{Y}{A}$ means the annual outflow in volume per unit area of the basin or the quantity expressed in average height in the whole area of the basin.

Thus there is a following relation among these factors:

$$h \cdot S + \frac{Y}{A} = U \cdot S = 1$$

where h · S is change in groundwater storage in the aquifer; h is the difference in elevation of groundwater level between the final level H_T and the initial level H_0 during a year, or $h = H_T - H_0$. If there is no change in groundwater storage as seen often in lime-storage and volcanic basin, $h \cdot S = 0$ and $\frac{Y}{A} = I$, or the yield of spring represents the whole amount of infiltration.

The results of measurements made at springs under the same condition as above were plotted on the diagram, taking annual precipitation, P mm, in abscissa and annual yield, I mm, in ordinate. A linear relation was obtained between them as shown in the following expressions.*

$I = 0.9\,(P - 302)$ at Oude Wetering
$I = 0.7\,(P - 210)$ at Rothamsted
$I = 1.38(P - 583)$ at Groningen

3. Estimation of groundwater supply by analysis of hydrograph**

After a rainfall or melting of snow deposit, a part of precipitation infiltrates into the ground. Infiltrated water moves downward and pore spaces between soil particles are filled with water. At the moment when the whole thickness of aeration zone above the water table is moistened and quite saturated with water, infiltrated water joins the groundwater storage. Again there exist subsurface inflow and outflow. Here we shall define outflow as quantity of water which flows away over a unit area during a unit duration of time to decrease by a certain amount of storage fixed to the groundwater level at that instant, and inflow as quantity of water which comes from adjacent area of upper reach and supplements even a small extent the loss due to outflow. According to a volume equilibrium between these three components, the groundwater storage at a certain spot in the unconfined aquifer changes in positive or negative trend and consequently the level of water table rises or falls. The sum of quantity of infiltrated water and subsurface inflow is groundwater supply, while subsurface outflow is loss.

Suppose that there exists neither groundwater supply due to infiltration or additional inflow, nor loss by extraction. The subsurface outflow causes a steady decline in elevation of the water table accompanying storage decrease. The descent is made with a certain rate, which is generally proportional to slope and permeability. A rate of descent is speedy when the level of the water table at the spot in consideration is higher, and it becomes less speedy according as the level decreases. It drops to a very small rate and fades to the minimum when slope of the water table approaches to nearly zero. The level-time relation is represented by a curve of natural descent, which is an

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*/ Castany, G.: Op. cit. p.87.

**/ Murakami, M.: La courbe de descente naturelle du niveau de l'eau souterraine et l'evaluation de la hauteur d'infiltration. Memoires de l'Association Internationale des Hydrogeologues, Tome V. Reunion d'Athenes. 1962. pp. 119-127.

exponential curve:

$$y(t) = -a + ke^{-bt} \tag{1}$$

where $y(t)$ is the groundwater level referred to ground surface at any instant t, and a and b are constants. From this formula groundwater level at t_o, or $t = 0$, is given by

$$y(t_o) = -a + k$$

and it descends in the end to the minimum level, which is given by

$$y(t_{oo}) = -a$$

This last condition is seen in dry season and the lowest record in level in the past is regarded as the approximate limit of the regulated storage.

Now the change in the groundwater storage depends on groundwater supply, outflow and extraction by pumping. An equilibrium is kept between them by following relations.

$$\begin{aligned} \Delta h \cdot S &= (\Delta I + \Delta V) + \Delta Y - \Delta\Omega \\ &= \Delta U + \Delta Y - \Delta\Omega \end{aligned} \tag{2}$$

where $h \cdot S$: change in volume per unit area in groundwater storage,
S: mean effective porosity or storage coefficient of the aquifer,
I: infiltration in volume per unit area,
V: subsurface inflow in volume per unit area,
U: groundwater supply in volume per unit area,
Y: subsurface outflow in volume per unit area,
Q: extraction by pumping in volume per unit area,

and Δ means change in value for every factor during a unit duration of time, Δt. These values in volume are represented in depth in the aquifer, or

$$\begin{aligned} \Delta h &= (\Delta i + \Delta v) + \Delta y - \Delta q \\ &= \Delta u + \Delta y - \Delta q \end{aligned} \tag{3}$$

where h: change in elevation of groundwater level,
i: infiltration in depth,
v: subsurface inflow in depth,
u: groundwater supply in depth,
y: subsurface outflow in depth or the rate of natural descent of groundwater level,
q: extraction by pumping in depth.

values in depth are quotients of their respective values in volume to effective porosity or storage coefficient, S.

Now we shall explain how to estimate graphically the quantity of groundwater supply. We shall consider for briefness a diagram for the case when the groundwater level begins to rise caused by an additional groundwater supply. Fig. 21 shows the relations between y-curve of natural descent, z-curve for fluctuation of groundwater level or hydrograph and (u-q)-curve for accumulated difference of groundwater supply and extraction. The first curve is smooth and theoretically constructed, while the other two curves are irregular in configuration and show capricious changes, reflecting behavior of groundwater supply, outflow and pumpage. Suppose that the groundwater level has decreased with the same rate as that of natural descent till the instant t_o, or that z-curve coincides with y-curve referred to the corresponding level.

At the instant t_o, when z-curve deviates from y-curve at the level $z(t_o)$, $z(t_o) = y(t_o)$, and groundwater supply is cancelled by pumpage; $du(t_o) - dq(t_{-1}) = 0$ and $u(t_o) = 0$.
The elevation of groundwater level at any instant t_n is expressed by $z(t_n)$ and it corresponds to $y_z(t_n)$ on y-curve. The rate of natural descent is obtained from formula (1):

$$dy_z(t_n) = -b \quad z(t_n) + a \quad dt \qquad \ldots\ldots\ldots\ldots\ldots\ldots \qquad (4)$$

At the instant t_1, after an interval of time Δt from t_o, the groundwater level rises to

$$z(t_1) = z(t_0) + du(t_0) + dy(t_0) - dq(t_0)$$

$$du(t_0) - dq(t_0) = z(t_1) - z(t_0) - dy(t_0)$$

$$= z(t_1) - z(t_0) +$$

$$= z(t_1) - z(t_0) + b\ \{z(t_0) + a\}\ dt$$

Here only $du(t_0)$ is an unknown factor in this equation.

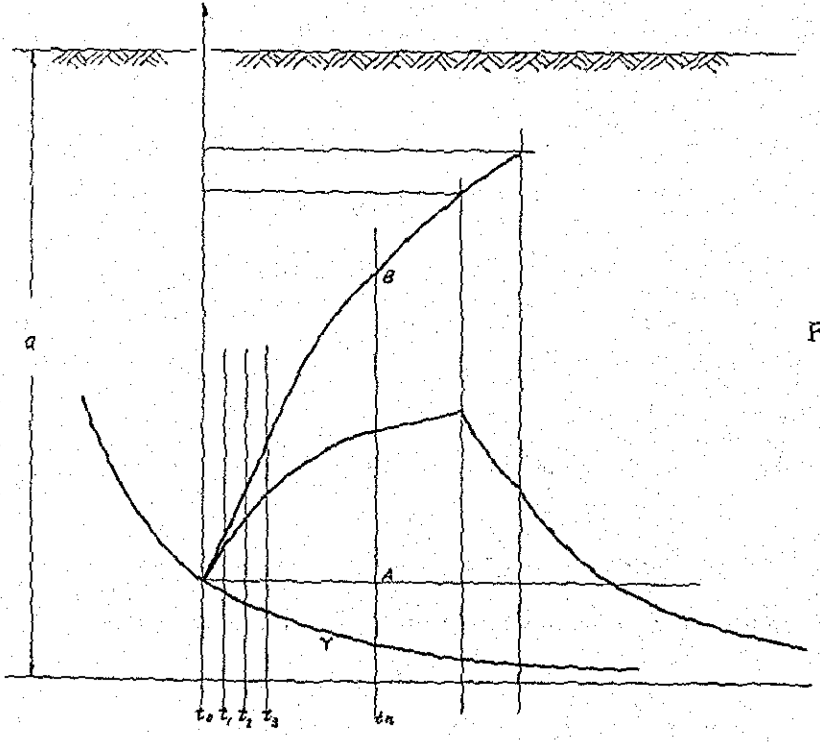


Fig. 21 Means for analyzing groundwater supply from hydrograp

At t_2, $z(t_2) = z(t_1) + du(t_1) + dy_z(t_1) - dq(t_1)$

$$du(t_1) - dq(t_1) = z(t_2) - z(t_1) - dy_z(t_1)$$

$$= z(t_2) - z(t_1) +$$

$$= z(t_2) - z(t_1) + b \quad z(t_1) + a \quad dt,$$

at t_3, $du(t_2) - dq(t_2) = z(t_3) - z(t_2) +$

$$= z(t_3) - z(t_2) + b \quad z(t_2) + a \quad dt,$$

at t_n, $du(t_{n-1}) - dq(t_{n-1}) = z(t_n) - z(t_{n-1}) +$

$$= z(t_n) - z(t_{n-1}) + b \{z(t_{n-1}) + a\} dt,$$

Summing up these equations,

$$du(t) - dq(t) = z(t_n) - z(t_0) + (\quad + \quad + \quad + \cdots\cdots + \quad)$$

$$= z(t_n) - z(t_0) + b \sum_{t}^{t_{n-1}} \quad z(t) + a \quad dt,$$

or

$$u(t_n) - q(t_n) = h_n + b \sum_{t_0}^{t_{n-1}} \{z(t) + a\} dt. \quad \cdots\cdots\cdots \tag{5}$$

The total groundwater supply $u(t_n)$ at t_n since t_0 is

$$u(t_n) = h_n + b \sum_{t_0}^{t_{n-1}} \{z(t) + a\} dt + q(t_n). \quad \cdots\cdots\cdots \tag{6}$$

In Fig. 21, $u(t_n) - q(t_n)$ is represented by $\overrightarrow{AB}$, $b \sum_{t_0}^{t-1} \{z(t) + a\} dt$ by $\overrightarrow{BC}(-)$ and $z(t_n) - z(t_0) = h_n$, or total change in elevation of groundwater level, by $\overrightarrow{AC}$, thus

$$u(t_n) - q(t_n) = \overrightarrow{AB} = \overrightarrow{AC} + \overrightarrow{CB}$$

$$u(t_n) = AB + q(t_n) \quad \cdots\cdots\cdots\cdots\cdots\cdots\cdots\cdots \tag{7}$$

$q(t_n)$, or total draft since t_0, may be obtained by calculation.

As $u(t_n) = i(t_n) + v(t_n)$,

if the groundwater supply is only from precipitation, then

$$u(t_n) = i(t_n),$$

and when there is no supply from precipitation,

$$u(t_n) = v(t_n);$$

that is only due to subsurface inflow from the adjacent area. If no draft is made in the area considered, h_n represents the natural change in the whole storage due to groundwater supply and subsurface outflow.

The total change in elevation of groundwater level during the same period is therefore given by

$$h_n = u(t_n) - b \cdot \sum_{t_0}^{t_{n-1}} \{z(t) + a\}\ dt - q(t_n). \quad \ldots\ldots \quad (8)$$

The change in volume per unit area in groundwater storage is expressed by

$$h_n \cdot S = u(t_n) - b \cdot S \sum_{t_0}^{t_{n-1}} \{z(t) + a\}\ dt - q(t_n). \quad \ldots \quad (9)$$

Calculation of groundwater supply in depth, or $u(t_n) = \sum_{t_0}^{t_{n-1}} du(t)$, is made practically as in the following processes. A *smooth curve of natural* descent of groundwater level is drawn on a tracing paper in rectangular by connecting every decreasing parts of hydrograph for periods of no groundwater supply and no draft. When the aquifer consists of several layers which are different in permeability and effective porosity, rate of natural descent is not the same for thickness of every layer. Accordingly y-curve is represented by a series of curves characteristic for every layer.

Sliding y-curve on z-curve in coordinating corresponding level height, one determines the starting point at t_0 on z-curve and reads difference in ordinate between the two curves at t_1, which represents $du(t_0) - dq(t_0) = u(t_1) - q(t_1)$, and plots $z(t_0) + u(t_1) - q(t_1)$ on ordinate for t_1. Next sliding y-curve to the position where it comes on the point $z(t_1)$ on z-curve, one reads $du(t_1) - dq(t_1)$ – position where it comes on the point $z(t_1)$ on z-curve, one reads $du(t_1) - dq(t_1)$ or difference in ordinate between the two curves at t_2, and plots $z(t_0) + \sum_{t_0}^{t_1} du(t) - \sum_{t_0}^{t_1} dq(t) = z(t_0) + u(t_2) - q(t_2)$ on ordinate for t_2. This procedure is continued till t_n to read the last increment of $du(t_{\pi-1}) - dq(t_{n-1})$ and plot $z(t_0) + \sum_{t_0}^{t_{n-1}} du(t) - \sum_{t_0}^{t_{n-1}} dq(t) = z(t_0) + u(t_n) - q(t_n)$ on ordinate for t_n. Thus we can draw (u-q)-curve. Here we can assume interval of time for calculation optionally according to inclination of z-curve, for example, as 1 or 2 hours, but it is preferable to take as shorter interval as $du(t) - dq(t)$ is larger. After completing analysis for $u(t_n) - q(t_n)$ and totalizing $q(t_n) = \sum_{t_0}^{t_{n-1}} q(t)$, we can find $u(t_n)$.

EXAMPLE – Here is an example for the case, where neither subsurface inflow nor extraction exists; made at a well in Nara plain, Japan. According to hydrograph in Fig. 22, was constructed a series of exponential curves of natural descent, composed of four segments, which are shown in Fig. 23 and represented respectively by following formulae from above to below:

A_1: $y_1 = -57.5 + 46.7\ e^{-0.770t}$ to depth 0 ~ 46.0 cm,

A_2: $y_2 = -64.0 + 44.0\ e^{-0.500t}$ for depth –46.0 ~ 56.5 cm,

A_3 : $y_3 = -123.9 + 82.4\ e^{-0.0594t}$ for depth $-56.6 \sim 100.2$ cm, and

A_4 : $y_4 = -102.3 + 2{,}870{,}000\ e^{-0.648t}$ for depth more than -100.2 cm.

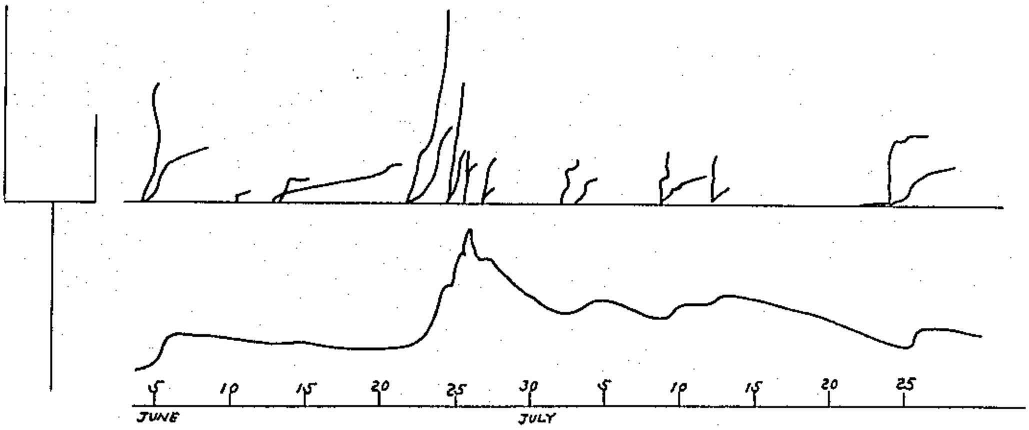
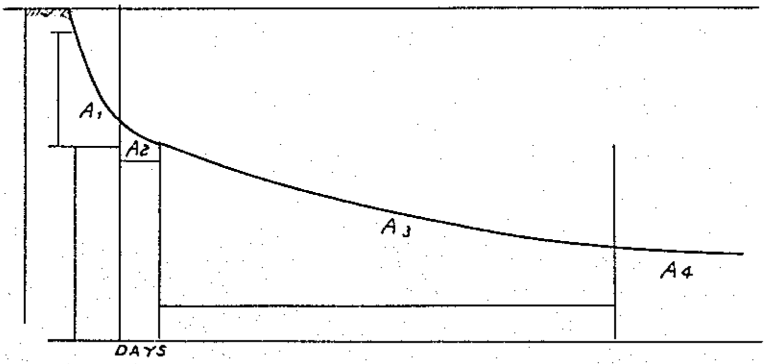


Fig. 22 Precipitation, infiltration depth



Hydrograph

Fig. 23 Curve of natural descent at Ando-well, Nara, Japan

These segments have theoretical lower limits of descent respectively at -57.7 cm (a_1), -64.0 cm (a_2), -123.9 cm (a_3) and -102.3 cm (a_4). Here the increase in groundwater storage is expected only by infiltration of precipitation. By analysis of hydrograph with aid of y-curve constructed, we calculated for every rainfall accumulated values of infiltration depth, u(t) or i(t), and precipitation, p(t).

4. Change in groundwater storage

The notation of symbols is the same as in the preceding article. The change in groundwater storage or the change in elevation of groundwater level at any instant depends on the position of z-curve referred to the horizon of $z(t_n)$ and y-curve, that is on the mutual relations in elevation of $z(t_n)$, $z(t_{n+1})$ and $z(t_n) + dv(t_n)$.

Concerning the position of $z(t_{n+1})$, as shown on the diagram in Fig. 24,

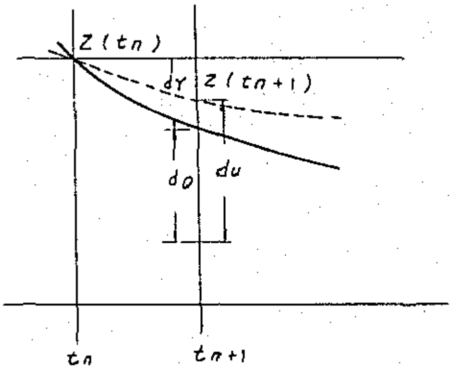


Fig. 24 Explication on condition in change of groundwater storage or elevation of level

i) z-curve is in increasing phase:

$$z(t_{n+1}) > z(t_n) \text{ or } du > -dy + dq;$$

the level and the storage is increasing.

ii) z-curve is horizontal:

$$z(t_{n+1}) = z(t_n) \text{ or } du = -dy + dq;$$

no change exists in the level and the storage.

iii) z-curve is in decreasing phase between $z(t_n)$-horizon and y-curve:

$$z(t) > z(t_{n+1}) > z(t_n) + dy(t_n) = z(t_n) - b\ \{z(t_n) + a\}\ dt$$

or

$$-dy + dq > du > dq;$$

the level and the storage are decreasing within the limit of the regulated storage.

iv) z-curve is in decreasing phase and coincides with y-curve:

$$z(t_{n+1}) = z(t_n) + dy(t_n) = z(t_n) - b\ \{z(t_n) + a\}\ dt$$

or

$$du = dq;$$

draft, dq, equals to groundwater supply, du, and the level and the storage decrease within the limit of the regulated storage. The decreasing rate of level is equal to the rate of natural descent.

v) z-curve is in decreasing phase and situates below y-curve:

$$z(t_{n+1}) < z(t_n) + dy(t_n) = z(t_n) - b\{z(t_n) + a\}\, dt$$

or

$$du < dq;$$

draft exceeds groundwater supply by overpumpage and the level and the storage continue decreasing. The fall of general level of groundwater by draft must not be overreached below the minimum level of natural descent along y-curve, or $-a$, which indicates the lowest limit of the regulated storage, otherwise an encroachment on the geological storage is induced.

CHAPTER IX. FREE AND CONFINED GROUNDWATER

PART I. FREE GROUNDWATER

§ 1. Definition of free groundwater

A part of precipitation which infiltrated under the ground through pore spaces among soil grains is reserved on impermeable rocks. The surface of groundwater which is called water table has some slope and groundwater percolates in its direction of find an outlet in a river valley, a lake or some other depression which situates in a neighboring area. This slope is decided by difference of height of water table or loss of head, distance and character of soil grains or permeability between two areas of recharge and discharge. The surface of water table has the same atmospheric pressure at the spot and the water particle has a hydraulic pressure determined by the depth from the water table. This kind of groundwater is free groundwater.

§ 2. Water table and isopiezometric lines

The form of water table is represented on an isopiezometric chart or map, in which points of the same level measured from some fixed horizontal reference plane are connected by a line which is called an isopiezometric line. It is made in the same manner as we draw a contour line on the topographic map.

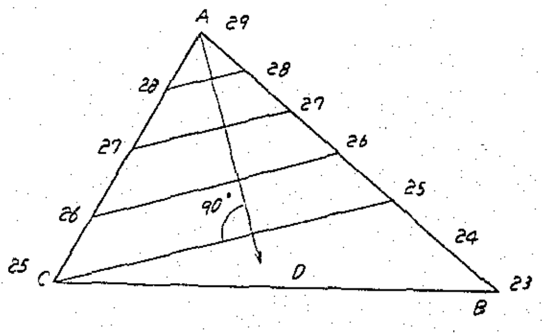


Fig. 25

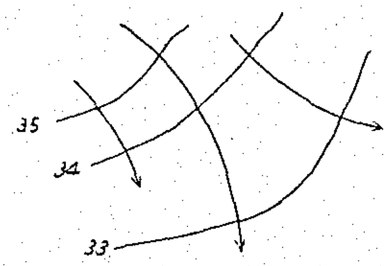
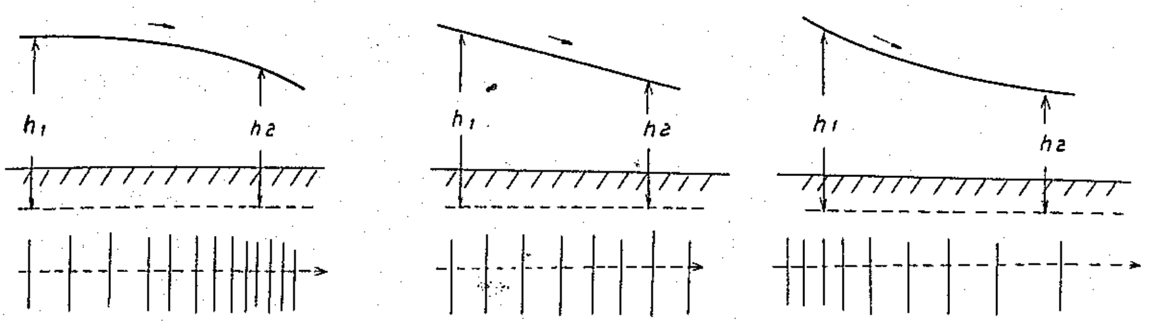


Fig. 26

The direction of flow of water tends downward perpendicular to isopiezometric lines. Water table has in general a curved surface but sometimes is flat. A vertical section including this flow line gives a type of profile of depression. In general this profile is a parabolic or hyperbolic curve, but is a straight line in special cases. When it takes a parabola, distance between isopiezometric lines diminishes. When it is a straight line, distance being constant and when it takes a hyperbola, distance increases. (Fig. 27)

i) When water table is flat, profile of depression assumes always a straight line.



Parabola Straight line Hyperbola

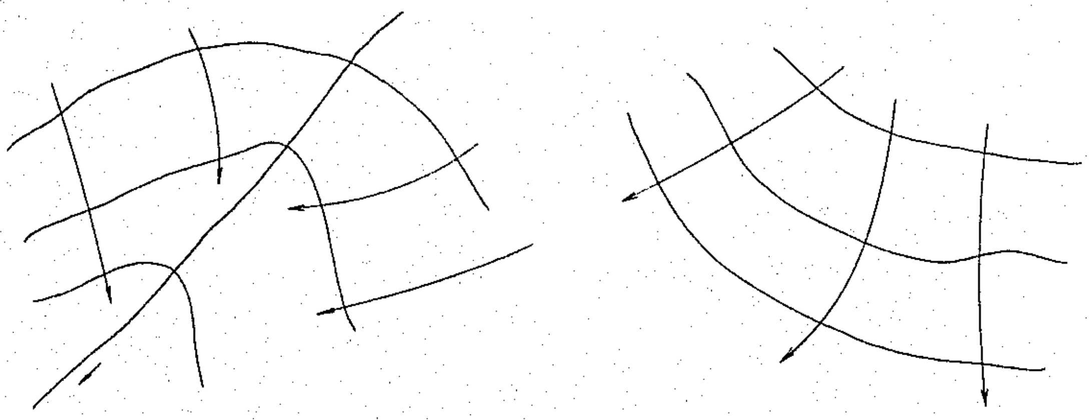
Fig. 27

ii) When water table takes a wave form with horizontal axes perpendicular to the direction of flow line, profile of depression assumes a parabola, or a hyperbola, or sometimes a straight line, and every flow line follows the same curve.

iii) When the form of water table is an inverted cone or a curved depression, the pattern of flow lines is convergent. The isopiezometric curves arrange concave toward downstream. It is the case when groundwater gathers to discharge in a river valley. (Fig. 28)

iv) When the form of water table takes a spherical or a conical domed form, the pattern of flow lines is divergent. The isopiezometric curves arrange with convex side toward downstream. It is the case when groundwater spreads downward from a recharge area with abundant quantity of water. (Fig. 28)

For both cases of iii) and iv), the profile of depression takes a parabola or a hyperbola, sometimes a straight line. These profiles are different in form for every flow line.



Convergent Divergent

Fig. 28

§ 3. Arrangement of isopiezometric lines

Distance between two neighboring isopiezometric lines along a flow line diminishes toward downstream for a parabolic form of profile of depression, is equal for a flat or a cylindrical form and increases for a hyperbolic form as are shown in the above Fig. 27.

Fig. 29 shows some examples of configuration of isopiezometric lines.

Whether groundwater flows into or flows out of a river or a lake decides direction of flow lines and accordingly the arrangement of isopiezometric line takes particular forms against the position of the river or the lake as seen in these examples.

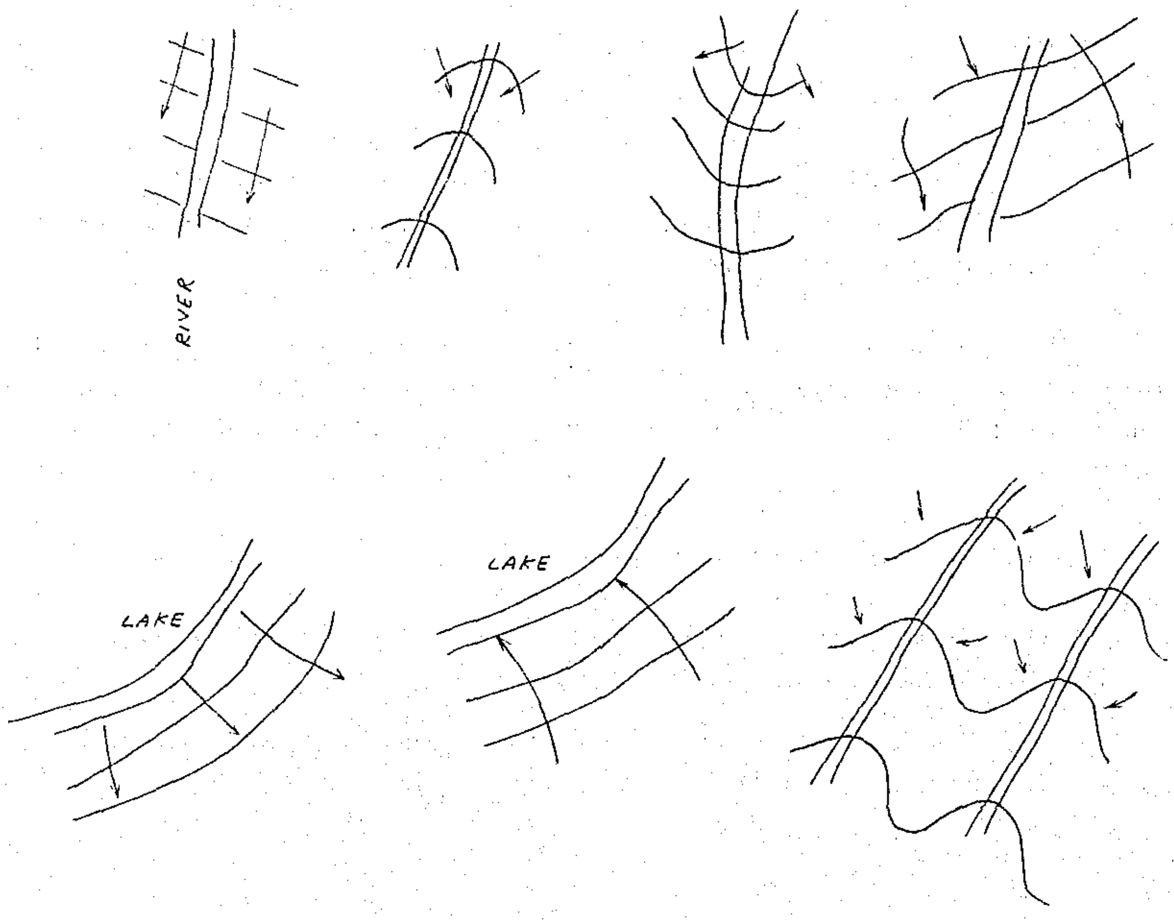


Fig. 29

§ 4. Hydraulic gradient of water table

Hydraulic gradient i, means slope of water table in direction of flow line for free groundwater. Fig. 30 shows a vertical section including flow line, or a profile of depression.

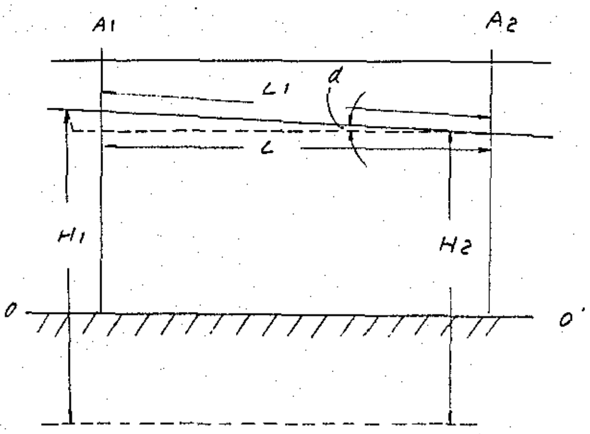


Fig. 30

Hydraulic gradient is represented by

$$i = tg = \frac{H_1 - H_2}{L}$$

where i: hydraulic gradient, H_1 and H_2: Height of water table at points A_1 and A_2 (down-stream side) measured from a fixed reference plane, $0_0'$.

L: horizontal distance between A_1 and A_2.

When angle of slope, , is very small as we see it in water table, we can put

$$i = tg = \sin = \frac{H_1 - H_2}{L'} \qquad \ldots\ldots\ldots\ldots\ldots\ldots \qquad (6.7)$$

where L': distance between two points, A_1 and A_2, along a flow line on the water table.

$H_1 - H_2$ means head loss between A_1 and A_2 on the water table.

Hydraulic gradient which we see in natural condition is from 50/1000 = 0.050 to 0.03/1000 = 0.000,03.

Hydraulic gradient is calculated referring an isopiezometric chart from a ratio of difference in height between two isopiezometric lines to horizontal distance along a flow line between them.

Where the distance between two successive isopiezometric lines is large, hydraulic gradient is small, while where the distance is small, hydraulic gradient is large. When profile of depression takes a parabolic curve, hydraulic gradient increases in direction of flow line. When it is a straight line, hydraulic gradient being constant and when it takes a hyperbola, hydraulic gradient decreases.

§ 5. Influence of discharge and permeability to form of water table

Darcy's formula: $v = Ki$ .... (6.3)

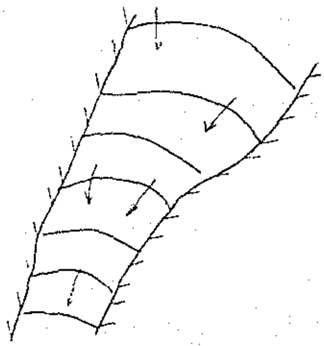
and Discharge: $Q = vS = KSi = K(HL)i$ .... (6.8)

where S is a sectional area perpendicular to flow line, and H and L are respectively thickness and width of the aquifer. According to these relations concerning four factors – Q, K, S and i, a change in value of one factor leads to mutual changes of other factors. Thus we can analyse conditions for several cases:

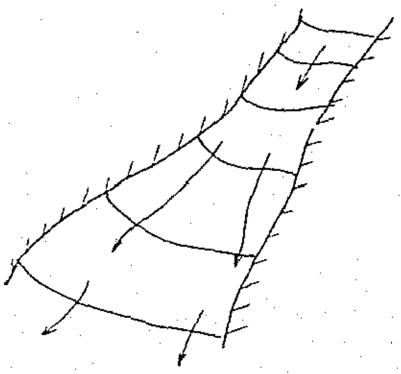
I. Q: constant and K: constant

As we see in formula (6.8), decrease of S provokes increasing velocity or increase of i; diminish of sectional area means diminish of H, L or both. This is cases of A or C of Fig. 31.

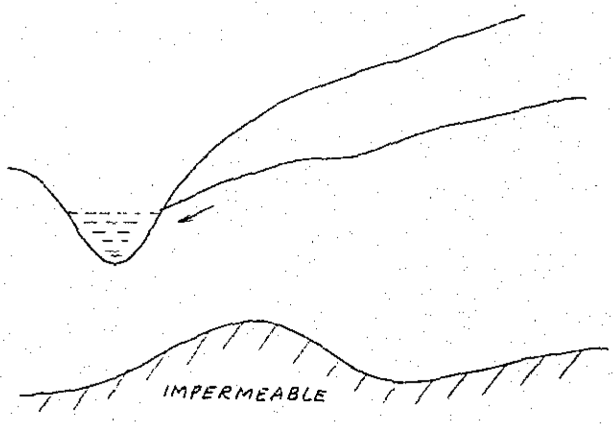
In the same manner increase of S provokes diminishing velocity or decrease of i; increase of S means increase of L, H or both, that is the case shown in B and D of Fig. 31. If there is no cause to change sectional area, v accordingly i is constant.



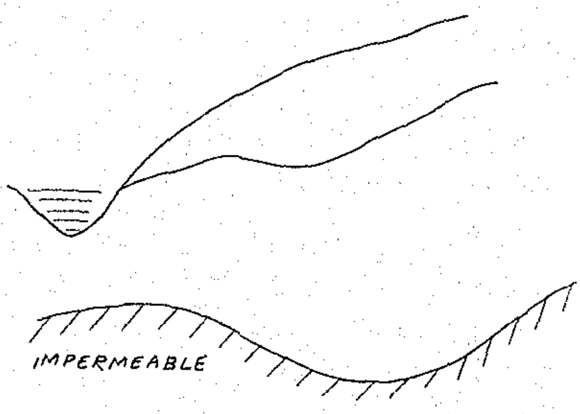
A. Convergent



B. Divergent



C. Swelling of water table



D. Depression of water table

2. **K: Constant and Q: variable (Fig. 32)**

$\frac{Q}{K}$ = Si or HLi; H, L or i may change, but in keeping a relation that products of these factors being proportional to Q. As change of i and L is very small in general, change of discharge results increase or decrease of height of water table.

3. **Q: constant and K: variable (Fig. 33)**

$\frac{Q}{K}$ = Si or HLi; H, L or i may change, but in keeping a condition that product of these three factors is inversely proportional to Q. As change of i and L is very small in general, decrease of permeability toward downstream makes height of water table lower, while increase of permeability raises its height.

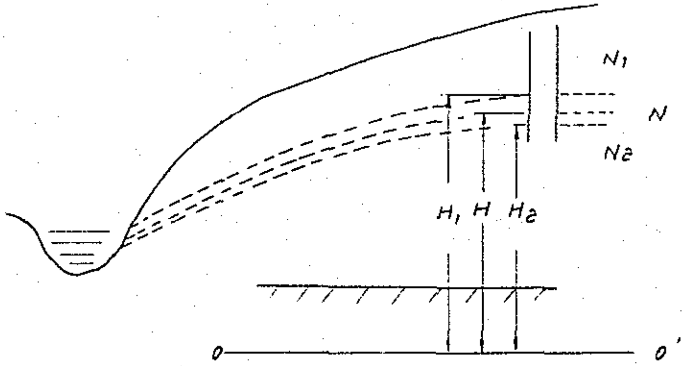


Fig. 32

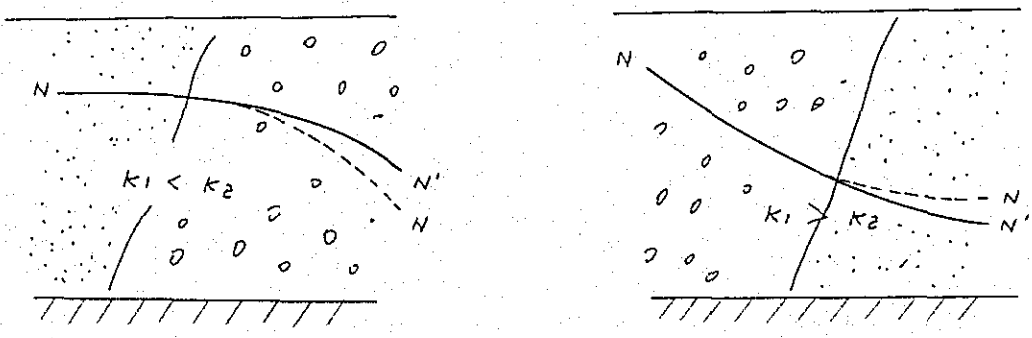


Fig. 33

By change of permeability from K_1 to K_2, profile of depression is modified as from NN to NN'.

§6. Estimation of discharge of groundwater

1. Apparent velocity: v

$$Q = KSi$$

Q is discharge of groundwater which passes the sectional area, S. It is convenient to calculate discharge, q, for sectional area of unit length of width or 1 meter width.

$$q = KHi \qquad \dots\dots \qquad (6.9)$$

where H is thickness of groundwater body, and q is called transmissibility.

EXAMPLE $K = 3 \times 10^{-2}$ m/s

$H = 12$m; $i = 0.01$,

$q = 3 \times 10^{-2} \times 12 \times 0.01 = 0.0036$ m^3/s.

It is an example of apparent velocity which is

$$v = Ki$$

or $v = 3 \times 10^{-2} \times 0.01 = 3 \times 10^{-4}$ m/s.

If we take $S = 240m^2$ or sectional area with 20m width,

we obtain $Q = 240 \times 3 \times 10^{-4}$ m^3/s $= 0.072$ m^3/s

or $= 0.072 \times 3600 = 259$ m^3/h.

2. Real mean velocity: v_m

When we consider that groundwater flows through pore spaces, which is represented by total porosity, we call it real mean velocity. Then the relation of real mean velocity to apparent velocity is expressed as

$$v_m = \frac{Q}{S_m} \qquad \dots\dots \qquad (6.10)$$

where v_m *and* S_m *are respectively real mean velocity and sectional area referred to total* porosity. So the relation between v_m and v is expressed as

$$v_m : v = S : S_m \qquad \dots\dots \qquad (6.11)$$

3. Real effective velocity: v_e

We can know velocity of groundwater by colorimetric method in measuring time to pass from a station in the upper reach to the another station in the lower reach along the stream line. It is true velocity of water which flowed through spaces represented by effective porosity and is called effective velocity, v_e.

For the same discharge of groundwater as above cited,

$$Q = 259 \text{ m}^3/\text{h} = 0.072 \text{ m}^3/\text{sec.}$$

$$S = 240 \text{ m}^2$$

and when we take effective porosity $m_e = 10\%$, we get as effective porosity

$$v_e = \frac{Q}{Sm_e} = \frac{v}{m_e} = \frac{3 \times 10^{-4}}{0.1} \text{ m/s} = 3 \times 10^{-3} \text{ m/s}$$

and $$v_e : v = S : S_e \quad \cdots\cdots\cdots\cdots \quad (6.12)$$

It is ten times speedy than the apparent velocity.

We cite here data of real mean velocity and effective velocity expressed both in m/day for the hydraulic gradient of 1% for different material of aquifers.

Table

Material of aquifer	Diameter of grains (mm)	Ream mean velocity (m/day, 1%)	Effective velocity (m/day, 1%)
Clay	0.005 ~ 0.25	0 0 ~ 0.0003	0 ~ 0.0012
Silt, fine sand, loess	*0.25 ~ 0.5*	*0.013 ~ 0.02*	0.02
Sand or sandstone *with middle grains*	0.25 ~ 0.5	0.30	0.35
Coarse sand, sandy gravel	0.5 ~ 0.2	2	about 2
Middle gravel	2 ~ 5	7.5	9
Coarse gravel	5	33	
Maximum velocity in gravel	74% of grains being 5 mm		

PART II. CONFINED GROUNDWATER

§7. Definition of confined groundwater

When water flows through an aquifer which is situated between two layers of aquifuge above and below, this kind of groundwater is called confined groundwater. That is when an impermeable layer above prevents see page of infiltrating water directly and groundwater at any depth receives hydrostatic pressure due to weight of water column above that depth.

When such aquifer is cut by some fault and water leaks out through this fissure or a hole is drilled to this aquifer through an aquifuge or an impermeable layer superposed above, water goes upward through this fissure or the bored hole to a height corresponding to hydrostatic pressure in the aquifer. Water springs with pressure if this height is above the ground surface. We call such water artesian water. The well through which artesian water flows out is called artesian well.

§8. Some examples of confined groundwater

We show here some examples of confined groundwater.

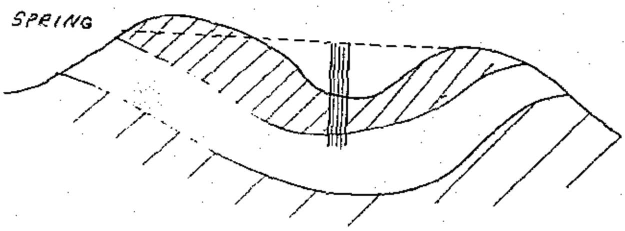


Fig. 34

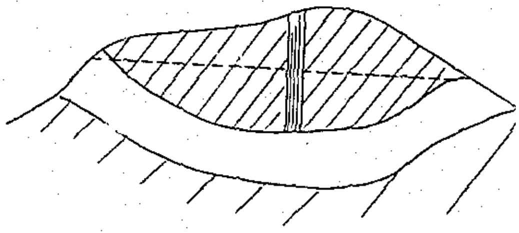


Fig. 35

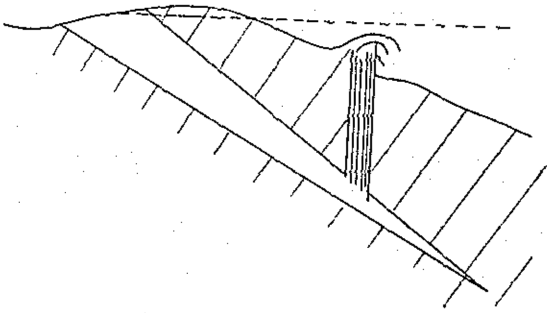


Fig. 36

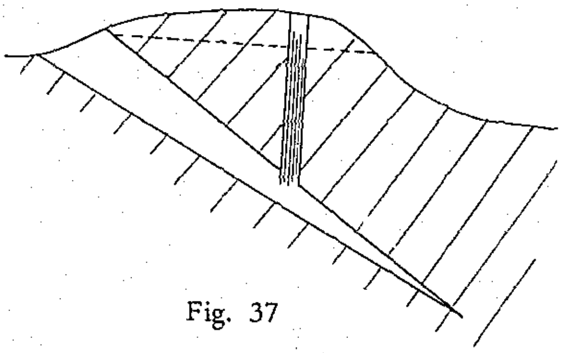


Fig. 37

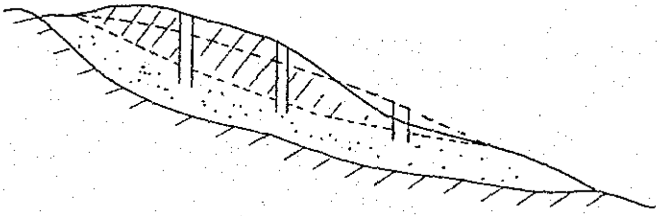


Fig. 38

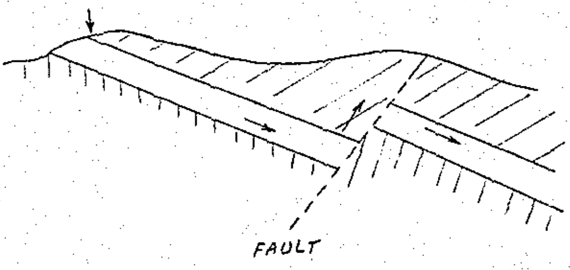


Fig. 39

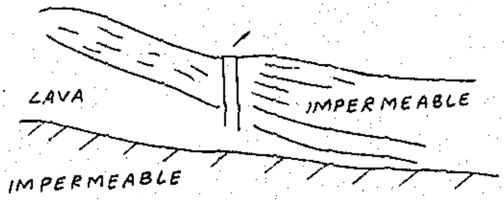


Fig. 40

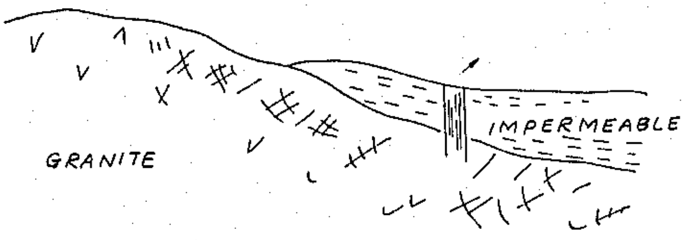


Fig. 41

g is eroded granite and permeable.

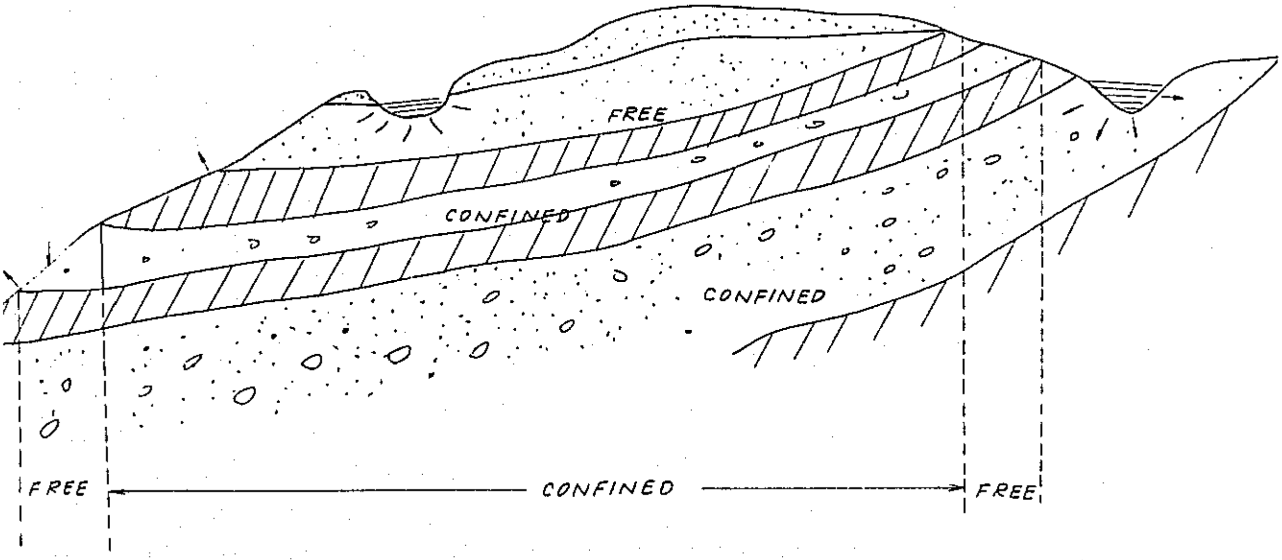


Fig. 42

§9. Water discharge

We will calculate now discharge of confined water for a simple case. When aquifer is confined with water between two parallel strata, water flows parallel to these layers and hydraulic gradient is the same as inclination of strata.

Darcy's law is applied also in this case.

Apparent velocity is $v = ki$

and discharge of groundwater is expressed by

$$Q = SKi$$

where S is a cross section perpendicular to flow lines with thickness H and width L, or

$$S = HL$$
$$Q = KHiL$$

Sometimes value of $KH = T$ (6.13) is used and is called transmissibility, or $Q = TiL$ (6.14)

Transmissibility has a dimension $[L^2T^{-1}]$ or is expressed in m^2/s or cm^2/s.

For example, when $T = 2\ m^2/s$, $i = 0.0038$, and $L = 11700\ m$,

$$Q = 2 \times 0.0038 \times 11700$$
$$= 88.92\ m^2/s$$

CHAPTER X. EXPLOITATION OF GROUNDWATER STORAGE

§ 1. Climatic environments

In wet regions we can find easily aquifers with sufficient yields where geological conditions is favorable for storing water in rocks. Groundwater storage depends on the behavior of precipitation and its seasonal variation in quantity.

While there are particular problems in regions of extreme climates. There are arid and semi-arid regions where the loss of surface and groundwater is much abundant due to small quantity of precipitation as well as strong evapotranspiration. In the desert region existence of the surface water is limited and the groundwater is stored in natural cavity or in wells and drains artificially devised as we see in long infiltration galleries, or kanats, in Iran. The River Nile in Egypt is the water flowing from other basins where is abundant of precipitation, and the groundwater there is found only in Nubian sandstones of which outcrops are exposed on ground surface in wet region. In the semi-arid region precipitation occurs seasonally and we see consequently temporary streams on surface of the land.

In dry area the water once infiltrated in soil or outflowed through fissures in rocks from adjacent areas is brought by capillary action to ground surface, where salts deposit by strong evaporation and crust is formed. Thus the groundwater is much brackish or saline in quality.

Chemical analysis of water and total dissolved solids in arid and semi-arid regions are shown in Fig. 1* and the following table.

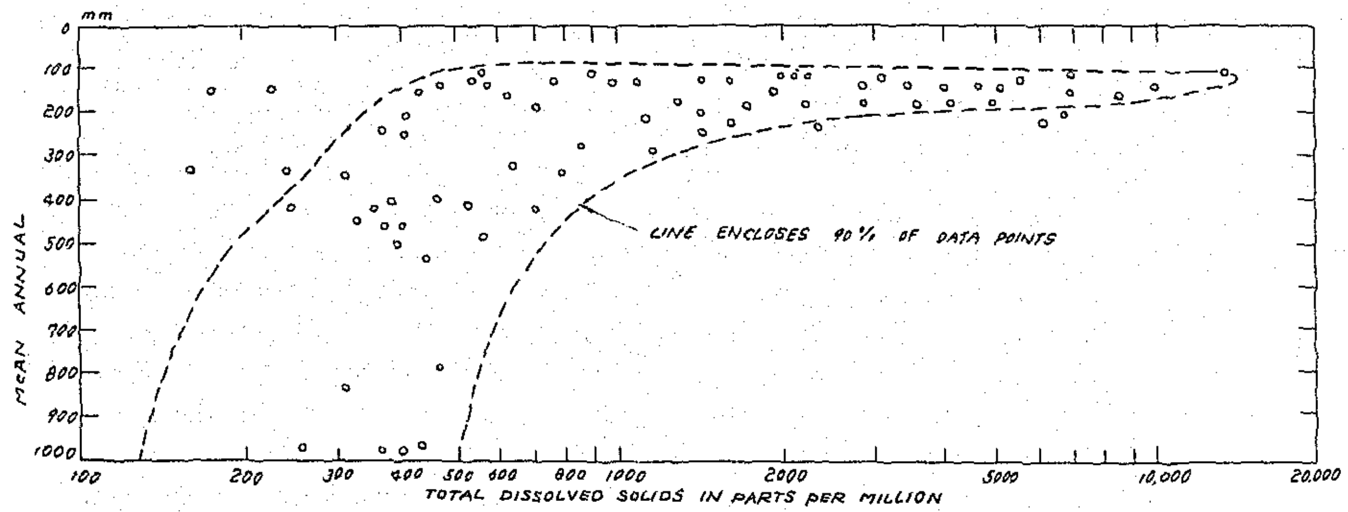


Fig. 43 Relationship between mean annual precipitation and total dissolved solids of groundwater in Syria

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* Davis, S.N. & DeWiest, R.J.M.: Hydrogeology, John Wuley, New York, 1966.

Table

Chemical Analyses of Water from Arid and Semi-arid Regions, Analyses in ppm

	Water Leached from Saline Soil of the U.S.S.R.	Fresh Well Water from Alluvium, Northern Chile	Brackish Well Water from Alluvium, Northern Chile	Brackish Spring Water from Rocks Rich in Gypsum and Calcite, New Mexico	Fresh Spring Water from Rhyolite Aquifer near Socorro, New Mexico
SiO_2	...	36	4	21	27
Ca	1540	7	295	456	18
Mg	360	0.5	58	113	4
Na	2010	160	474	47	54
K		8	61	...	2.8
HCO_3	61	248	63	200	154
SO_4	2430	96	415	1460	28
Cl	5040	53	1130	15	15
NO_3	...	4	0.3	0.2	1.2
B	...	...	18	...	...
Total disoolved solids	...	487	2469	...	224
References	(26)	(8)	(8)	(15)	(15)

Water from spring of Ein Nur (Tannur)* shows chlorine contents of more than 2000 mg/ℓ. The lowest dissolved solids are generally found in small quantities in fractured metamorphic and plutonic rocks and in sediments derived from these rocks. We can also find favorable aquifers at the basal zones in very large sand dune tracts and in basalt and limestone regions. Aquatic vegetations indicate wet condition in soil and we see them in wadis.

There are regions of cold and polar climates, where water freezes during a long period or permanently, although evaporation loss is small owing to low temperature. In the cold regions within the Arctic Circle in U.S.A., Canada and U.R.S.S., soil water in the ground freezes and permafrost is formed. Fig. 2 shows the distribution of area of permafrost. It extends to latitude 48° in both central Asia and central North America. In the north region permafrost is generally continuous, but as one passes to the south distribution of permafrost is not continuous and then becomes sporadic. It depends on the ground temperature distribution through the year. The ice melts when temperature rises higher than 0°C and the water infiltrates into the ground. To look for potable

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* Government of Israel, Ministry of Agriculture, Water Commission, Hydrological Service; Hydrological Year-Book of Israel, 1962/63. Jerusalem, 1965.

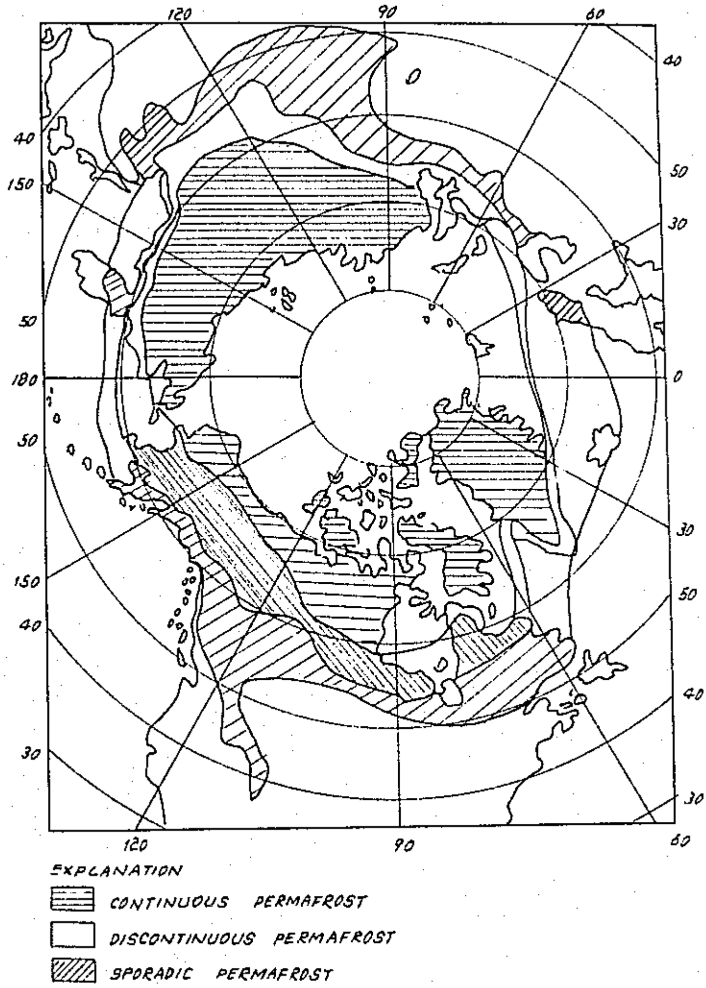

Fig. 44 Distribution of permafrost in the northern hemisphere

water in the Arctic permafrost region is to make a search for permeable and unfrozen zones that are saturated with water. Vegetations with long roots like willows indicate the place where permafrost is absent. Polygonal patterns such as frost-crack polygons and stone polygons on the ground surface indicate commonly formation of permafrost. Formation of hydrolaccoliths or pingos shows the accumulation of ice lenses under the ground. We must also mark change in discharge of rivers or lakes which suggests active infiltration of water into the ground, existence of springs, patches of luxuriant vegetation and other indications for non-frozen soils.

§2. Geological environments

The aquifers favorable for storing groundwater is classified as in the following.

i) Unconsolidated sediments:
Gravel, sandy gravel, sand-quaternary and pliocene beds.

ii) Semiconsolidated and consolidated sediments:
Conglomerate, sandstone with joints and cracks or fractured-tertiary and older rocks.

iii) Limestone and dolomite with joints and other fractures.

iv) Volcanic rocks as basalts, tuffs which are porous or with cracks. The characteristics of aquifers depends on porosity and permeability given by intergranular pores and accordingly is influenced by arrangement of composing materials, existence of joints and cracks or fissures in rocks.

Here are some examples of measurements of transmissivity and storage coefficient which suggest permeability, porosity or storage capacity of aquifers.

Table*

	Thickness	Transmissivity	Storage coefficient
Alluvial fan deposits, Gallatin Valley, Mont.	63	36,000	0.06
Glacio-fluvistile deposits, Hanford, Wash.	45	3,000,000	0.20
Aquia greensand	20	10,000 ~ 20,000	0.00023
Patapsco formation	25	35,600 (av.)	0.0002 (av.)
Silurian dolomite	250	61,000	0.00035

It is difficult to find groundwater in the aquicludes which are represented by rocks as described in the following, unless there is fissures or cracks in which groundwater is reserved: crystalline, igneous and metamorphic rocks, relatively impermeable carbonate rocks, consolidated clastic rocks and fine-grained clastic rocks such as shales and clays.

For selecting site of drilling we must refer to geological construction and hydrogeological conditions and topography in the area.

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*/Maxey, G.B: Geology, part I. Hydrogeology. Nadbook of Applied Hydrology edited by Chow, V.T. McGraw-Hill, New York. 1964. pp. 4.

§3. Water uses

Groundwater is used for irrigation, industry and public and domestic water supplies.

In the United States of America the rate of use of groundwater in 1955 showed 16% of the total national water use.* This usage amounted approximately to 50 billion gallons per day in *1960, which equals about 20% of the total water use.* A demand for groundwater is increasing every year and the drilling of new wells amounts to 400,000 annually.**

Distribution of Groundwater Use

	Percent of total groundwater use
Irrigation	65
Self-supplied industrial water supplies	21
Public water supplies	10
Rural water supplies	4
Total	100

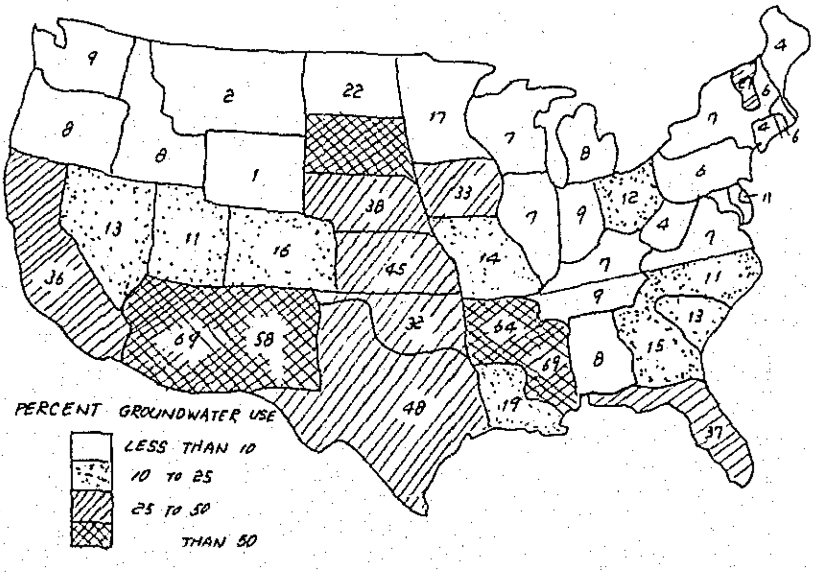

Fig. 45 Groundwater use relative to total water use in the United States (After MacKichan)

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*/United Nations, Department of Economic and Social Affairs: Water for Industrial Use. New York, 1958, p.20.

**/Todd, D.K.: Groundwater. Handbook of Applied Hydrology edited by Chow, V.T. McGraw-Hill, New York. 1964. p.13 - 5.

In Finland, of the total amount of water used in 1950, 3,051,000 m^3/day, groundwater occupied 178,800 m^3/day compared with surface water 2,872,200 m^3/day or 60% of the total.

The yearly amounts of groundwater for industrial use in Japan are shown in the following data from Ministry of International Trade and Industry.

Year	Total amount	Percentage for total amount of fresh water used
1955	3,838,000 m^3	39.2%
1958	7,836,000	32.8
1962	11,092,000	29.7
1964	14,207,000	29.7

Thus the need of groundwater is increasing every year and the amount of groundwater for industrial us in 1971 in Japan is estimated 18,704,000 m^3, or 23.2% of the total *amount of fresh water to be used, 121,000,000* m^3.

The statistics for Tokyo municipal area shows that the yearly amounts of groundwater for industrial use in 1964 was 555,000 m^3/day or 39.1% of the total fresh water 1,420,000 m^3/day.*

§4. Types of reservoirs

There are two types of reservoirs: confined and unconfined.

Unconfined water is defined as the water that is indirect contact vertically with the atmosphere through openspaces in permeable material.

Confined water is reserved in the aquifer which is separated from the atmosphere by impermeable bed. Confined water is also called artesian water. When the pressure surface of confined water or piezometric surface is at a higher level than the level of the position where well is drilled, groundwater flows up to a height of piezometric surface.

The surface of unconfined or free surface groundwater is called water table and it takes a form *similar to the topographic surface*, but *the piezometric surface assumes* a plane which connects elevations of the position where water infiltrates into the aquifer and the outlet or spring in a valley. Outflow from the artesian basin concerns the level of water table at the inlet of the aquifer. By connecting the elevation of piezometric surface at every well, we can find the position of inlet where, for example, river water infiltrates in.to the aquifer.

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* Suiri-kagaku Kenkyusho (Water Science Institute): Annual Report of Water Economy, 1960 (in Japanese). Tokyo. 1959.

CHAPTER XI. UTILIZATION OF GROUNDWATER STORAGE

§ 1. Conservation of water in quantity and quality

Groundwater must always be conserved in quantity and quality. One should pay attention not to destroy a hydrologic equilibrium in groundwater storage and to prevent from deterioration of quality. So that the supply may be continued indefinitely without any harm. It is necessary to control water uses and to supervise construction work of pumping installation and treatment of waste water by legislation.

§ 2. Conservation of water quantity

We can not extract by pumping all the natural storage of groundwater. It is only a part of this storage that is allowable to exploit safely, and it is called exploitable storage. If the extraction of groundwater more than the explitable storage is continued every year, by some economical reasons, a decrease of groundwater level results. This phenomenon means an overdraft, of withdrawal in excess of groundwater from the geological storage, and quantity of water in the groundwater storage would be consumed at the end as seen in mining of ores.

To prevent this danger and conserve the quantity of groundwater in optimum condition, it is necessary to maintain the water resources indefinitely, and it is required for this aim to estimate "safe yield" or the quantity of exploitable resources, and to take care not to exceed this limit, so that a hydrological equilibrium between all waters entering and leaving the basin may not be destroyed. Safe yield defines the basin draft taken annually on a groundwater supply which can be continued indefinitely without harming the supply.* Here groundwater supply means the quantity of water which adds to groundwater storage by infiltration (after evapotranspiration loss) plus subsurface inflow.

Whether level of water table rises or falls due to withdrawal of groundwater depends on the quantitative relation between water supply, subsurface outflow and draft. If we designate daily groundwater supply ΔV (m^3/day), subsurface outflow ΔY (m^3/day), daily draft ΔQ (m^3/day) and daily change in elevation of groundwater level Δh (cm/day), then their annually totalized amounts are indicated respectively by V m^3, Y m^3, Q m^3 and h cm. Considering a hydrological equilibrium in a basin between them,

$$\Delta V - \Delta Y - \Delta Q \infty \Delta h$$

or

$$(V - Y) - Q = h \cdot S$$

where S means effective porosity or storage coefficient of the aquifer. Totalizing their values during a year.

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* Todd. D.K.: Op. cit p.200.

This relation is represented by a straight line with a slope fixed

$$(V - Y) - Q = h \cdot S$$

$$Q = (V - Y) - h \cdot S$$

by S in coordinates of ΔQ and Δh. Whereas, so far as $V - Y$ is unknown, we can not *exactly correlate* Q and h. *But if* $V - Y$ *is* nearly constant every year, then it is expected that plots of Q and h would be arranged on a fixed straight *line. This relation* is explained graphically, assuming Q in abscissa and h in ordinate, as shown in Fig. 46.

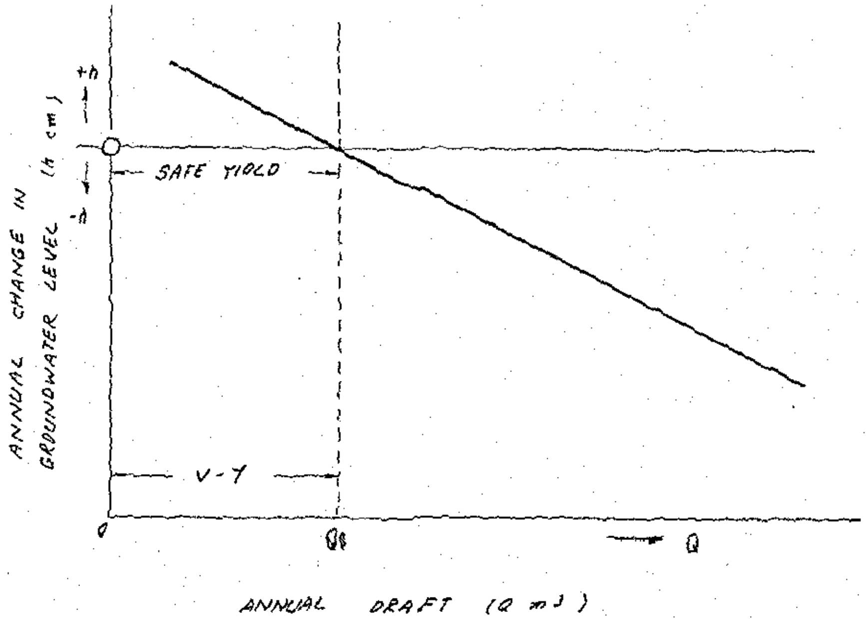

Fig. 46 Relation between annual draft and annual change in groundwater level

If the draft, say Q_0, is equal to quantity $V - Y$, then $h = 0$ or the hydrologic equilibrium is maintained and any change is recognized neither in elevation of groundwater level *nor in* groundwater storage. The groundwater level tends to decrease, or h is negative, when the draft, Q, *exceeds* $V - Y$, *and to* increase, or h is positive, on the contrary, when Q is kept within $V - X$. In general it is very difficult really to find exact value of $V - Y$, unless in special cases or any appropriate research is not made for estimating it. So we must then assume $V - Y$ being nearly constant.

The safe yield is *given by an intercept from the* origin to the straight line for $h = 0$. *Thus it is* possible to foresee a danger of overdraft and prevent it by *controlling* yield not to surpass the safe *yield.* To *determine* the safe yield in a new basin where no development of groundwater has ever been made, we can find it by *measurement* of Q and h for a shorter period.

R.H. Hill's method to determine safe yield is based on this principle.

The diagram for his research for the Pasadena Basin, California, U.S.A. *is shown in* Fig. 47.*

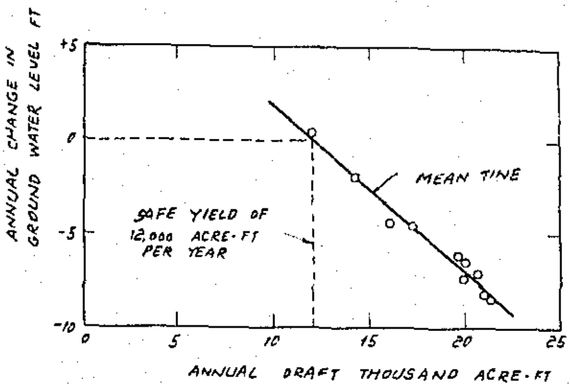

Fig. 47 Determination of safe yield by the Hill Method for the Pasadena Basin, Los Angeles Country, Calif., U.S.A. (after Conkling)

He plotted data as five-year moving averages for the period *1922 ~ 1938.* The safe yield was indicated as 12,000 acre-ft per year, while the average annual pumpage was amounted to 18,500 acre-ft, which caused a sizable overdraft in the basin.

The over-extraction of groundwater causes not only lowering in elevation of level of water table or piezometric surface, but also land subsidence. It occurs due to shrinkage of clayey layers by exhaust of water in sandy aquifers adjacent above and below. Serious damages are provoked by it causing stagnation of surface and ground water, inondation of floods, and especially in coastal area invasion of high tides.

Here is an example of land subsidence in Osaka City, Japan, caused by the overdraft groundwater storage.**

Subsidence was remarked since 1925 or before and it was accelerated abruptly after the war since about 1950 according as the need of water for industrial use began to increase. The subsidence continued till 1964 in most places. The maximum record of subsidence was observed 251.7 cm during 30 years covering 1935 ~ 1965 in the severest damaged center.

Researches revealed this cause being overdraft of groundwater storage, and for prevention of this desaster a severe control was executed by law on draft and pumping devices since 1959; the uppermost position for strainer cut was restricted below 500 ~ 600m and sectional area of pipe at outlet of a pump within 21cm^2. Pumpage of water

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*/Todd, D.K.: Op. cit. pp.206 ~ 207.

**/Commission for General Countermeasure against Land Subsidence: Outline of Land Subsidence in Osaka (*in Japanese*). 1966.

Department of General Planning, Osaka City: Data for Land Subsidence in Osaka (in Japanese). 1966.

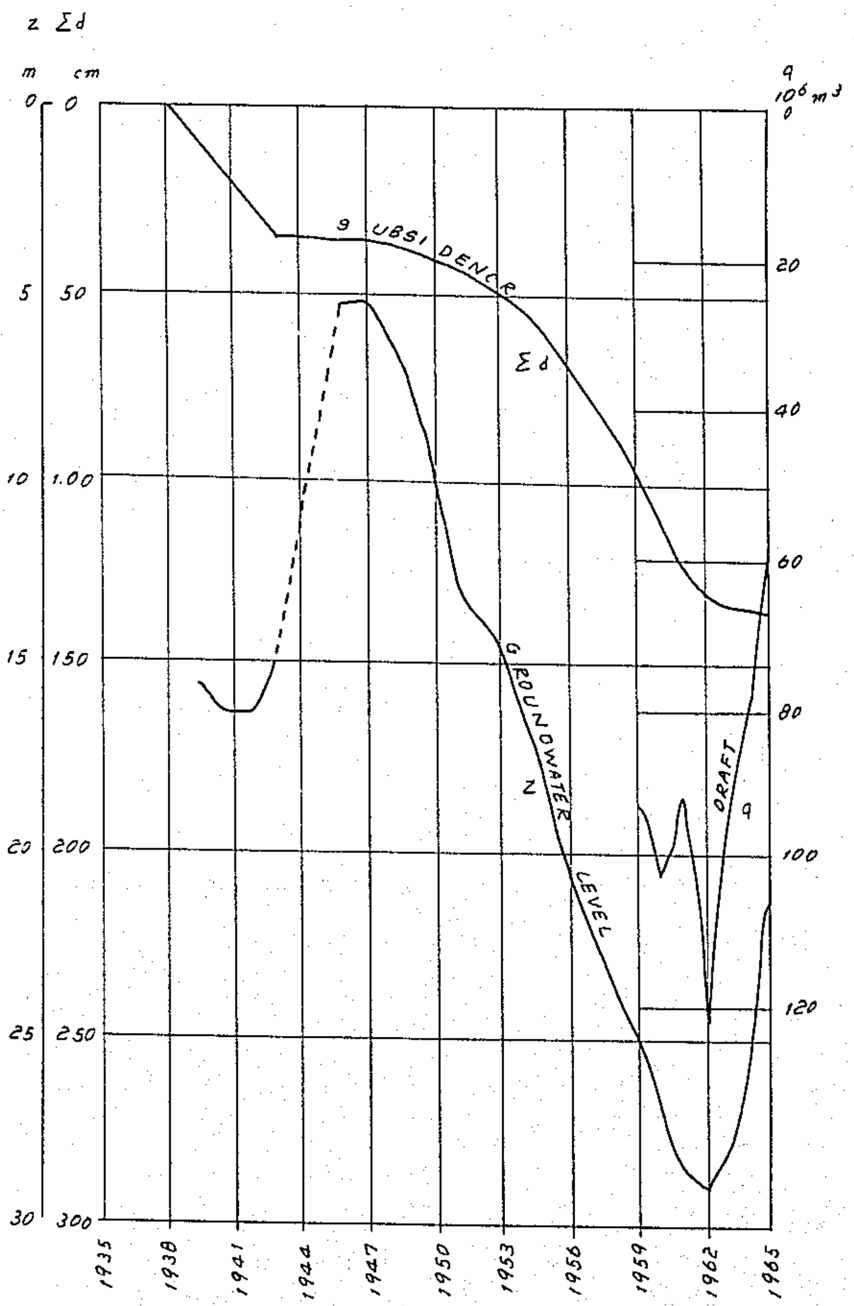

Fig. 48 Observation records for No.3 well at Kujo, Osaka

for cooling use in buildings was ceased and wells not fitted for the categories above mentioned were abandoned. On the other hand, the water works specially planned for industrial water was started.

By this control the draft diminished, the decrease of groundwater level stopped in 1962, and the rate of subsidence dropped and then changed in reverse direction since 1964 for most wells in Osaka. Observation records for one of wells are shown on diagram in Fig. 3.

§3. Conservation of water quality

The groundwater storaged in the ground is exploited and used. Water is wasted or reused after treatment by various processes, and if such water infiltrates into the ground through reservoirs or drainage channels, the groundwater is by and by polluted.

We find very often in groundwater bacteria whose existence proves a contamination. They are brought from polluted soils at ground surface when water infiltrates. Valleys and lakes or reservoirs, where installations for intake of drinking water are situated, are requested to be safeguarded against pollution by waste water and sewage.

The groundwater is used for drinking, industry and irrigation. The criteria for allowable contents of dissolved elements and hardness is different according to objects of water utilization. The measures of chemical, physical and bacterial constituents must be specified, as well as standard methods for reporting results of water analyses. Protection from pollution and conservation of groundwater quality is essential for utilization and development of groundwater. If groundwater of inferior quality is stored in an adjacent area, it is probable to withdraw it in consequence of lowering of water table by pumpage. So draft must be controlled by appropriate processes in like manner as devised against sea water intrusion.

1. Protection from sea water intrusion

Fig. 4 shows the chemical analyses of waters in wells arranged on a line extending from the center of the aquifer to the coast in Salinas Valley, California, U.S.A.*

Chemical analyses were represented on Collins' diagram and chloride-bicarbonate ratio, $\frac{Cl}{CO_3 + HCO_3}$, was calculated for each well. The total concentration and chloride-bicarbonate ratio increases from the well inland toward the well near the coast and the ratio approaches by and by to that for sea water.

This phenomenon is an example of sea water intrusion about which shall be described in the following.

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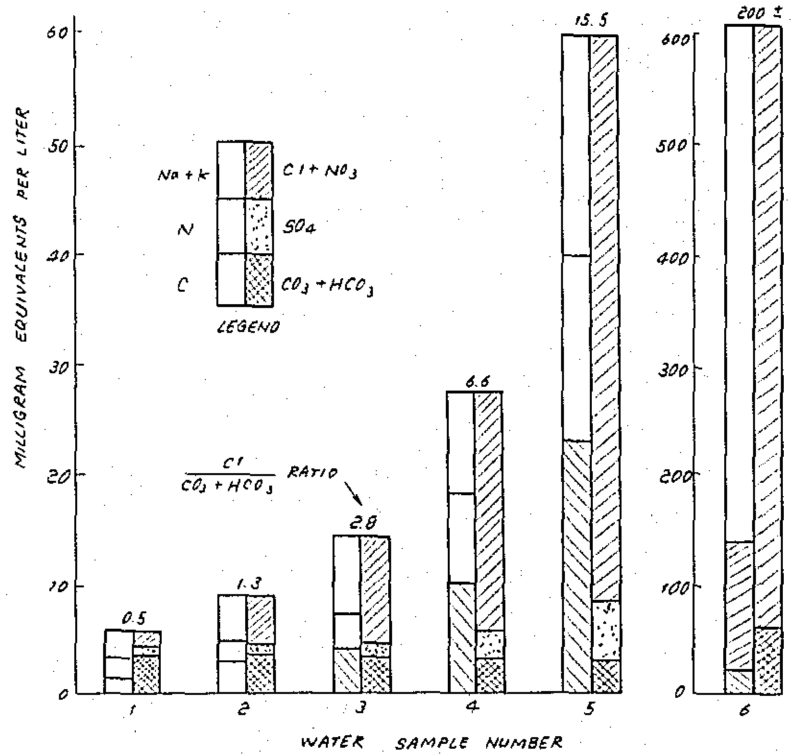
*/Todd, D.K., Op. cit. p.293.



Sample Number	Description
1	Normal good groundwater in aquifer
2	Slightly contaminated groundwater
3	Moderately contaminated groundwater
4	Injuriously contaminated groundwater
5	Highly contaminated groundwater (near seashore)
6	Sea water

Fig. 49 Chemical analyses of a line of well waters from Salinas Valley, Calif., U.S.A.

Along the coast fresh groundwater flows out into the sea, while sea water seeps inlandward through the sea bed, and is formed in the aquifer an interface between fresh water body above and sea water body below. If groundwater is pumped up from below the level of this interface, brine is extracted. While even if pumpage is made from the depth above this interface, an overdraft would cause sea water intrusion. So care must be taken not to break an equilibrium between fresh and salt water by overdraft.

Five methods for prevention are introduced by Todd to prevent this phenomenon.*

i) To change pattern of wells for a confined aquifer by shifting their concentrated position from near sea shore to inland. By this method salt-water wedge is diminished in size and sea water intrusion is prevented.

ii) To lower the level of the interface by replenishment of fresh water. This procedure is made by artificially recharging water into the intruded aquifer, through spreading areas for unconfined aquifers, and by recharge wells for confined aquifers.

iii) To construct a line of wells near and parallel to a coast, and form a trough line in the groundwater level. By this procedure it is possible to withdraw sea water and prevent a flow of fresh water toward the coast by stable salt-water wedge formed inland-side along a line of wells.

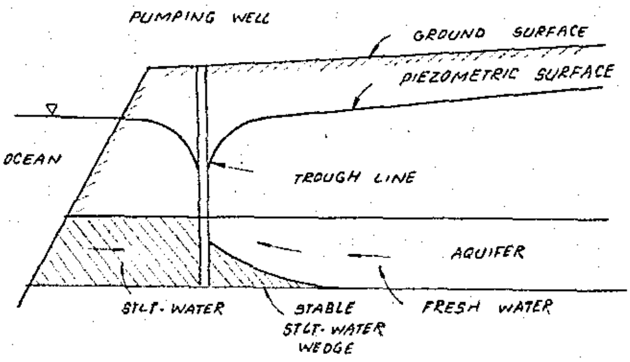


Fig. 50 Control of sea water intrusion by a pumping trough paralleling the coast

iv) Inversely as the previous method, to form a fresh-water pressure ridge line by surface spreading of water for unconfined aquifers, or constructing a line of recharge wells adjacent and parallel to the coast for confined aquifers. A ridge must be of sufficient height above sea level to repel sea water. Water table or piezometric surface would assume a cone like form around each well and each cone is connected with saddle form.

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*/Todd, D.K.: Op. cit. pp.282 ~ 287.
Todd, D.K.: Sea Water Intrusion in Coastal Aquifers, Trans. Amer. Geophysical Union, vol. 34, No.5, 1953.

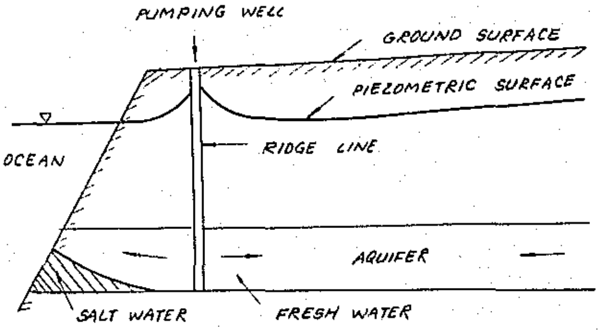

Fig. 51 Control of sea water intrusion by a pressure ridge paralleling the coast

v) To construct a barrier of sheet pilling, asphalt, concrete or puddle clay in relatively shallow aquifers, or a wall of impermeable membrane by injecting through a line of holes emulsified asphalt, cement grout, bentonite slurries, silicagel, or calcium acrylate for full utilization of groundwater storage impounded by the barrier.

2. Cause of deterioration of groundwater quality

We shall introduce here a table on various causes of deterioration of groundwater quality prepared by Water Resources Board, California, U.S.A.*

Table

Causes of Deterioration of Groundwater Quality

CONTAMINATION AND POLLUTION

Domestic and Municipal Sweage

Industrial Wastes
- Organic wastes
  - food processing
  - lumber processing
- Mineral wastes
  - metal processing industries
  - mining and ore extraction industries
  - oil industries
  - chemical industries
  - miscellaneous
- Cooling water

Solid and Semisolid Refuse

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*/Todd, D.K.: Op. cit. p.196.

DEGRADATION

Effects of Development, Use, and Reuse of Water

- Irrigation return water
  - surface drainage
  - percolation
- Interchange between aquifers due to improperly constructed, defective, or abandoned wells
- Interchange between aquifers due to differentials in pressure levels resulting from excessive withdrawal
- Overdraft conditions
  - sea water intrusion
  - salt balance
  - upward or lateral diffusion of connate brines and/or juvenile water due to overpumping
- Contamination from the surface due to improperly constructed wells

Natural Causes

- Inflow and/or percolation of juvenile water from highly mineralized springs and streams

Other Causes

- Accelerated erosion
- Mineralization resulting from plant transpiration and/or evaporation

WELL HYDRAULICS

by Dr. Gunjiro Sakai
Prof., Hirosaki University

PRESCRIPT

Groundwater is discriminated stratum water and fissure water roughly. But "well Hydraulics" concerns only the former, because it is very difficult at present to treat mathematically the movement of the latter in the earth.

Stratum water is classified also into two, that is, free surface groundwater, shortening free surface water, and piestic surface groundwater, shortening piestic surface water. The former has aeration zone and saturation zone, and the boundary of two zonesis called by the term of groundwater table. The aquiclude under the saturation zone is named basal bed, and the aquifer is called by the term of water table aquifer or free surface aquifer. It is not correct to call it by the term of unconfined aquifer or non-aetesian aquifer geohydrologically. Piestic surface water has no real groundwater surface, but it has an imaginary groundwater surface called by the term of piestic surface whose continuation is recognized as a reflection of the continuation of groundwater within the aquifer concerned. Aquicludes being above and below of the aquifer are called by the term of confining beds, then the aquifer is called by the term of confined aquifer too. In addition, upper confining bed is called positive and lower one negative respectively.

A well boared into free surface aquifer is termed as free surface water well or water table well and the one into piestic surface aquifer as piestic surface water well or tapped well. The latter is sometimes called by the term of artesian well, but artesian phenomenon does not limit to stratum water, then it is not a correct term. Concerning pumping test, term of "water table condition" is used for the former, but for the latter there is no proper term at present, then for the convenience, term of "artesian condition" or "piestic condition" is used in general.

Groundwater surface in a well is termed phreatic surface or phreatic level for both kinds of stratum water mentioned above. Formerly, the term "phreatic surface" was used as the groundwater surface of water table aquifer, but "phreat" means "well" in Greek, then it is incorrect.

CHAPTER I. REVIEW OF WELL THEORIES

When a well is pumped, the phreatic surface drops according with the discharge. But after some periods of pumping the descent of the phreatic surface stops in general, and they say that the pumping has reached to its steady state. On the other hand, there is another case when the descent does not cease as long as the pumping continues, though the withdrawal remains constant, and they say that the state is of non-steady. The drop of groundwater surface (in the broad meaning) occurs not only in the well (the phreatic surface) but also near the outside.

Thiem's theory is a most practical one, although it was published (1906) early. The theory treates a steady state pumping and belongs to a kind of equilibrium theory. Problems contained in this theory are as follows; 1) it does not explain the recharge of well water, 2) it does not treat any state until the steady state occurs. An area around the well where the groundwater surface drops owing to the pumping is called by the term of pumping depression or influence circle. But an actual condition of influence circle is not clear. Brinkhaust and Smreker published such theories independently that either formula brought forward by them does not contain a so-called radius of influence circle. These two belong also to equilibrium theory. After these theories two other theories were made public by Schultze and Weber, and either of them may be classified as the kind of non-equilibrium theory. However, that of the former is applied to a water table well and that of the latter to a tapped well only. These two theories are excepted in this lecture as they are too special.

In 1935, a famous theory represented by Theis' non-equilibrium formula was published and it is a most popular theory among geohydrologists nowadays. This theory presupposes to consume the groundwater storage of concerned area during pumping. Therefore, it presupposes also that steady state of pumping never occurs and the radius of influence circle expands as long as the pumping continues. This is a kind of non-equilibrium theory, but as it is grounded on an elasticity of aquifer or a compressibility of groundwater within it, it is known by the name of elastic theory. Leaky theory was proposed by Jacob in permeable positive confining bed from upper water table aquifer. This is also a kind of non-equilibrium theory.

Thiem, G: Hydrologische Methode, Leipzig, 1906

Brinkhaus, P.: Anlagen zur Gewinnung von naturlichen und künstlichen Grundwasser, München und Berlin, 1920

Smreker, O.: Das Grundwasser, seine Erscheinungsformen, Bewegungsgesetze und Mengenbestimmung, Leipsig und Berlin, 1914

Schultze, J.: Die groundwasserabsenking in Theorie und Praxis, Berlin, 1928

Weber, H.: Die Reichweite von grundwasserabsenking mittels Phorbrunnen, Berlin, 1928

Theis, C.V.: The Relation between the Lowering of Piezometric Surface and the Rate and Duration of a Well using Groundwater Storage, Amer. Geophys. Union Trans., 1935

Jacob, C.E.: On the Flow of Water in Elastic Artesian Aquifer, Trans. Amr. Geophys. Union, 1940

According to our experiences, pumping of well is taken place in steady state in many cases. So the theoretical formula of well theory must adapt both of steady and non-steady pumping as a rule, and Leaky Theory answers the rule. But Jacob failed to notice the supplement of squeeze water being derived from hydrous confining aquicludes, though the idea has been advocated by many early scholars. In 1957, the writer made public Squeeze Theory which oncerned both leaky water and aquiclude water. However, either Leaky Theory or Squeeze Theory does not fit to practical application for its mathematical procedure. So Elastic Theory only will be taken in this lecture. By the way, Elastic Theory, Leaky Theory and Squeeze Theory are applicable only to tapped well in principle.

Recently, several well theories are recommended by Howard, Ferris and Others. However, they did not refer to the process of development of theories and limited only the descriptions of applicable formulas to pumping test. While Muscat developed a detailed argument, either in basic theory or applied one. He distinguished also the cases of compressible water and of non-compressible one. So the former corresponds to the non-steady pumping and the latter to the steady pumping. But his theory is a little complicate in the mathematical development.

Groundwater hydraulics is grounded on Darcy's Law. Nowadays, problem concerning a laminer flow or a turburent flow of groundwater are discussed in an applied well theory, but so far as the theory stands on Darch's Law, such discussion may be out of the question. Darcy's Law is expressed by the next formula:

$$v = k, i \qquad \ldots\ldots\ldots\ldots \quad (1.1)$$

here v is the filtration velocity of water, k is permeability and i is the gradient of groundwater flow. The relation between v and true velocity of groundwater v_t is as follows:

$$v = p. \, v_t \qquad \ldots\ldots\ldots\ldots \quad (1.2)$$

here p is an effective porosity of aquifer. For the convenience of mathermatical treatment, the characters of aquifer and groundwater within it are assumed in all well theories as follows, though the assumption would not be reasonable, because the structure of aquifer and water flow within it is very complicate in nature.

(1) Composing materials of aquifer is homogeneous and isotropic for the groundwater flow.

(2) The thickness of aquifer is unchangeable in everywhere and spreads infinitely.

Sakai, G.: Study on Hydraulic Theory of Well, Bull. Hirosaku Univ. Educ. Fac. (1) 1957; (2) 1958

Howard, P. H.: A Histrical Review of Investigations of Seepage toward Wells, Jour. Boston Soc. Civil Eng., 1954

Ferris, J. G.: D.B. Knowles, and R.W. Stallman: Theory of Aquifer Test, U.S.G.S., Water-Supply Paper 1536-E, 1962

Muscat, M.: The Flow of Homogeneous Fluids through Porous Media, New York and London, 1937

(3) Confining beds or basal bed are horizontal and flat, and also parallel with groundwater surface (in broad meaning).

(4) Groundwater within aquifer does not flow naturally.

The mathematical treatment of groundwater flow through aquifer, that is, groundwater hydraulics is discussed on the ground of Laplace formula or Navier and Stokes formula, but in practical use those formulas would not be necessary to refer to. So in this lecture, Tiem's Theory and Theis' Theory will be explained, and then practical applied well hydraulics concerned will be studied with some exercises using actual pumping data.

CHAPTER II THIEM'S THEORETICAL FORMULAS

1. Tapped Well

The theory will be explained in accordance with Fig. 1, where m is the thickness of aquifer; $r\theta$ is the well radius; $h\theta$ is the height of phreatic surface measured upward from the upper surface of negative confining bed; h is the piestic surface of groundwater at the distance r from the center of well, measured as well as $h\theta$; r_e is the radius of influence circle, at which corresponding distance the piestic surface of groundwater coincides with the static water surface (in broad meaning) and at which the height of piestic surface becomes H measured by the same way mentioned above. The withdrawal of well is Q.

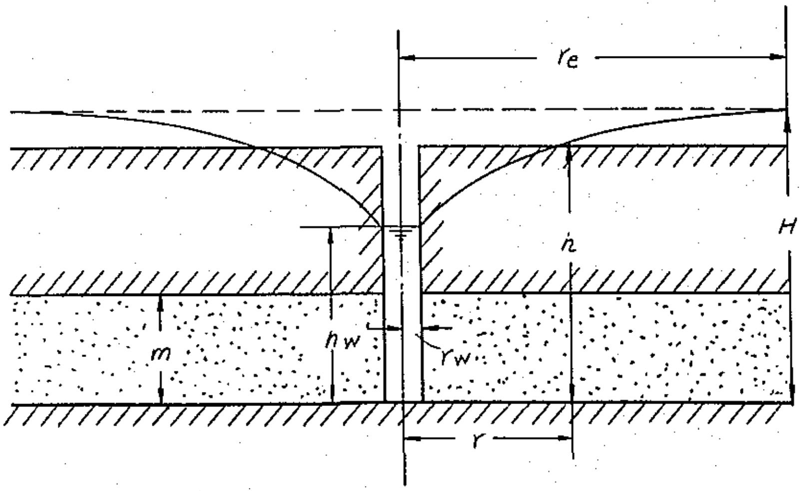


Fig. 1 Explanatory Schema of Tapped Well
– Equilibrium Theory –

The equation of motion of water in the pumping state is

$$v_r = -k \cdot \frac{dh}{dr} \qquad \text{.................} \qquad (2.1)$$

and the equation of continuity of flow is

$$Q = -2\pi r m v_r \qquad \text{.................} \qquad (2.2)$$

here v_r is the velocity of water flow at the distance r from the well.

From Equs. (2.1) and (2.2):

$$\frac{dh}{dr} = \frac{Q}{2\pi km} \cdot \frac{1}{r} \qquad \text{.................} \qquad (2.3)$$

is derived. By integration of Equ. (2.3) with follwoing boundary conditions are

$r = r\theta$; $h = h\theta$
$r = r_e$; $h = H$

the next equations are obtained.

$$H - h\omega = \frac{Q}{2\pi km} \cdot \log_e \frac{r_e}{r\omega} \quad \cdots\cdots\cdots \quad (2.4)$$

$$H - h = \frac{Q}{2\pi km} \cdot \log_e \frac{r_e}{r} \quad \cdots\cdots\cdots \quad (2.5)$$

By the substitution as $H - h\omega = s\omega$, $H - h = s$, $km = T$ (transmissibility), Equs. (2.4) and (2.5) are transformed as follows:

$$s\omega = \frac{Q}{2\pi T} \log_e \frac{r_e}{r\omega} \quad \cdots\cdots\cdots \quad (2.6)$$

$$s = \frac{Q}{2\pi T} \log_e \frac{r_e}{r} \quad \cdots\cdots\cdots \quad (2.7)$$

Equs. (2.4) – (2.7) are the theoretical formulas of Thiem for tapped well. dh/dr in Equ. (2.3) is a gradient at a depressed piestic surface. Therefore, the distance at which its value becomes zero must be $r = \infty$, but in this theory it is presupposed to be finite, that is, $r = r_e$.

When the drops of piestic surface are s_1 and s_2 at the distance r_1 and r_2 from the pumping well respectively, next simultaneous equations may be derived from Equ. (2.7).

$$\left.\begin{aligned} s_1 &= \frac{Q}{2\pi T} \log_e \frac{r_e}{r_1} \\ s_2 &= \frac{Q}{2\pi T} \log_e \frac{r_e}{r_2} \end{aligned}\right\} \quad \cdots\cdots\cdots \quad (2.8)$$

Eliminating r_e from the equations,

$$s_1 - s_2 = \frac{Q}{2\pi T} \log_e \frac{r_2}{r_1} \quad \cdots\cdots\cdots \quad (2.9)$$

or

$$s_1 - s_2 = \frac{2.3Q}{2\pi T} \log \frac{r_2}{r_1} \quad \cdots\cdots\cdots \quad (2.9')$$

is obtained, and either of them represents the base of opsilon method, by which k of the aquifer is calculated. Fig. 2 explains the configuration of wells concerned. Equ. (2.9') may be transformed as follows:

$$T = \frac{0.366Q}{s_1 - s_2} \log \frac{r_2}{r_1} \quad \cdots\cdots\cdots \quad (2.10)$$

here s_1, s_2, r_1, r_2 and Q may be measured all by the actual pumping test. If the thickness of aquifer m is known, for instance, by test boring, k (= T/m) will be computed too.

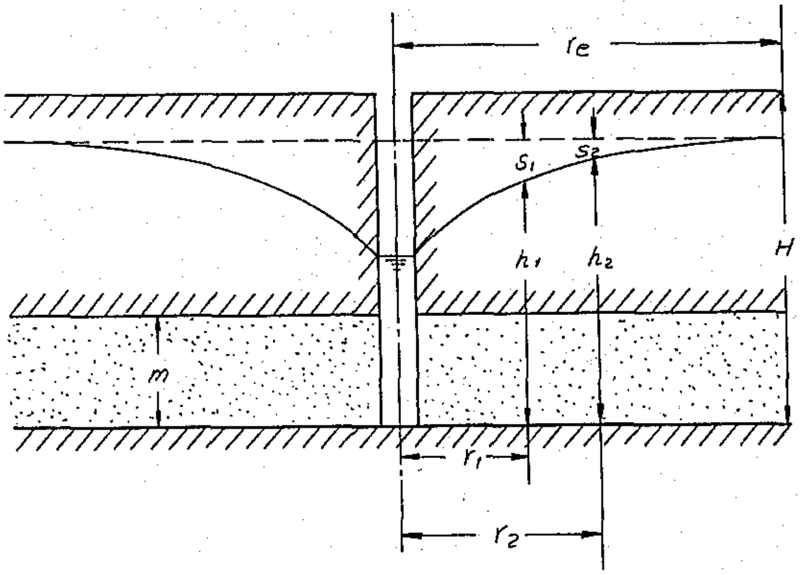


Fig. 2 Explanatory Schema of Epsilon Method

EXERCISE I

The withdrawal of a pumping well in steady state is $Q = 3{,}600\ m^3/day$ and the other values corresponding to Fig. 2 are $r_1 = 10\ m$, $r_2 = 50\ m$, $s_1 = 55\ cm$, $s_2 = 9\ cm$ respectively, when the thickness of aquifer is $m = 10\ m$.

Equ. (2.7) is modified as follows:

$$s = \frac{2.3Q}{2\pi T} \log \frac{r_e}{r} \qquad \cdots\cdots\cdots\cdots\cdots\cdots \quad (2.7')$$

and moreover transformed into

$$\log r_e = \log r + \frac{2.73sT}{Q} \qquad \cdots\cdots\cdots\cdots\cdots\cdots \quad (2.7'')$$

Thus, the radius of influence circle r_e must be computed theoretically, but Equ. (2.7") can not apply successfully.

EXERCISE II

Under the same conditions as well as EXERCISE I, compute r_e.

When two different withdrawals Q_1 and Q_2 are corresponding to drawdowns $s\omega_1$ and $s\omega_2$ respectively, from Equ. (2.6), the next simultaneous equations will be formed, if r_e remains unchangeable,

$$\left.\begin{aligned} s\omega_1 &= \frac{Q_1}{2\pi T} \log_e \frac{r_e}{r\omega} \\ s\omega_2 &= \frac{Q_2}{2\pi T} \log_e \frac{r_e}{r\omega} \end{aligned}\right\} \qquad \cdots\cdots\cdots\cdots\cdots\cdots \quad (2.11)$$

CHAPTER III OTHER EQUILIBRIUM THEORIES

(A) Brinkhaus' Theoretical Formulas

Brinkhaus' Theory belongs also to a kind of equilibrium theory, but it can not be applied to the pumping well itself, because he noticed a discontinuation between the phreatic surface and groundwater surface (in broad meaning) outside of well, and this fundamental limitation is common to either tapped well or water table well.

1. Taped Well

The theory will be explained in accordance with Fig. 1. The equation of motion of water in pumping state is as follows:

$$\frac{dh}{dr} = -k' v_r^n \quad \cdots\cdots\cdots\cdots\cdots \quad (3.1)$$

here k' is a kind of coefficient concerning the permeation of aquifer but different with coefficient of permeability and n is a certain index number. The equation of continuity of flow is the same as Equ. (2.2), so the next equation is derived.

$$\frac{dh}{dr} = -k' \left(\frac{Q}{2\pi m}\right)^n \cdot \frac{1}{r^n} \quad \cdots\cdots\cdots\cdots\cdots \quad (3.2)$$

Integrating Equ. (3.2), under the boundary condition as $r = \infty$ for $h = H$, the next equation is obtained.

$$s = H - h = k' \left(\frac{Q}{2\pi m}\right)^n \cdot \frac{1}{(n-1)\, r^{n-1}} \quad \cdots\cdots\cdots\cdots\cdots \quad (3.3)$$

This is the Brinkhaus' theoretical formula for tapped well.

The index number n is determined by the following procedure. When the drops of piestic water surface are s_1 and s_2 corresponding to the distances from the well, that is, r_1 and r_2 respectively, the next simultaneous equations are formed,

$$\left.\begin{aligned} s_1 &= k' \left(\frac{Q}{2\pi m}\right)^n \cdot \frac{1}{(n-1)\, r_1{}^{n-1}} \\ s_2 &= k' \left(\frac{Q}{2\pi m}\right)^n \cdot \frac{1}{(n-1)\, r_2{}^{n-1}} \end{aligned}\right\} \quad \cdots\cdots\cdots\cdots\cdots \quad (3.4)$$

and then Equ. (3.5) is derived,

$$\frac{s_1}{s_2} = \left(\frac{r_2}{r_1}\right)^{n-1} \quad \cdots\cdots\cdots\cdots\cdots \quad (3.5)$$

The index number is computed from Equ. (3.5) directly, but it is obtained also by the following graphical solution.

$$\log s_1 - \log s_2 = (n\text{-}1) \cdot (\mathrm{long}\ r_2 - \log r_1) \quad \ldots\ldots\ldots\ldots \quad (3.6)$$

represents a straight line of gradient of (n-1) in the relation curve plotted on a logarismic section paper, and n is obtained. In practice, many values of drops s_1, s_2, ..., s_n may be measured at the corresponding observation wells whose distances from the pumping well are r_1, r_2, ... , r_n respectively, and in some cases a few plots deviate from the straight line, by the reason that r adopted is too small and probably occurs turburent flow there. An example is indicated in Fig. 4.

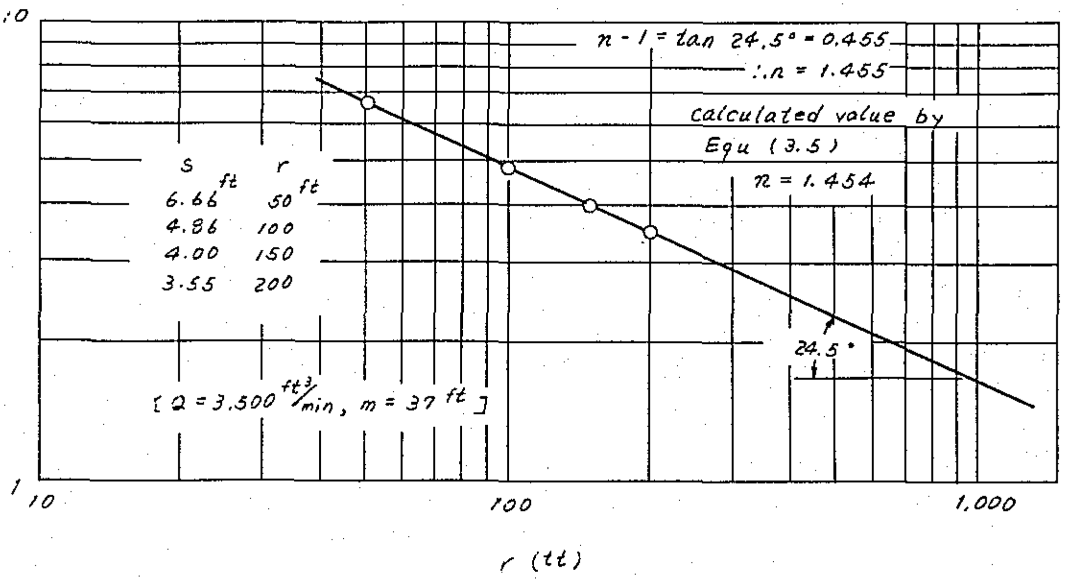


Fig. 4 Graphical Solution of Index Number n for Tapped Well
— Brinkhaus' Theory —

EXERCISE III

Derive an equation from Equ. (3.3), which can express the relation between Q and s at a certain distance from the pumping well concerned, and then using the data of the above example and also the value of n obtained, calculate the ratio of s to make the discharge twofold'

2. Water Table Well

The theory will be explained in accordance with Fig. 3. The equation of motion of water in the pumping state is identical with Equ. (3.1) and that of continuity of flow with Equ. (2.16), then the next equation may be derived,

$$\frac{dn}{dr} = k' \left(\frac{Q}{2\pi}\right)^n \cdot \frac{1}{rn} \cdot \frac{1}{h^n} \quad \ldots\ldots\ldots\ldots\ldots \quad (3.7)$$

Integrating the equation under the same condition as the case of tapped well,

$$H^{n+1} - h^{n+1} = k' \left(\frac{q}{2\pi}\right)^{n} \cdot \frac{n+1}{n-1} \cdot \frac{1}{r^{n-1}} \quad \cdots\cdots\cdots\cdots\cdots \quad (3.8)$$

When the saturation thickness h of a certain observation well whose distance is r from the pumping well changes to h_1 or h_2 corresponding to the discharge Q_1 or Q_2 respectively, from Equ. (3.8) the next equation is derived,

$$\frac{H^{n+1} - h_2^{\,n+1}}{H^{n+1} - h_1^{\,n+1}} = \left(\frac{Q_2}{Q_1}\right)^{n} \quad \cdots\cdots\cdots\cdots\cdots\cdots \quad (3.9$$

This equation can not be solved easily. So assuming various values of n and computing both sides of the equation separately, two curves are drawn on a section paper, values n on abscessa and those of computed values on ordinate. Thus, the intersection of two curves will give the desirous value of n. An example is indicated in Fig. 5.

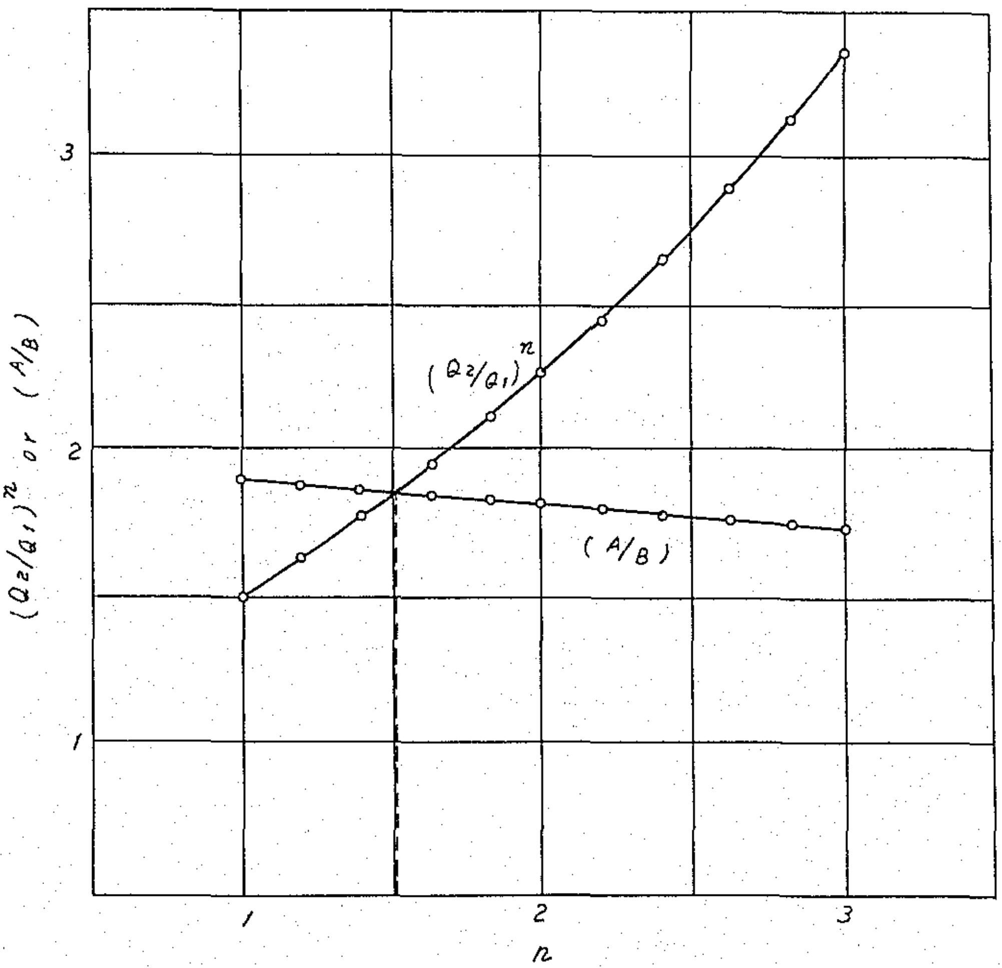


Fig. 5 Graphical Solution of Index Number n for Water Table Well — Brinkhaus' Theory —

A: $H^{n+1} - h_2^{n+1}$, B: $H^{n+1} - h_1^{n+1}$

EXAMPLE DATA: $Q_1 = 1{,}000$ ft^3/min., $Q_2 = 1{,}500$ ft^3/min.

EXERCISE IV

Using Equ. (3.3), the data of above example and also the value of n obtained, calculate the ratio of s, the one at the observation well and the another at twofold distant one from the pumping well.

(B) Smreker's Theoretical Formulas

Smreker's theory belongs to the equilibrium theory too and in the thoretical formulas the radius of influence circle r_e is not contained as well as those of Brinkhaus. But Smreker's theory refers to the pumping well itself as that of Tiem's one.

1. Tapped Well

The theory will be explained in accordance with Fig. 1, in it. The equation of motor of water in this theory is as follows:

$$\frac{dh}{dr} = \frac{\tau}{2\vartheta} \cdot v_r^{\frac{3}{2}} \qquad \ldots\ldots\ldots\ldots\ldots\ldots \quad (3.10)$$

here τ is a certain constant concerning with permeation and ϑ indicates acceleration of gravity. The equation of continuity of flow is as follows:

$$v_r = \frac{Q}{2\pi prm} \qquad \ldots\ldots\ldots\ldots\ldots\ldots \quad (3.11)$$

here p is an effective porosity of aquifer. From two equations above, the next equation is derieved.

$$\frac{dh}{dr} = \frac{\tau}{2\vartheta} \cdot \left(\frac{Q}{2\pi prm}\right)^{\frac{3}{2}} \qquad \ldots\ldots\ldots\ldots\ldots\ldots \quad (3.12)$$

Integrating the equation under the conditions;

$h = H$; $r = \infty$
$h = h\omega$; $r = r\omega$

the next equations are obtained.

$$H - h = s = \frac{\tau}{\vartheta} \left(\frac{Q}{2\pi pm}\right)^{\frac{3}{2}} \cdot \frac{1}{r^{\frac{1}{2}}} \qquad \ldots\ldots\ldots\ldots\ldots\ldots \quad (3.13)$$

$$H - h\omega = s\omega = \frac{\tau}{\vartheta} \left(\frac{Q}{2\pi pm}\right)^{\frac{3}{2}} \cdot \frac{1}{r\omega^{\frac{1}{2}}} \qquad \ldots\ldots\ldots\ldots\ldots\ldots \quad (3.14)$$

here $s\omega$ is the drop of piestic surface of groundwater outside of well and s the drawdown of phreatic surface respectively. These equations are the Smreker's theoretical formulas for tapped well. If the following substitution,

$$k'' = \frac{\tau}{\vartheta} \left(\frac{1}{p}\right)^{\frac{3}{2}} \qquad \cdots\cdots\cdots\cdots\cdots \qquad (3.15)$$

is made, k'' represents a certain constant concerning with the permeation of aquifer.

When the different drawdowns of $s\omega_1$ and $s\omega_2$ are occured in the pumping well itself, corresponding to the different discharges Q_1 and Q_2 respectively, the next simultaneous equations are obtained,

$$\left.\begin{aligned} s\omega_1 &= k'' \left(\frac{Q_1}{2\pi m}\right)^{\frac{3}{2}} \cdot \frac{1}{r\omega^{\frac{1}{2}}} \\ s\omega_2 &= k'' \left(\frac{Q_2}{2\pi m}\right)^{\frac{3}{2}} \cdot \frac{1}{r\omega^{\frac{1}{2}}} \end{aligned}\right\} \qquad \cdots\cdots\cdots\cdots\cdots \qquad (3.16)$$

and easily the following equation is derived,

$$\frac{s\omega_2}{s\omega_1} = \left(\frac{Q_2}{Q_1}\right)^{\frac{3}{2}} \qquad \cdots\cdots\cdots\cdots\cdots \qquad (3.17)$$

The same relation may be concluded from Equ. (3.13) between the different drops of s at the observation well distant r from the pumping well and the corresponding discharges of Q.

$$\frac{s_2}{s_1} = \left(\frac{Q_2}{Q_1}\right)^{\frac{3}{2}} \qquad \cdots\cdots\cdots\cdots\cdots \qquad (3.18)$$

When the drawdown of pumping well is maintained constant, the relation between discharge Q and the radius of pumping well $r\omega$ may be derived by the next simultaneous equations as follows:

$$\left.\begin{aligned} s\omega &= k'' \left(\frac{Q_1}{2\pi m}\right)^{\frac{3}{2}} \cdot \frac{1}{r\omega_1^{\frac{1}{2}}} \\ s\omega &= k'' \left(\frac{Q_2}{2\pi m}\right)^{\frac{3}{2}} \cdot \frac{1}{r\omega_2^{\frac{1}{2}}} \end{aligned}\right\} \qquad \cdots\cdots\cdots\cdots\cdots \qquad (3.19)$$

and then

$$r\omega_2 = r\omega_1 \left(\frac{Q_2}{Q_1}\right)^3 \qquad \cdots\cdots\cdots\cdots\cdots \qquad (3.20)$$

or

$$Q_2 = Q_1 \left(\frac{r\omega_2}{r\omega_1}\right)^{\frac{1}{3}} \qquad \cdots\cdots\cdots\cdots\cdots \qquad (3.20')$$

The two Equs. (3.20) and (3.20') are practical and agree with our experiences comparatively well.

2. Water Table Well

The theory will be explained in accordance with Fig. 3, neglecting r_e in it. The equation of motion of water is the same with the case of tapped well but that of continuity of flow is expressed as follows:

$$Vr = \frac{Q}{2\pi prh} \qquad \cdots\cdots\cdots\cdots\cdots \qquad (3.21)$$

then the next equation is derived,

$$\frac{dh}{dr} = \frac{k''}{2} \left(\frac{Q}{2\pi rh}\right)^{\frac{3}{2}} \qquad \cdots\cdots\cdots\cdots\cdots \qquad (3.22)$$

Integrating the equation under the same conditions as the case of tapped well the next two equations are obtained.

$$H^{\frac{5}{2}} - h^{\frac{5}{2}} = \frac{5}{2} k'' \left(\frac{Q}{2\pi}\right) \cdot \frac{1}{r^{\frac{1}{2}}} \qquad \cdots\cdots\cdots\cdots\cdots \qquad (3.23)$$

$$H^{\frac{5}{2}} - h\omega^{\frac{5}{2}} = \frac{5}{2} k'' \left(\frac{Q}{2\pi}\right) \cdot \frac{1}{r\omega^{\frac{1}{2}}} \qquad \cdots\cdots\cdots\cdots\cdots \qquad (3.24)$$

These equations are the Smreker's theoretical formulas for the water table well.

The Smreker's formulas, however, are unpractical, and the relation between discharge and drawdown can not be easily derived like as Equ. (3.17) for the case of tapped well. Moreover, notwithstanding the same relation between discharge and well radius is derived as like as Equ. (3.20) or Equ. (3.20'), it is far from our experiences. If the index number n in the theoretical formulas of Brinkhaus amounts 3/2, they agree with the theoretical formulas of Smreker. As the natural hydrogeology is very complicated, Brinkhaus' formulas in which the values of index number n must be searched experimentally will do better for practical use.

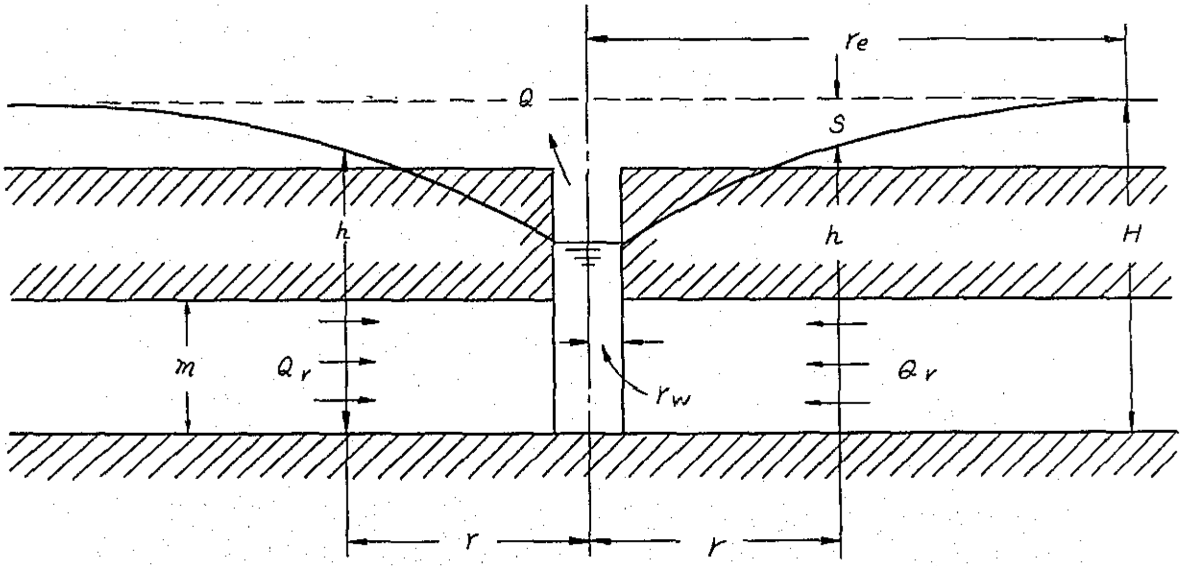
In general situation, water flows vertically into the water table well from its bottom only. Under such well structure, the discharge of well increases in direct proportion to radius. But we must remember also that the discharge by such structure is fairly little compared with that by the structure which water flows laterally into the well through strainer of well casing.

CHAPTER IV. THEIS' THEORETICAL FORMULA

Theis' Non-equilibrium Formula represents a non-equilibrium theory. This theory is applied only to a tapped well in principle. For a water table well, it may be applied approximately so far as the saturation thickness is very thick compared with drawdown. But such a condition never occurs for the non-steady pumping.

This theory will be explained in accordance with Fig. 6, where Q_r is the quantity of water through the wall of vertical cylinder which is distant r from the pumping well and all other symbols are the same with those of Fig. 1.

Fig. 6 Explanatory Schema of Tapped Well
— Non-equilibrium Theory —



The equation of motion of water is

$$Q_r = -2\pi rkm \cdot \frac{\partial s}{\partial r} \qquad \cdots\cdots\cdots\cdots\cdots \qquad (4.1)$$

and the equation of continuity of flow is

$$\partial Q_r = -2\pi rS \cdot \frac{\partial s}{\partial t}\, \partial r \qquad \cdots\cdots\cdots\cdots\cdots \qquad (4.2)$$

here S is a coefficient of storage.* Eliminating ∂Q_r from Equs. (4.1) and (4.2), the next differential equation is derived,

* A coefficient of storage is defined by Wenzel as the quantities of water ft^3 withdrawn from the column of aquifer equivalent to the thickness and of cross section 1 ft^2, by the pressure drop corresponding with 1 ft H_2O. But it may be defined by C.G.S. unit too, because it is of non-dimensional mathematically.

$$\frac{\partial s}{\partial t} = \frac{k.m}{S}\left(\frac{\partial^2 s}{\partial^2 r} + \frac{1}{r} \cdot \frac{\partial s}{\partial r}\right) \qquad \text{(4.3)}$$

The equation may be solved under conditions as below;

Initial condition: $t = 0 \quad s = 0$
Boundary condition: $r = \infty \quad s = 0$

Assuming $r\omega$ zero practically,

$$\left.\begin{aligned} s &= \frac{Q}{4\pi mk}\int_u^\infty \frac{e^{-u}}{u}\,du \\ u &\equiv r^2\ S/4mkt \end{aligned}\right\} \qquad \text{(4.4)}$$

here $\int_u^\infty \frac{e^{-u}}{u}\,du$ is Ei Function and also is called Well Function, signified as W(u) well hydraulics. The equation is the famous Theis' Non-equilibrium Formula, and using transmissibility $T = km$, it is transformed as follows:

$$\left.\begin{aligned} s &= \frac{Q}{4\pi T}\int_u^\infty \frac{e^{-u}}{u}\,du \\ u &\equiv r^2/4t \cdot S/T \end{aligned}\right\} \qquad \text{(4.5)}$$

and further is abbreviated as below,

$$\left.\begin{aligned} s &= M \cdot W(u) \\ M &= \frac{Q}{4\pi T} \end{aligned}\right\} \qquad \text{(4.5')}$$

The above formula corresponds to the so-called pumping test, while Theistreated also the case of the so-called recovery test referring to the ascending state of groundwater surface after the pumping stops. He assumed that the negative pumping will begin at an instance when the actual pumping stopped, and formed such equations as follows:

$$\left.\begin{aligned} s &= \frac{Q}{4\pi T}\,[W(u) - W(u')] \\ u &\equiv \frac{r^2}{4t} \cdot \frac{S}{T}, \quad u' \equiv \frac{r^2}{4t'} \cdot \frac{S}{T} \end{aligned}\right\} \qquad \text{(4.6)}$$

$$t \equiv t_s - t'$$

here t' is the time of recovery and t_s is of pumping continued. According to the Theis' Theory, the recovery of groundwater surface has taken place by the consumption of storage outside of influence circle which continues to enlarge after the pumping stops.

1. Graphical Solution of Theis

For this solution a type curve must be prepared at first, which is drawn on logarismic section paper, W (u) on ordinate and u on abscessa respectively, using Ei function table. For the next, the measured data s and r^2/t are plotted in the same way, that is to say in other word, an actual measured curve is drawn on logarismic section tracing paper of the same scale. Then putting the latter upon the former and trying parallel transition, the match point of two curves is searched. Thus, from the former the values of W (u) and u, and from the latter those of s and r^2/t are decided. Then T and S are computed by the next equations.

$$T = \frac{Q}{4\pi s} \cdot W(u) \qquad \cdots\cdots\cdots\cdots\cdots \qquad (4.7)$$

$$S = 4Tu \cdot \frac{1}{r^2/t} \qquad \cdots\cdots\cdots\cdots\cdots \qquad (4.8)$$

From the match point:

by "type curve"

$u = 0.1, \quad W(u) = 1.82$

by "measured curve"

$r^2/t = 3.75 \times 10^5 \ \text{ft}^2/\text{d}$
$s = 11.6 \ \text{ft}$

$Q = 3{,}050 \ \text{ft}^3/\text{d}$

$$T = \frac{3{,}050 \times 1.82}{4 \times 3.14 \times 11.6} = 38 \ \text{ft}^3/\text{d}/\text{ft}$$

$$S = \frac{4 \times 38 \times 0.1}{3.75 \times 10} = 4.0 \times 10^{-5}$$

By Fig. 7, an example of the solution will be explained. However, in the procedure mentioned above the match point is unreliable in practice, because the measured curves are not smooth in many cases.

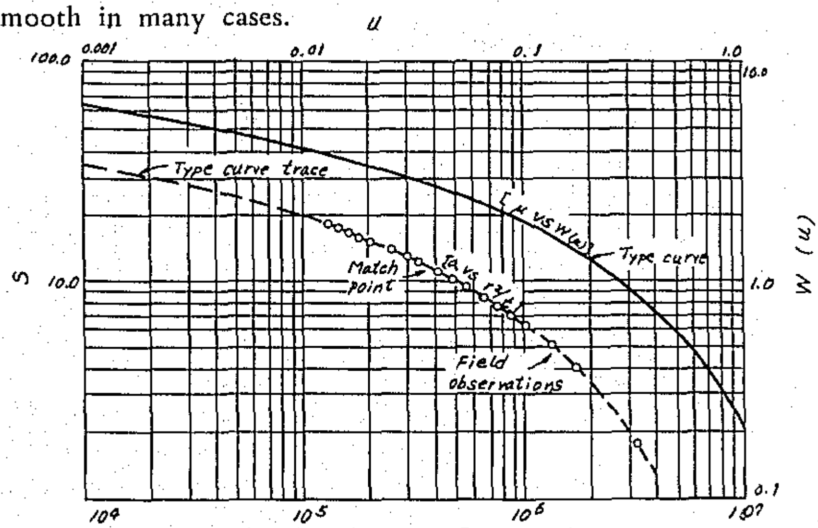


Fig. 7 Graphical Solution of Theis by Type Curve

2. The Approximate Solution of Jacob

This solution is of graphical too. Under the condition u < 1, W(u) is developed in a series as follows:

$$W(u) = -0.5772 - \log_e u + u - \frac{u}{1\ (2!)} + \frac{u}{3\ (3!)} - \cdots \qquad (4.9)$$

neglecting the terms contained u, u^2, u^3, ..., and also considering Equ. (4.5′), Equ. (4.9) is approximated as follows:

$$s = M\ (-0.5772 - \log_e u) \qquad \ldots\ldots\ldots\ldots\ldots \qquad (4.10)$$

When two different drops of piestic groundwater surface in an observation well distant r from pumping well are s_1 and s_2, and the corresponding pumping times are t_1 and t_2 respectively, the next equation is derived,

$$s_2 - s_1 = M.\log_e \frac{t_2}{t_1} = 2.3\ M.\log \frac{t_2}{t_1} \qquad \ldots\ldots\ldots\ldots\ldots \qquad (4.11)$$

Plotting the data of pumping test on a semi-logarismic section paper, the so-called s-t curve is obtained, in which s on ordinate (normal graduation) and t on abscessa (logarismic graduation). When u < 1 is satisfied, in other words, when t is sufficiently large, the s-t curve indicates a straight line whose gradient is 2.3M. If the difference of two values of s corresponding to one cycle of logarismic graduation is expressed by Δs for the straight line mentioned above, transmissibility T will be computed by the next equation,

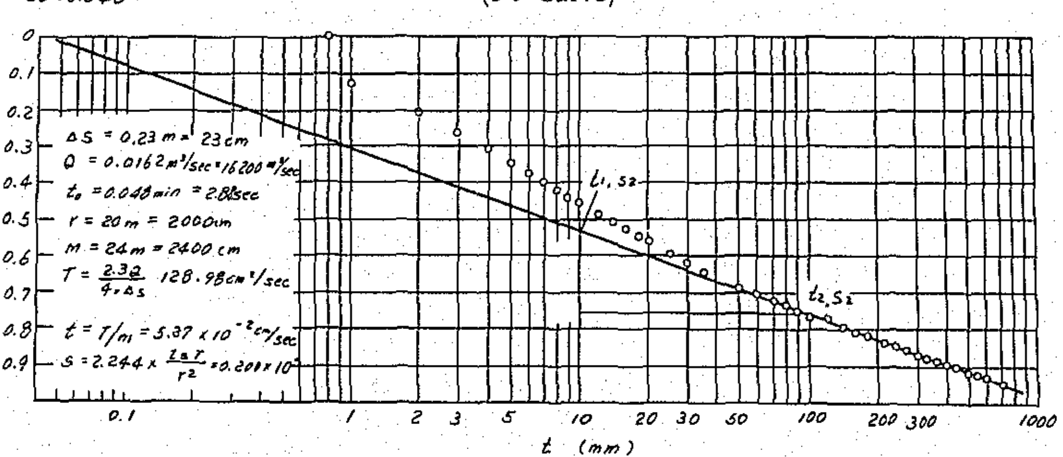
$$T = \frac{2.3Q}{4\pi\Delta s} \qquad \ldots\ldots\ldots\ldots\ldots \qquad (4.12)$$

In this procedure, when the straight line is extended upwards and the time corresponding the intersection with s = 0 line is symbolized by t_0, the next equations are derived,

$$\left.\begin{aligned} -\log_e u_0 &= 0.5772 \\ u_0 &\equiv \frac{r^2}{4t_0}\ \frac{S}{T} \end{aligned}\right\} \qquad \ldots\ldots\ldots\ldots\ldots \qquad (4.13)$$

and then coefficient of storage S is computed. An example is indicated in Fig. 8.

Fig. 8 Graphical Solution of Jacob for Pumping Test
(s-t Curve)



EXERCISE V

Using the pumping test data of inserted table, compute T and S of aquifer of the well concerned, by graphical method of Jacob.

Data Table of Pumping Test for EXERCISE V

Withdrawal of Pumping Well: $Q = 506 m^3/d$
Distance of Observation Well: $r = 20m$

Observation h m	Time elapsed (min.)	Drop (m)
7.15	(pumping start)	
	1	0.122
	2	0.325
	3	0.423
	4	0.528
	5	0.602
	6	0.667
	7	0.723
	8	0.763
	9	0.808
7.25	10	0.838
	12	0.903
	14	0.946
	16	0.988
	18	1.033
7.35	20	1.062
	25	1.131
	30	1.200
	35	1.236
	40	1.275
	50	1.352
9.15	60	1.402
	70	1.445
	80	1.493
	90	1.518
	100	1.554
9.15	120	1.598
	140	1.643
	160	1.670
10.15	180	1.708
	210	1.734
11.15	240	1.758
	270	1.781
12.15	300	1.798
	330	1.830
13.15	360	1.845
	390	1.870

Observation h m	Time elapsed (min.)	Drop (m)
14.15	420	1.870
	450	1.883
15.15	480	1.890
	510	1.890
16.15	540	1.913
	570	1.912
17.15	600	1.923
	630	1.935
18.15	660	1.956
	690	1.958
19.15	720	1.976

For the recovery test, the next equation is derived from two Equs. (4.6) and (4.10),

$$s = M.\log_e \frac{t}{t'} = 2.3M.\log \frac{t}{t'} \qquad \ldots\ldots\ldots\ldots\ldots\ldots \qquad (4.14)$$

A curve drawn in accordance with the above equation using recovery test data is called by the name of s-t/t′ curve, which is drawn on semi-logarismic section paper, t/t′ on abscessa. In the actual procedure, the value of s is selected corresponding to t/t′ = 10, that is log t/t′ = 1, and s in this case is called s′ tentatively, then s′ 2.3M. This method, however, can not obtain S directly, for which it will be explained later. When the condition u′ < 1 is not satisfied, this method will not give a correct solution.

Fig. 9 Graphical Solution of Jacob for Recovery Test
(s-t/t′ Curve)

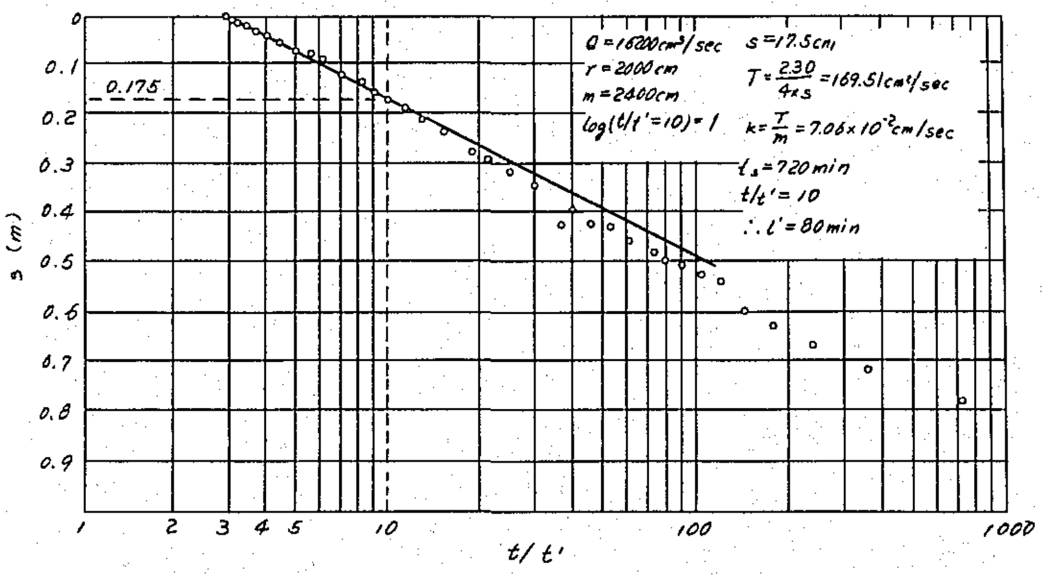


Fig. 9 indicates an example solution by this method according to the same data as of the example of Fig. 8. But in this case, $t_s = 720$ min. and $t/t' = (t_s + t')/t' = 10$, then $t' = 80$ min. which is too small to satisfy the condition $u^1 > 1$.

While Jacob proposed another method for recovery test. When two different depths of recovered phreatic surface in the observation well, being called generally by the term of drops too, at r distance from the pumping well are s_1 and s_2, corresponding to the times $t_1 = t_s + t_1'$ and $t_2 = t_s + t_2'$, from Equs. (4.6) and (4.10), the next equation will be derived.

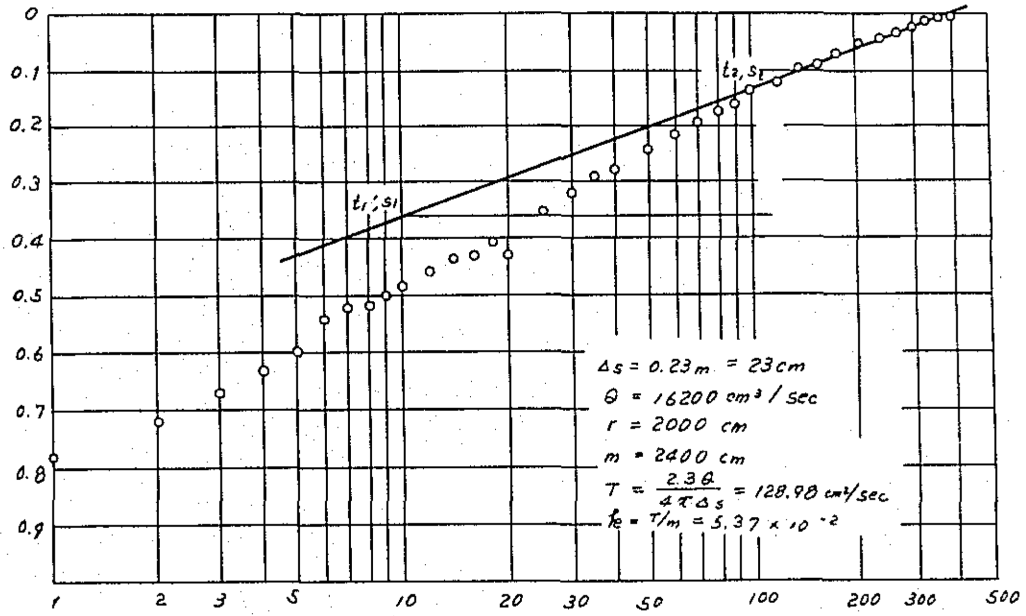
$$s_1 - s_2 = 2.3M. \log \left(\frac{t_s + t_1'}{t_s + t_2'} \cdot \frac{t_2'}{t_1'}\right) \qquad \cdots\cdots\cdots\cdots \quad (4.15)$$

If t_s is satisfactorily large compared with t_1' and t_2', Equ. (1.15) is transformed approximately as below,

$$s_1 - s_2 = 2.3M. \log \left(\frac{t_2'}{t_1'}\right) \qquad \cdots\cdots\cdots\cdots \quad (4.16)$$

Under the condition $u' < 1$, that is, the values of t^1 are very large, the relation curve between s and t^1 expressed on a semi-logarismic section paper represents a straight line as well as the case of pumping test. This curve is called by the name of s-t^1 curve. When the $s_1 - s_2$ is expressed by Δs corresponding to one cycle of logarism, T will be computed by Equ. (4.12) mentioned already. An example is indicated in Fig. 10.

Fig. 10 Graphical Solution of Jacob for Recovery Test
(s-t′ Curve)



This method can not be obtained S directly as mentioned already. But when t_s (the time of actual pumping) and s_s (the drop at the instance of pumping stop) are known, S will be computed by the following equations, using W(u) table,

$$\frac{4\pi s_s T}{Q} = W(u_s) \qquad \cdots\cdots\cdots\cdots\cdots \qquad (4.17)$$

$$u_s \equiv \frac{r^2}{4t_s} \frac{S}{T}$$

EXERCISE VI

Using the recovery test data of inserted table concerned the same well taken in EXERCISE V, compute T of aquifer of the well by the graphical solution of Jacob.

Data Table of Recovery Test for EXERCISE VI

Withdrawal of Pumping Well: $Q = 506 m^3/d$
Distance of Observation Well: $r = 20 m$
Time of Pumping Duration: $t_s = 720$ min.

Observation h m	Time elapsed (min.)	Drop (m)
19.15	(pumping stop)	
	1	1.871
	2	1.715
	3	1.561
	4	1.451
	5	1.361
	6	1.290
	7	1.229
	8	1.186
	9	1.143
19.25	10	1.105
	12	1.055
	14	0.996
	16	0.957
	18	0.909
19.35	20	0.881
	25	0.818
	30	0.758
	35	0.714
	40	0.662
	50	0.595
20.15	60	0.538
	70	0.492
	80	0.456
	90	0.420
	100	0.396

Observation h m	Time elapsed (min.)	Drop (m)
21.15	120	0.343
	140	0.301
	160	0.278
22.15	180	0.248
22.45	210	0.211
23.15	240	0.184
	270	0.158
0.15	300	0.140
	330	0.125
1.15	360	0.111
	390	0.093
2.15	420	0.076
	450	0.067
3.15	480	0.056
	510	0.048
4.15	540	0.039
	570	0.034
5.15	600	0.029
	630	0.026
6.15	660	0.023
	690	0.017
7.15	720	0.010

GEOPHYSICAL EXPLORATION,
ELECTRIC PROSPECTING
AND
ELECTRIC WELL LOGGING FOR GROUNDWATER

By working on the earth's surface, it is sometimes possible to estimate where groundwater occurs and, under special conditions, to obtain information on water quality. Investigating groundwater from the surface is at best not easy, nor are results always successful; however, such methods are normally less costly than subsurface investigations. Geophysical methods, developed in the last thirty years for petroleum and mineral exploration, have proved useful for locating and analyzing groundwater. Although several methods can be enumerated under the geophysical heading, only the electrical resistivity and deismic refraction methods have more than limited application to groundwater. Geologic investigation and reconnaissance represent a second approach to the problem. All methods provide only indirect indications of groundwater, as underground hydrologic data must be inferred from surface information. Correct interpretation requires supplemental data from subsurface investigations to substantiate surface findings.

Geophysical Exploration

Geophysical exploration refers to the scientific measurement of physical properties of the earth's crust for investigation of mineral deposits or geologic structure. With the discovery of oil by geophysical methods in 1926, economic pressures for the location of petroleum and mineral deposits stimulated the development and improvement of many geophysical methods and equipment. Application to groundwater investigations was slow because the commercial value of oil overshadows that of water. In recent years, however, refinement of geophysical techniques as well as an increasing recognition of the advantages of the methods for groundwater study has changed the situation. Today, many organizations responsible for meeting water supply demands are employing geophysical methods. The methods are frequently inexact or difficult to interpret, and they are most useful when supplemented by subsurface investigations.

Geophysical methods detect differences, or anomalies, of physical properties within the earth's crust. Density, magnetism, elasticity, and electrical resistivity are properties most commonly measured. Experience and research have enabled differences in these properties to be interpreted in terms of geologic structure, rock type and porosity, water content, and water quality.

Electrical Resistivity Method

The electrical resistivity of a rock formation limits the amount of current passing through the formation when an electrical potential is applied. It may be defined as the resistence in ohms between opposite faces of a unit cube of the material. If a material of resistance R has a cross-sectional area A and a length L, then its resistivity can be expressed as

$$\rho = \frac{RA}{L} \qquad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots \qquad (9.1)$$

In the metric system units of resistivity are $\text{ohm-m}^2/\text{m}$, *or simply* ohm-m.

Resistivities of rock formations vary over a wide range, depending upon the material, density, porosity, pore size and shape, water content and quality, and temperature.

There are no fixed limits for resistivities of various rocks; igneous and metamorphic rocks yield values in the range 10^2 to 10^8 ohm-m, sedimentary and unconsolidated rocks, 100 to 10^4 ohm-m. In relatively porous formations, the resistivity is controlled more by water content and quality within the formation than by the rock resistivity; therefore, for aquifers composed of unconsolidated materials, the resistivity of the groundwater governs. Aquifer resistivity can be expressed in terms of resistivity of the groundwater and of porosity for uniform packings of spherical mineral grains. If ρ is the aquifer resistivity, ρw is that of the groundwater (assumed to fill all of the voids), and α is porosity, then it can be shown that

$$\frac{\rho}{\rho w} = \frac{3 - \alpha}{2\alpha} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots \quad (9.2)$$

This relation applies only for isotropic conditions; for anisotropic conditions, orientation of the current with respect to stratifications will produce resistivity differences.

Actual resistivities are determined from an apparent resistivity which is computed from measurements of current and potential differences between pairs of electrodes placed in the ground surface. The procedure involves measuring a potential difference between two electrodes (P in Fig. 9.1) resulting from an applied current through two other electrodes (C in Fig. 9.1) outside but in line with the potential electrodes. If the resistivity is everywhere uniform in the subsurface zone beneath the electrodes, an orthogonal network of circular arcs will be formed by the current and equipotential lines,

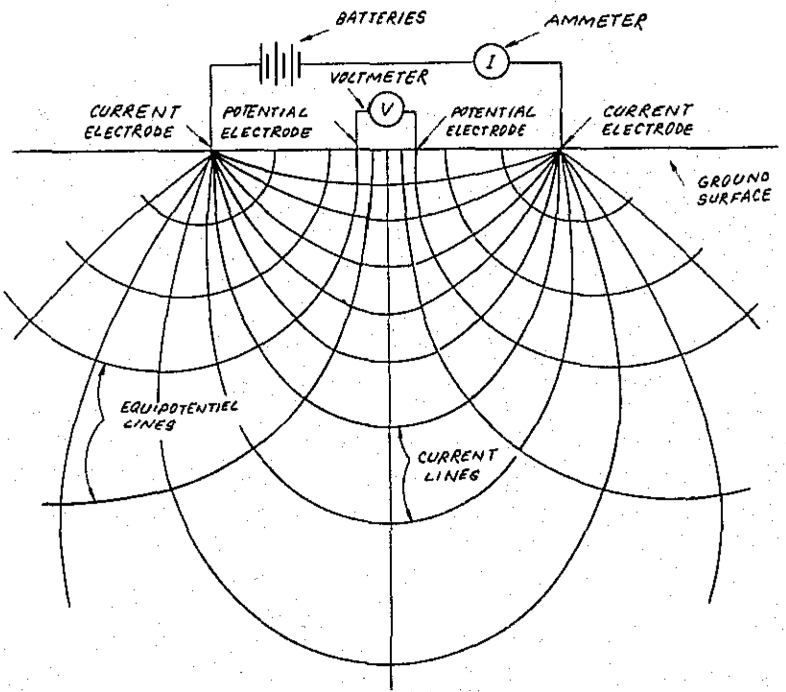


Fig. 9.1 Electrical circuit for resistivity determination and electrical field for a homogeneous subsurface stratum

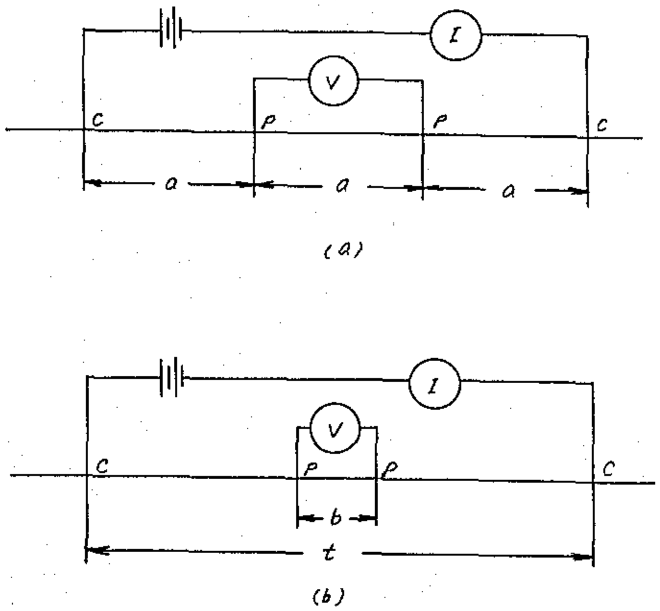


Fig. 9.2 Common electrode arrangements for resistivity determination: (a) Wenner and (b) Schlumberger

as shown in Fig. 9.1. The measured potential difference is a weighted value over a subsurface region controlled by the shape of the network. Thus, the measured current and potential differences yield an apparent resistivity over an unspecified depth. If the spacing between electrodes is increased, a deeper penetration of the electrical field occurs and a different apparent resistivity is obtained. In general, actual subsurface resistivities vary with depth; therefore, apparent resistivities will change as electrode spacing are increased, but not in a like manner. Because changes of resistivity at great depths have only a slight effect on the apparent resistivity compared to those at shallow depths, the method is seldom effective for determining actual resistivities below a few hundred feet.

Current electrodes consist of metal stakes driven into the ground; potential electrodes are porous cups filled with a saturated solution of copper sulfate to inhibit electrical fields from forming around them. To minimize polarization effect either and a-c low frequency current or a reversible direct current is preferable. In practice, various standard electrode spacing arrangements have been adopted; most common are the Wenner and Schlumberger arrangements.

The Wenner arrangement, shown in Fig. 9.2a, has the potential electrodes located at the third points between the current electrodes. The apparent resistivity is given by the ratio of voltage to current times a spacing factor. For the Wenner arrangement, the apparent resistivity

$$\rho a = 2\pi a \frac{V}{I} \quad \ldots\ldots\ldots\ldots\ldots\ldots \quad (9.3)$$

where a is the distance between adjacent electrodes, V is the voltage difference between the potential electrodes, and I is the applied current.

The Schlumberger arrangement, shown in Fig. 9.2b, has the potential electrodes close together. The apparent resistivity is given by

$$\rho a = \pi \frac{(L/2)^2 - (b/2)^2}{b} \frac{V}{I} \quad \dots\dots\dots\dots\dots\dots\dots\dots \quad (9.4)$$

where L and b are the current and potential electrode spacings, respectively (Fig. 9.2b). Theoretically, $L \gg b$, but for practical application good results can be obtained if $L \gg 5b$.

When apparent resistivity is plotted against electrode spacing (a for Wenner, and L/2 for Schlumberger) for various spacing at one location, a smooth curve can be drawn through the points. The interpretation of such a resistivity-spacing curve in terms of subsurface conditions is a complex and frequently difficult problem. The solution can be obtained in two parts: (a) interpretation in terms of various layers of actual (as distinguished from apparent) resistivities and their depths; and (b) interpretation of the actual resistivities in terms of subsurface geologic and groundwater conditions. Part (a) *can be accomplished with theoretically computed* resistivity-spacing curves of 2-, 3-, and 4-layer cases for various rations of resistivities. Curves and explanation of curve-matching techniques have been published by Mooney and Wetzel for the Wenner configuration; by La Compagnie Générale de Geophysique for the Schlumberger configuration.* Part (b) depends upon supplemental data. *Comparing actual resistivity raviations with depth to* data from a nearby logged test hole enarby logged test hole enables a correlation to be established with subsurface geologic and ground water conditions. This information can then be applied for interpretation of resistivity measurements in surrounding areas.

Fig. 9.3a shows a graph of apparent resistivity and electrode spacing, while Fig. 9.3b gives the interpretation of the measurement. Depths and actual resistivities were determined by matching with Schlumberger curves. Geologic data were obtained from nearby test holes, but estimated chloride contents of the groundwater were based upon a relation between actual resistivity of the saturated aquifer and resistivity of the groundwater obtained from measurements at surrounding test holes. It should be noted that the resistivities here are unusually low because of severe sea water intrusion; values 10 to 100 times higher can be expected under ordinary conditions.

Resistivity surveys can cover either vertical investigations at selected locations by varying electrode spacings, or they can furnish isoresistivity maps of an area. In areal studies a constant spacing may be adopted to measure resistivities only at a particular *depth of interest, such as an aquifer. Areal resistivity changes can be interpreted in* terms of aquifer limits and changes in groundwater quality, whereas variable depth surveys may indicate aquifers, water tables, salinities, impermeable formations, and bedrock depths.

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* Also, curves for the Schlumberger arrangement can be computed from tables of potentials about a point electrode in Mooney and Wetzel.

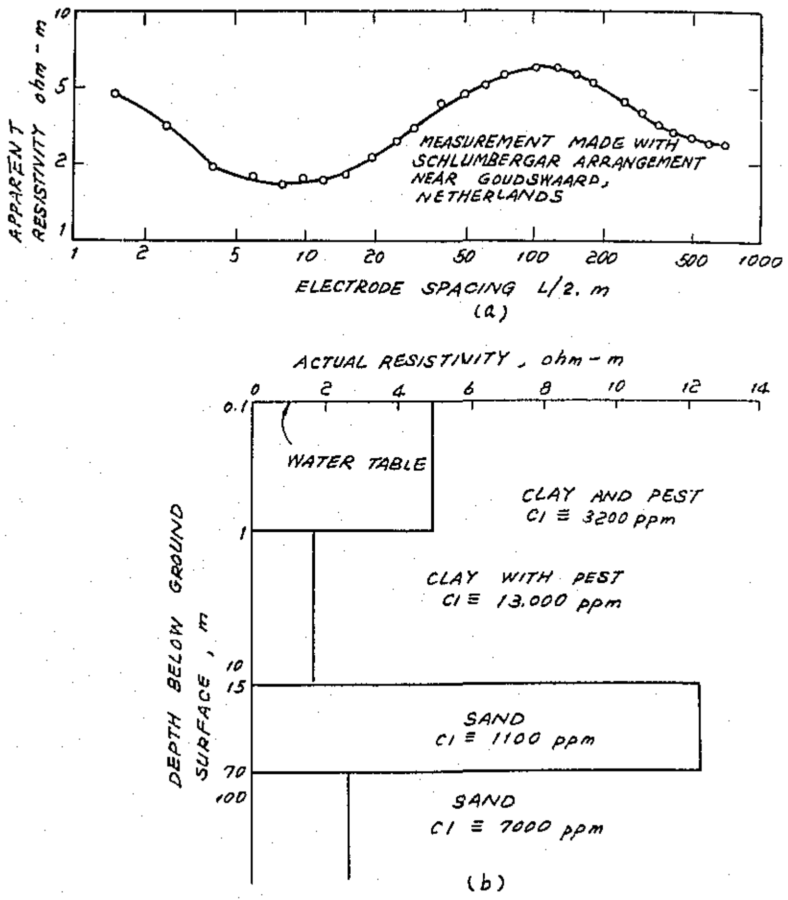

Fig. 9.3 Example of electrical resistivity measurement showing (a) results and (b) interpretation

Any factors which disturb the electrical field in the vicinity of the electrodes may invalidate the resistivity measurements. Buried pipe lines, cables, and wire fences are common hazards. Extremely dry conditions may necessitate moistening the ground around electrodes to establish proper earth contact.

Of all surface geophysical methods, the electrical resistivity method has been applied most widely for groundwater investigations. Its portable equipment and ease of operation facilitate rapid measurements. The method is often helpful in planning efficient and economical test drilling programs. It is especially well adapted for locating subsurface salt water boundaries because the decrease in resistance when salt water is encountered becomes apparent on a resistivity-depth curve.

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* Also, curves for the Schlumberger arrangement can be computed from tables of potentials about a point electrode in Mooney and Wetzel.

Resistivity Logging

Within an uncased well, current and potential electrodes can be lowered to measure electrical resistivities of the surrounding media and to obtain a trace of their variation with depth. The result is a resistivity (or electric) log. Such a log is affected by fluid within a well, by well diameter, by the character of surrounding strata, and by groundwater.

Of several possible methods for measuring underground resistivities, the multielectrode method is most commonly employed, as it minimizes effects of the drilling fluid and well diameter and also makes possible a direct comparison of several recorded resistivity curves. Four electrodes, two for emitting current and two for potential measurement, constitute the system. Recording Curves are termed normal or lateral depending on the electrode arrangement as a shown in Fig. 10.3. In the normal arrangement the effective spacing is considered to be the distance AM (Fig. 10.3a) and the recorded curve is designated AM. The spacing for lateral (AO) curves is taken as the distance AO, measured between M and a point midway between the electrodes A and B (Fig. 10.3b). Sometimes a long normal curve (AM') is recorded based on the same electrode arrangement as the normal but with the distance AM increased several times. Similarly, a long lateral curve (AO') has a longer AO distance than the regular lateral. Boundaries of formations having different resistivities are located most readily with a short electrode spacing whereas information on fluids in permeable formations can be obtained best with long spacings.

An electric log of a well usually consists of vertical traverse that record the normal and lateral, possibly either or both the long normal and long lateral, and the spontaneous potential curves. An illustration of an electric log is given as Fig. 10.4. Accurate interpretation of resistivity logs is difficult, requires careful analysis, and is best done by specialists.

Resistivity curves indicate the lighlogy of rock strata penetrated by the well and enable fresh and salt waters to be distinguished in the surrounding material. In old wells exact locations of casings can be determined. Resistivity logs may be used to determine specific resistivities of strata, or they may indicate qualitatively changes of importance. As mentioned in the previous chapter, resistivity of an unconsolidated aquifer is controlled primarily by the porosity, packing, water resistivity, degree of saturation, and temperature. Although specific resistivity values cannot be stated for different aquifers, on a relative basis shales, clays, and salt water sands give low values, fresh water sands moderate to high values, and cemented sandstones and nonporous limestones high values. Of course, casings and metallic objects will indicate very low resistivities. Correlation of rock samples, taken from wells during drilling, with resistivity curves furnishes a sound basis for interpretation of curves measured in nearby wells without available samples.

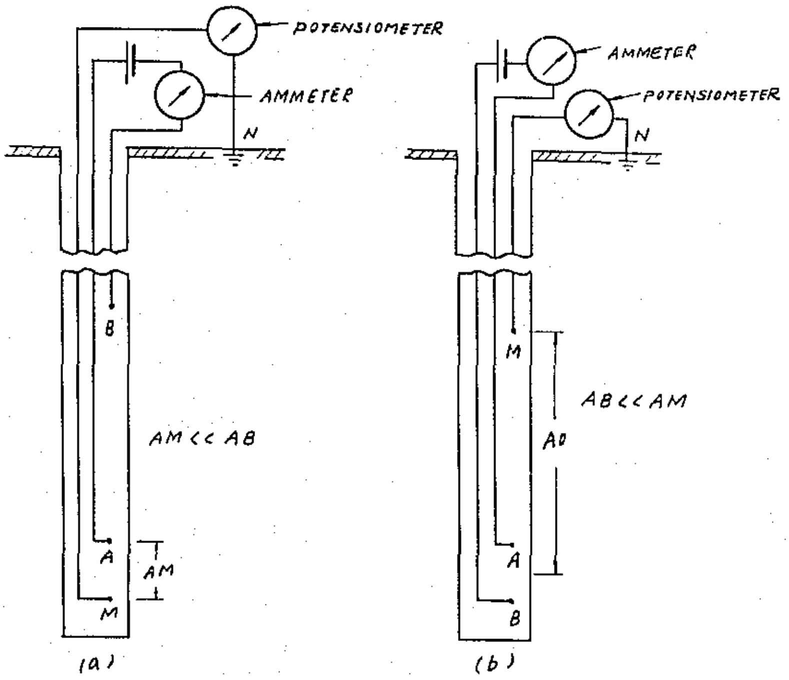

Fig. 10.3 Electrode arrangements for recording (a) normal and (b) lateral resistivity well logs

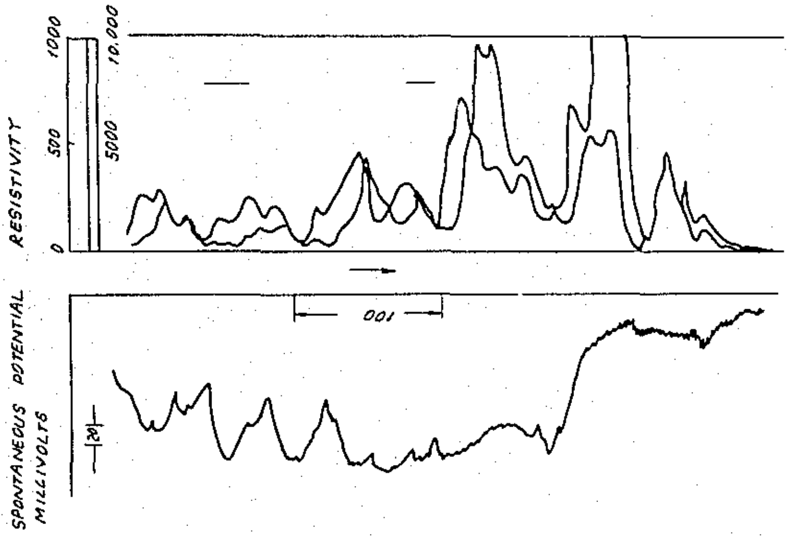

Fig. 10.4 Spontaneous potential and resistivity logs of well (courtesy Schlumberger Well Surveying Corp.)

Potential Logging

The potential method measures natural electrical potentials found within the earth. Potentials are referred to also as self-potentials, spontaneous potentials, or simply "SP". Measurements, usually in millivolts, are obtained from a recording potentiometer connected to two like electrodes. Usually one electrode is lowered into the well and the other connected to the ground surface, as illustrated by electrodes M and N in Fig. 10.3.

Interpretation of potential logs to obtain maximum information is difficult and should be done by specialists. The exact nature of phenomena giving rise to observed potentials is not entirely known. Electrofiltration resulting from the flow of drilling mud into surrounding formations or of water through porous media and electrochemical potentials arising from fluid concentration differences, as between drilling mud and groundwater, are believed to be the major causes. Energy from chemical reactions occurring at formation boundaries, oxidation, and pH gradients also have been suggested as possible causes. Potential values range from zero to several hundred millivolts; positive values occur with flow from the formation into the well, negative values for the reverse flow. Potential logs are read in terms of positive and negative deflections from an arbitrary base line which might be either a permeable or impermeable formation of considerable thickness. Therefore, potential logs indicate permeable zones but not in absolute terms. Frequently, changing the natural water level in a well will modify flows and also potential values, thereby enabling better definition to be achieved of permeable zones. In areas where there are no sharp contrasts in permeable zones, potential curves lack relief and are of little value. In urban and industrial areas, spurious earth currents may occur, such as from electric railroads, which interfere with potential logging.

Spontaneous potentials resulting from electrochemical potentials can be expressed by

$$SP = M \frac{\rho f}{\rho w} \quad \cdots\cdots\cdots\cdots\cdots\cdots\cdots\cdots \quad (10.2)$$

where ρf is the drilling fluid resistivity in ohm-m, ρw is the groundwater resistivity in ohm-m, and M is a factor dependent upon the chemical composition of the two fluids and upon the character of the formations adjoining an aquifer. Knowing M and measuring SP and ρf enables estimates of the groundwater resistivity to be computed.

In practice, potential and resistivity logs are usually recorded together as shown in Fig. 10.4. The two logs often indicate the same subsurface conditions and thereby supplement one another; however, occasionally the pair of logs will furnish information not available directly from either alone.

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PHOTOGEOLOGY

PHOTOGEOLOGY

by Dr. Kyuya Matsuno
Geological Survey of Japan

AERIAL PHOTOGRAPH

General statement: The aerial photograph is an instantaneous record of the ground details as determined chiefly by the focal length of the camera lens, the flying height of the airplane at the time of exposure, and the film and filter used.

I. Classification of aerial photograph

A. Vertical photograph – taken with the camera axis pointing essentially vertical down.

B. Oblique photograph – taken with the camera axis purposely tilted from vertical, generally 20 degrees or more.

Note: The following discussion is limited to techniques applicable to the study of vertical photography.

II. Terminology and geometry of the aerial photograph

A. Term (see Fig. 1)

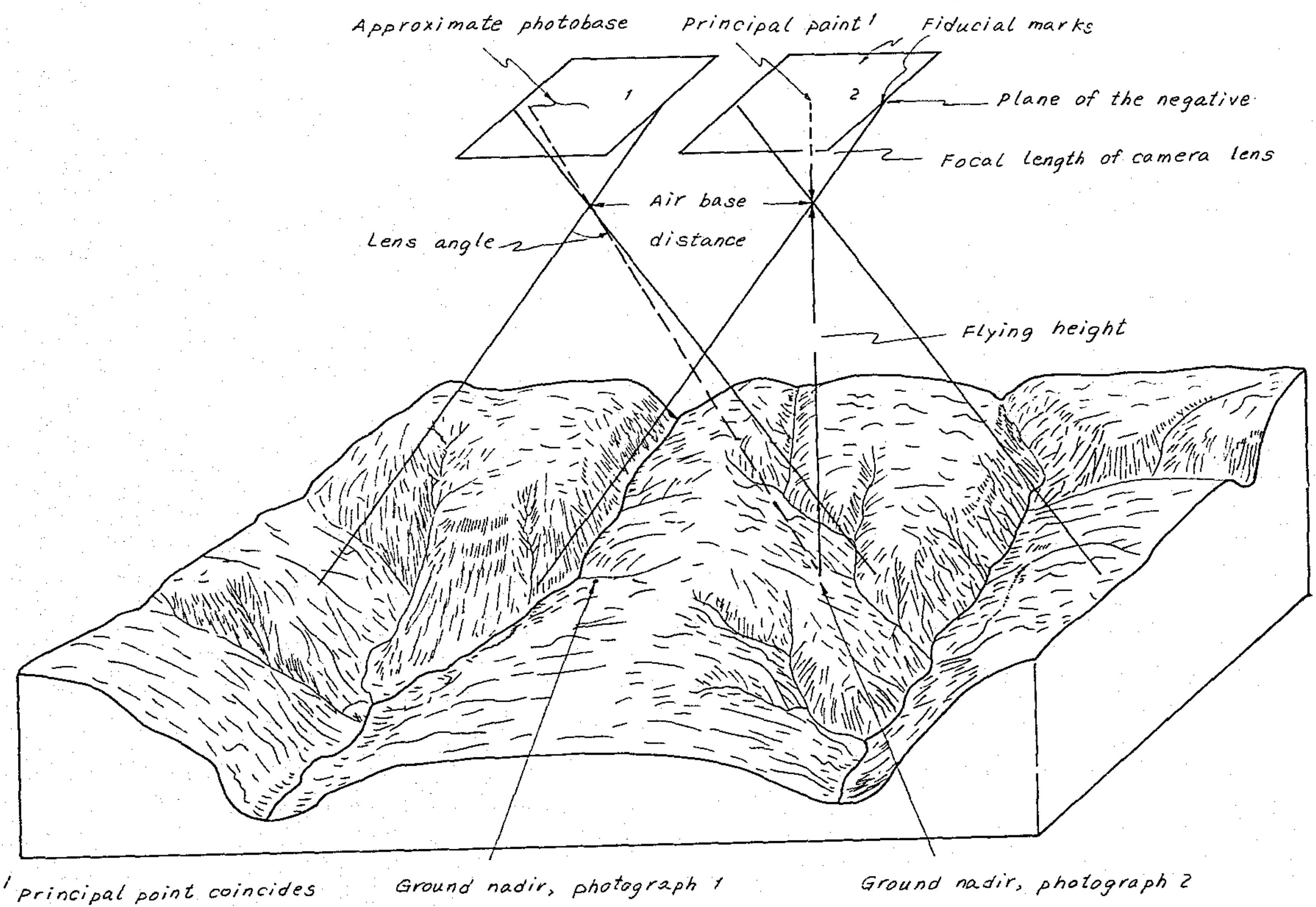
1. Fiducial mark
2. Principal point
3. Photograph nadir
4. Ground nadir
5. Iso-center
6. Air base distance
7. Approximate photo base
8. Flying height
9. Focal length
10. Lens angle or angle of view

B. Geometry of the aerial photograph

1. Average scale of a photograph (S)

$$S = \frac{\text{Focal length (meters)}}{\text{Flying height (meters)}} = \frac{f}{H}$$

Note: If a photograph is enlarged or reduced, the "effective" focal length for that photograph is also changed in direct proportion to the amount of enlargement or reduction.



[1] Principal point coincides with photograph nadir in truly vertical photograph

Fig. 1 Geometry of the vertical aerial photograph

2. Radial displacement or relief displacement – When an object is photographed from two different positions, as on two overlapping vertical aerial photographs, an apparent shift in the position of the object takes place to a radial direction, which is known as parallactic displacement. This radial displacement is directly related to the height of the object.

$$h = \frac{mH}{r}$$

h = difference in altitude desired;

m = relief displacement of upper image point with respect to the lower image point;

H = height of the airplane above lower image point;

and

r = radial distance from principal point of photograph to lower image point.

III. **Film and filter** – will be discussed in the following chapter.

VIEWING THE PHOTOGRAPHS

General statement: May be viewed individually, as mosaics or as stereo-pairs (brief explanation)

I. **Two dimensional viewing** – single prints, mosaics.

   A. Advantages

   B. Disadvantages

II. **Three-dimensional viewing**

   A. How we see three-dimensionally

   B. Instruments used to aid three-dimensional viewing

      1. *Discussion of stereoscopes*

         a. Construction (arrangement and function of parts)
         b. Operation and care

      2. Brief mention of stereoplotters (thoroughly discussed forth coming outline)

   C. Advantages of three-dimensional viewing

   D. "Naked stereo"

   E. Pseudoscopic viewing with reversed positions of stereo pair to give reverses relief.

III. **Vertical exaggeration** – the exaggeration of vertical distances with respect to horizontal distances. Vertical exaggeration in normal viewing arrangement is fundamentally related to base height ratio, which is the ratio of the air base distance to the flying height. As this ratio increases, vertical exaggeration increases.

PHILOSOPHY OF PHOTOGEOLOGIC INTERPRETATION

I. Observation of data

Photointerpretation is defined by Colwell " ..... the act of examining photographic images of objects for the purpose of identifying the objects and deducing their significance."

Photogeologic interpretation is only applicable to those geologic features that develop surface expression. Features of surface expression are identified by mean of (1) basic recognition criteria and (2) interpretation criteria.

A. Basic recognition criteria

1. Photographic tone
2. Colour
3. Texture
4. Pattern
5. Shape
6. Size
7. Combination of recognition criteria

Note: From basic recognition criteria, we distinguish surface expressions of geologic features directly or we may distinguish interpretation criteria.

B. Interpretation criteria

1. Landform
2. Drainage
3. Vegitation
4. Culture

Buth basic recognition criteria and interpretation criteria are used to distinguish or interpret geologic features of surface expression; and depend upon the individual experiences.

II. Processing of data

The following are interrelated, although they are discussed separately.

A. Deductive and/or inductive method

B. Observations and interpretations considered in terms of "order" of reliability

C. Empirical largely observation comparison to key and rational interpretation by inductive and deductive means method of processing data

D. Convergence of evidence principle combined use of vegitation, soils, and geological features empirically derived, to solve interpretation problems

Summary: Thus interpretation may be considered as two step process:

1) Observation, fact-gathering, measurement, identification of features
2) Processing these data (of step 1) by deductive and inductive means.

APPLICATION OF PHOTOGEOLOGY TO GENERAL GEOLOGIC MAPPING

General statement: Depend upon climatic and geologic condition, the geologist may use aerial photographes: (General uses)

1. For logistic purposes (location, access, traverses, orientation)
2. As field base maps (mention of photo mosaic and orthophoto)
3. As medium for distinguishing "photogeologically mappable units"
4. As medium for observing and/or interpreting lithologic and structural character of rocks, relating this information to field *investigations.*

Quantity of geologic information controlled by climatic and geologic environment.

A. Climatic environment

   1. Arid regions

      a. Except for high mountains, more rock exposed in arid regions sparse rainfall *and mechanical weathering conductive to well-exposed bedrock.*

      b. Geo-botanical relationships

   2. Tropical regions = where vegitation is dense, structural information will tend to be primary.

      a. Owing to extensive leaching, soils derived from widely different parent rocks tend to be similar at maturity.

      b. Geo-botanical relationships are less significant than in arid regions because soils tend to be chemically similar at maturity. Exception: These plants that depend on physical differences in soils.

      c. Surface may be deeply weathered; hence, extensive outcrops are not common.

B. Geologic environment – Where surficial debris is widespread, structural information will be primary.

   1. Geomorphic cycle of erosion (youthful, mature, old age)

   2. Rock type (listed in order of greatest accommodation to interpretation)

      a. Sedimentary

      b. Igneous

         i. Extrusive
         ii. Intrusive

      c. Metamorphic – Metamorphic processes tend to destroy erosional and landform *characteristics of the sedimentary and igneous rocks.*

QUANTITY AND TYPE OF GEOLOGIC INFORMATION AVAILABLE FROM PHOTOGRAPHS

General statement: Type of geologic information (depend on climate, terrain, and erosional cycle)

A. Lithologic

B. Structural

I. **Lithologic character** – Observations and interpretations are based on basic recognition criteria and/or interpretive criteria.

A. Sedimentary rocks

1. Consolidated

a. Bedding is fundamental

1. Topographic expression or the "etching process"
2. Cycle of erosion

Note: Massive character may mas bedding

b. Grain size

1. Fine grained rocks

a. Dark tone
b. Fine textured drainage
c. Close regularly spaced joints

2. Coarse grained rocks

a. Light tone
b. Coarse drainage texture
c. Wide regularly spaced joints

Note: There are exceptions to above generalizations. Also drainage is more significant as an indicator of structure than rock texture.

2. Unconsolidated or surficial – Interpretation of this groups is interpretation of constructional features.

a. Landform is fundamental
b. Position

B. Igneous rocks

1. Extrusive rocks – Only relatively underformed extrusives (mainly younger tertiary or younger) are readily interpreted from photographs without aid of field survey.

a. Landform is commonly fundamental

- b. Position and relation to other rocks
- c. Vegitation
- d. Associated features – characteristic of extrusives
  1. Flow channels
  2. Surface character of flow
  3. Cones, calderas
  4. Terracette appearance of termini of multiple flows
  5. Steptoes
  6. Lava cascades
  7. Basal contact of flow commonly unconformable with country rock
  8. Abrupt termination of faults and joints system at flow contact
- e. Differences in appearance that may distinguish basic from acidic extrusives.

2. Intrusive rocks
   - a. Dikes
     1. Linear ridges and depressions
     2. Tonal contrast with country rock
     3. Distinctive vegitation
   - b. Sills – commonly more difficult to interpret that dikes
     1. Tonal contrast
     2. Distinctive vegitation
     3. Commonly distinctive angular rubble
   - c. Plugs, stocks, laccoliths and batholiths
     1. Tilt or cut sedimentary rocks they intrude
     2. Radial drainage in domed area
     3. Dendritic drainage in large intrusive areas
     4. Massive appearance
     5. Uniform plant distribution
     6. Multiple joint sets intersecting at varying angles contrast with joints in sedimentary and metamorphic rocks

C. Metamorphic rocks – Interpretation of bedding, fundamental in structural analysis, is commonly difficult because of the physical changes brought about by the metamorphism.

1. Comparison of photo appearance quartzite and marble with their sedimentary equivalents – Same physical appearance on photograph's tone, same topographic expression.

      a. Talus
      b. Vegitation
      c. Tone

   2. Topographic appearance may suggest metamorphic rocks.

II. **Structural character** – Observations and interpretations are based on basic recognition criteria and/or interpretive criteria.

A. Folds

- 1. Photograph evidence of folds. Landform – tone – streams
- 2. Photograph mediums for interpreting folds photos and mosaic
- 3. Estimation of dips
- 4. Measurement of dips

B. Faults

- 1. Photographs are conductive to delineating faults, particularly high angle faults
- 2. Fractures that may be indicative of faults alignments
    - a. Vegitation alignments
    - b. Straight streams
    - c. Alignment of lakes and ponds
    - d. Tone changes
    - e. Topographic alignment
    - f. Offsets
- 3. Fallacious interpretation of faults
    - a. Lineaments observable in any direction upon suggestion or sustained searching
    - b. Features misinterpreted as faults
        - 1. Flood plain margin
        - 2. Jointing
        - 3. Channel fill
        - 4. Vegitation alignments or differences
        - 5. Eroded dipping strata
        - 6. Margin of lava flows
- 4. Low angle faults
- 5. Statistical methods and fracture analysis
- 6. Active faults – Topographic gaps which are appeared on the surface of constructional features particularly.

C. Joints – may be particularly conspicuous on aerial photographs through widening effect of weathering and erosion. Commonly, vegitation is comparatively dense along joints.

1. Sedimentary rocks
    a. Flat-lying or gently dipping
        1. Abundant short lineation regularly spaced at more or less constant trend. Joints are vertical or near vertical.
        2. Commonly two sets are present intersecting at right angles.
        3. More closely spaced in fine grained rocks than in coarse grained rocks.
    b. Joints are indicator of steepness of dip.
2. Igneous rocks
    a. Intrusive – Three or four joints sets of variable dip and irregular spacing intersecting each other at widely varying angles. This pattern, contrasted to joints in sedimentary rocks is quite distinctive.
    b. Extrusives – Jointing in basalt sometimes distinctive on photographs.
3. Metamorphic rocks
    a. Pre-metamorphism joints tend to be destroyed.
    b. Tendency for small structures to develop rather than joints.
    c. Joints accentuated by streams or stream pattern have been observed. Contrast with expression of foliation.

D. Cleavage – Flow cleavage may be reflected by alignment in topographic grain or "micro relief".

E. Unconformities

1. Discordance of bedding
2. Channeling

RADIAL PLOTTING PROCEDURE FOR A SCALE SOLUTION

I. Objective of radial plot

1. All information must be plotted at the same scale.
2. Base map compilation
    - a. No good map of the desired scale exists of the area of interest.
    - b. A set of photos giving complete coverage of the area of interest exists.
    - c. Within the area are insufficient ground control points for a successful stereo-bridging but enough for radial triangulation.

II. **Basic principle involved in radial plot** – is that true and constant angles will be formed at the center of a truly vertical aerial photograph by radiating lines drawn to ground point in the periphery.
Thus: With vertical photographs (no tilt) with nadir points as vertex, these angles remain constant regardless of scale change and/or amount of relief. However: If tilt is present, problem is complicated by fact that image displacement owing to tilt radiated from isocenter (not nadir point). Most present day photos have less than three degrees tilt so far most purposes may be considered as vertical.

III. Graphic radial line method

1. Locate principal point on photos (red circles)
2. Transfer centers to adj. photos in same strip
3. Pick at least three other pts. (pass point) on each side of photo neat model in sidelap area between fleights. (total 9 pts. per photo)
4. Pick ground control pts. (yellow circles) prick photos and encircle. (red-azimuth, blue – azimuth, blue – pass pts., yellow – ground control pts.)
5. Make templets with clear acetate
    - a. Cut in pieces ca. 3 ~ 5 centimeters larger than photos so photo can be taped face down to it.
    - b. Turn over assembled templet-photo, center pt. marked and radials drawn through pass pts. each lines for accuracy (pin and ink)
        - 1) Azimuth lines solid .
        - 2) Radials short lines through pass pts. for 3cm or so
        - 3) Number all points
6. Remove templets from photos
7. Assemble templets on base sheet on which all round control pts. were previously plotted in relation to a prepared grid.
    - a. Assembly begins at plotted control on base sheet. If 2 of more pts. are in overlap area of adjoining templets (of a strip) scale and orientation can be established at once.

b. Laydown begins with two adjacent with the greatest number of control pts. laid in such position that azimuth lines coincide, and radials to control pts. intersect at plotted position to control pts. on base sheet. Fasten to base and each other with tape. Continue with additional templets.

c. Discrepancy at end of fleight line is proportioned through whole distance by adjustment throughout. (personal judgement)

Note: Errors may be cumulative so with lg no. of templets proportioning errors will be difficult if not impossible.

d. After completed, turn over whole assembly and mark pts. on bottom of transparent base.

DETERMINATION OF ALTITUDE DIFFERENCES BY STEROMETER

General statement: To obtain the height of an object or difference in altitude between two points the difference in parallax must be related to the geometry of the stereoscopic model.

Measuring height from paper prints – for measuring height difference from paper prints the above relation is best expressed by a formula which, in its simplest terms, is

$$h = \frac{H}{b} \cdot \Delta P \qquad \text{..............................} \qquad (1)$$

where h = height of the object, in meters, or difference in altitude being determined;

H = height of airplane, in meters, above mean terrain (determined from specifications of photographic mission or relation $H = f/S$, see below);

b = photobase (commonly determined by averaging the distances between the center and conjugate center of each photograph of the stereoscopic pair); units of measure may be millimeters must be the same as ΔP;

and p = paralax difference, in millimeters or inches, as determined with stereometer-type instrument (distance A-B of Fig. 2).

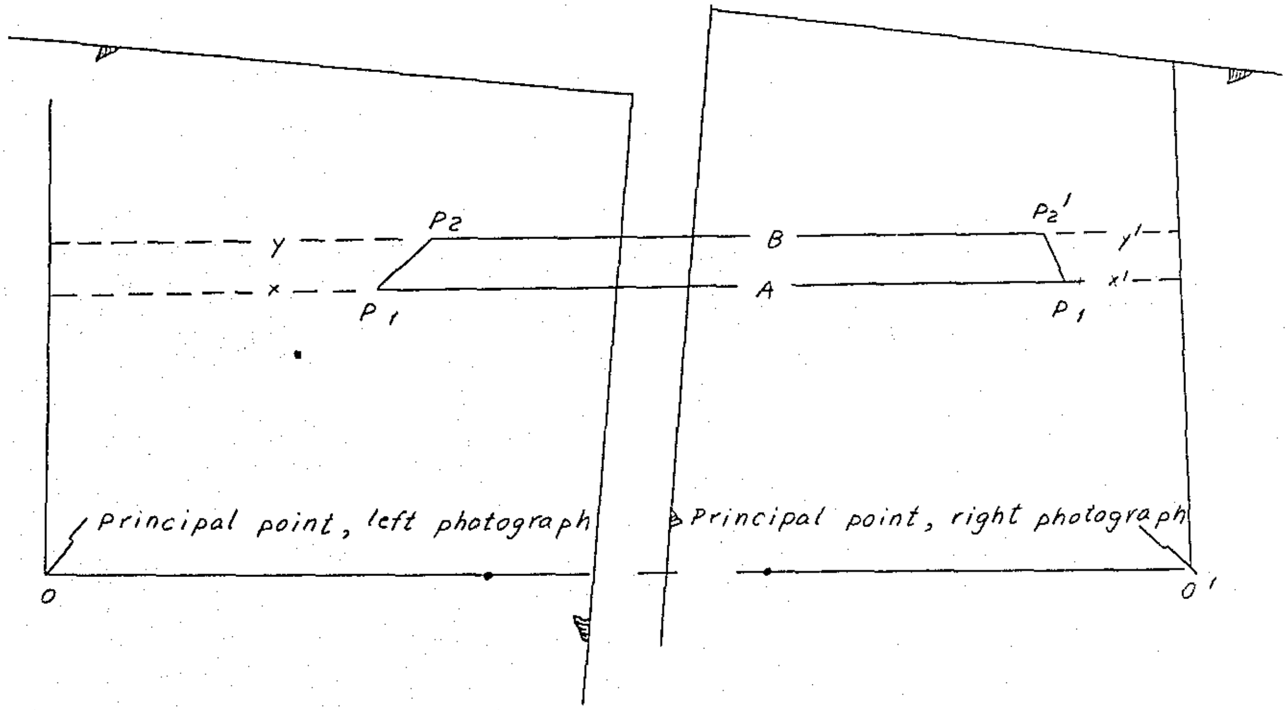


Fig. 2 Diagram showing relation between absolute stereoscopic parallaxes and horizontal distances actually measured with stereometer-type instruments in determining differences in altitude from paper prints.

Let images $P_1 - P_2$ and $P_1 - P_2 1$ represent the photographic expression of a pole on the left and right photographs of a stereoscopic pair. The absolute stereoscopic parallax of the base of the pole is $x - (-x') = (x + x')$. The absolute stereoscopic parallax of the top of the pole is $y - (-y') = (y + y')$. (Distances measured to the right of the principal point are positive, to the left, negative). The parallax difference between the top and bottom of the pole then is $(y + y') - (x + x')$.

From the figure:

$$y + y' + B = x + x' + A$$
$$y + y' - x - x' = A - B$$
$$(y + y') - (x + x') = A - B$$

Thus the difference in absolute stereoscopic parallaxes is equal to A – B. The distances A and B are the actual distances measured with stereometer-type instruments, generally in hundredths of millimeters, in determining altitude differences from paper prints.

It will be noted that as H increases, the measure of p for any given vertical interval will decrease; however, the absolute value of any one unit of ΔP will increase correspondingly. On the other hand, if focal length is increased and flying height remains constant the measure of ΔP for a given vertical interval will increase but the absolute value for any one units of ΔP will decrease correspondingly.

The above formula (1) may be used without any appreciable error if relief in an area is low; p will be small. If relief in an area is high one of the following two formulas, (2) or (3), should be used.

$$h = \frac{H'}{ab + \Delta P} \cdot \Delta P \qquad \ldots\ldots\ldots\ldots\ldots\ldots\ldots \qquad (2)$$

where h = height of the object, in meters, or difference in altitude being determined;

H′ = height of airplane, in meters, above the lower of the two points whose paralax difference has been measured;

ab = photobase adjusted to the lower of the two points whose paralax difference has been measured (commonly determined by the measuring the distance between photograph centers – 0-0′ of Fig. 2 – and subtracting from it the distance between conjugate image points at the lower altitude – distance A of Fig. 2); units of measure may be millimeters or inches but must be the same as ΔP.

and p = paralax difference, in millimeters or inches, measured with sterometer-type instrument (distance A-B of Fig. 2);

or

$$h = \frac{H'}{ab + \Delta P} \cdot \Delta P \qquad \ldots\ldots\ldots\ldots\ldots\ldots\ldots \qquad (3)$$

where h = height of the object, in meters, or difference in altitude being determined;

H′ = height of airplane, in meters, above the upper of the two points whose paralax difference has been measured;

ab = photobase adjusted to the upper of the two points whose paralax difference has been measured (commonly determined by measuring the distance between photograph centers – 0-0 of Fig. 2 – and subtracting from it the distance between conjugate image points at the upper altitude – distance B of Fig. 2); unit of measure may be millimeters or inches but must be the same as ΔP;

and ΔP = paralax difference, in millimeters or inches, measured with stereometer-type instrument (distance A-B of Fig. 2)

Formula (2) is used when H' and ab are determined by referring to the lower of the two points whose altitude difference is desired, as in measuring upward from the base to the top of a cliff. Formula (3) is used when H' and ab are referred to the upper points, as in measuring downward from the top to the bottom of a cliff. Most commonly the lower of two points is selected as the reference point and measurement are made from that point to some higher point in the stereoscopic model; thus formula (2) is more widely used than formula (3).

RADIOISOTOPE PROSPECTING
FOR
GROUNDWATER

RADIOISOTOPE PROSPECTING FOR GROUNDWATER

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1. Introduction

Concerning the Radioisotopes utilization in the field of agricultural water utilization, it is the most efficient way to trace the behavior of water as tracer taking advantage of the properties of RI. Therefore, RI is being widely used for this purpose. It is needed to investigate the water resources condition and to clarify the mechanism of water balance in each district to promote the utilization of water resources and for proper utilization of irrigation water for agricultural use. On the other hand, it would be important to feed the groundwater for irrigating paddy field, to increase the water resources by means of the artificial recharge of groundwater, and to use irrigation water repeatedly taking advantage of percolation in the basin. In these cases, it was difficult to clarify the route and flow velocity of percolation water by the traditional way. It would be possible to trace the behavior of water clearly, if RI is used.

The calculation of the amount of groundwater flow is being conducted by using a tentative radioisotope flow meter, radioactive logging instrument together with the routing with RI.

Specific yield of water in aquifer is obtained by $\gamma - \gamma$ logging and neutron logging methods, followed by the water yield test for aquifer conducted with RI, to promote the effective use of groundwater.

Groundwater is being obtained in priority by the natural radioactivity survey in order to develop the new groundwater source at the foot of volcanoes and plateaus for the irrigation resources. This method is conducted by using natural radioactive element without using radioisotopes and has contributed the development and conservation of water resources, together with the dating of groundwater age with natural tritium.

In connection with the countermeasure for acid water in irrigation water, percolation route into ground of acid water and flow velocity can be easily investigated. Therefore, tracing survey by RI is helpful in establishment of basic planning of acid-water prevention works.

In addition, various measuring methods by RI utilization to secure economics and safeness about the direct agricultural water utilization works such as leakage routing for dams, reclamation dikes and water canals, compaction for earth dams, density measurement of asphalt, determination of the amount of cement for injection, etc.

2. Natural radioactivity survey for groundwater veins

The groundwater in the hills, volcanic slopes and mountainous zones exists in diluvial deposits or rocks of Tertiary system. It mainly exists in waterveins type aquifers or in fissures in the bed rock, so that the survey for groundwater has been considered to be difficult. The water in the underlaying Tertiary system should be developed as irrigation water and industry water even in the paddy field zones of thin alluvial deposits found in such districts as Hokkaido, Tohoku District, Boso Peninsula, Hokuriku District, etc., not only in mountainous districts. In addition, if irrigation water for farm irrigation is to be newly developed, the fissures developed in the rock and the fissure water flowing in the pores would be only water resources.

The fault is formed due to the internal stress of tectonic movement, numerous fissures develop and the zones with high permeability also develop into the deep part of the ground.

Near the tectonic lines, uranium salts or radium salts in the deep ground rise through the fractured zones along faults with high permeability and then deposit in the soil near the surface. Radon rises through the cracks, and is scattered in the soil accompanied by disintegration, and then a kind of balance is maintained. Therefore, the intensity of the natural radioactivity is considered to increase more than in other places. The author measured the intensity of natural radioactivity in the fault zones at about 80 places in Japan. It was clarified that the intensity of natural radioactivity increases up to 1.3 to 2.0 times as compared with the background.

Such zones of strong natural radioactivity was investigated hydrogeologically, and then it was verified by the author that the zones accompany water veines.

Investigation was made at the east slope of Mt. Fuji which was considered to be one of the representative water deficient zones in Japan. Three scintillation counters ware set on the jeep and then the counts were self-recorded while running with the constant speed of 10 km/hr in the whole area (Fig. 1). As the results, five radioactivity zones were found as in Fig. 2. The width of the strong radioactivity zone reached 250 m in the lava areas and 400 m in the mud stream areas. The zones were detected for about 25 km in length. The zones have such hydrogeological structures that the groundwater veins.5)

The total amount of pumped water of 24 wells in these high radioactivity zones reaches 3,400 to 13,000 m^3/day and shows the better results as compared with the amount of pumped water less than 3,400 m^3/day of the deep wells in other areas.

It was clarified that there is an exponentially increasing relationship between the permeability coefficient and the intensity of the natural radioactivity.

Accordingly, it is confirmed that the groundwater is more abundant in the fractured zones along faults where the intensity of the natural radioactivity is high.

3. Utilization of natural tritium for measurement of groundwater movement in large areas

The utilization of natural tritium plays the most important role in surveying large areas for the development of groundwater resources. The flow velocity and flow distribution of the groundwater in a large water system could not be directly calculated by other methods in the past. The determination of ages of the groundwater, however, became possible by using natural tritium. Also, the movement velocity of groundwater can be obtained by the distance of seepage from the groundwater supplying zone, and conservation conditions and existence mechanism of the groundwater in the large areas can be investigated.

At present, the concentration of natural tritium is obtained by the following way: 10l-groundwater in the deep layer intercepted from the shallow by clay layer is concentrated into one-hundreds by 2-step electrolysis after distillation; its tritium concentration is measured by the liquid scintillation counter.

The tritium unit of the groundwater in Kanto Plain showed 2 to 406 T.U. Therefore, it was clarified that the groundwater immediately before the test of hydrogen bomb in 1954 flow towards the Tokyo Bay and reached as far as the suburbs of Tokyo City (Fig. 3).

In the flow of groundwater in such large areas is investigated, combined utilization of groundwater would be possible.

4. Measurement of groundwater flow by using RI

RI is used to obtain detailed data on the groundwater flow such as flow distribution of the groundwater, water-holding mechanism, flow velocity of the groundwater, etc. The following tracing method is applied for this purpose:

a. Multi-well tracing method
b. Single-well method (RI flow meter)

a. Multi-well tracing method

A self-recording scintillation counter with 10 scintillation probes was tentatively made (Fig. 4). Radioactivity in the water of 10 places is being self-recorded by putting probes into the observation well. The isotope mainly used in this case is ^{131}I.

At first, the actual velocity V_0 is obtained by tracing RI and the effective porosity β is obtained by $\gamma - \gamma$ logging. Then, the apparent velocity is obtained by $v = V_0\beta$. Lastly, the groundwater flow Q is obtained by $Q = AV$ ..... (1).
where, A: cross sectional area.

At present, ^3H is considered to be an ideal tracer of the groundwater. However, its radiation is β-rays and the maximum energy as small as 0.018, MeV. Therefore, the measurement in the field is difficult to conduct and its utilization is undeveloped. Accordingly, if there exists such radioisotopes that flow in the aquifer and can easily be detected as well as ^3H, it would be efficiently used instead of ^3H. The author compared ^{131}I and ^3H as the double tracer in the gravel layers in alluvial fan. Their runoff curves accorded perfectly as shown in Fig. 7, so that it was verified that ^{131}I has no lag as compared with ^3H (Fig. 5).7)

The scattered angle and dilution ratio (including sorption loss) would be important problems in the multi-well tracing method. The scattered angle of the tracer was expressed to be about 3° with the flow axis according to Danel. It showed 10° in maximum in the gravel layer and mostly within 6° according to the field tracing test of ^{131}I up to now. However, the fissure water in Andeside shows 23° which is very large. Now, the dilution ratio of the tracer Df is expressed Co/C (Co being intensity of radioactivity injected, C intensity of radioactivity of the tracer peak detected in the downstream).

The relationship between the dilution ratio Df and tracing distance ℓ was expressed as below after carrying out the field tracing test of ^{131}I in the alluvial plain, alluvial fan, etc.:8)

$$Df = 7.6\ell^{3.6}$$

The detection distance and the intensity of original RI solution can be previously estimated using such a relationship.

b. Single-well method (Using RI flow meter)

A RI flow meter that the flow velocity of the groundwater can be estimated by using an observation well was developed.9-10)

The principle is to measure the flow velocity and flow direction of the groundwater simultaneously from the dilution ratio of RI solution flowing out of 3 RI chambers of the counting apparatus (Fig. 6).

The relationship between the flow velocity of the groundwater V and the dilution ratio D is expressed as below;

$$D = -6.72V + 0.88$$

It was clarified that the flow velocity of the groundwater in the upper layers was larger than that in the lower layers when it was measured at the southern west slope of Mt. Fuji by using this apparatus. The amount of groundwater flow is obtained from Formula (1) by using the radioactivity logging.

5. Survey of aquifer by means of radioactive logging

The percolation capacity is generally expressed with the permeability of aquifer. The possible amount of water sampling increases more and shows higher capacity of percolation, if the permeability of aquifer is larger.

The intensity of scattered γ-rays in various layers of alluvium and diluvium is shown in Fig. 7. That is, the composed particle becomes rough in the order of mud, clay, clay sand, and gravel. The intensity of scattered γ-rays decreases in accordance with the increase of the groundwater flow quantity (In this case, the porosity becomes small). The intensity of scattered γ-rays increases if the aquifer becomes impervious (In this case, the porosity becomes large).11-12) The percolation capacity and porosity of the layer can be obtained by the $\gamma - \gamma$ logging as mentioned above.

The porosity of aquifer is obtained quantitatively by the following formula:

$$P = \frac{100\ (gs - ga)}{gs - gw}$$

where, P: Porosity (%)
gs: Density of Particles (g/cc)
ga: Density of deposits saturated by groundwater (g/cc)
gw: Density of groundwater placed in the porosity (g/cc)

ga is obtained by the $\gamma - \gamma$ logging and gs is obtained by the picnometer method. The pulse wave height analyser is used in obtaining ga, because the photo-electric region less than 100 KeV is needed to be cut in accordance with the density.

When the porosity of the test boring of 300 m in depth was measured at Okuma-machi, Fukushima Pref., the good accordance was obtained with the difference within 5% as compared with the porosity by means of drying method of boring core of sandstone and shale.

When the specific yield of sandy gravel was measured, Comparison was made between the traditional method and $\gamma - \gamma$ logging and n-n logging. The n-n logging showed the good accuracy with the error of 0.6% and $\gamma - \gamma$ logging had the error of 10 ~ 16%.

The selective $\gamma - \gamma$ logging is used for the investigation of the fissure water. This method is carried out as follows: Probe wall of the crystal part of scintillation inspector is made by A1 and low energy region only is counted by the pulse wave height analyser without absorbing the scattered γ-rays in the region by the wall. This apparatus shows the high counting rate in the low density and indicates the low counting rate in the high density. Accordingly, such a logging is effective in surveying the groundwater survey of the caves or cracks of lava, limestone, etc.

Fig. 9 shows the results of $\gamma - \gamma$ logging of the groundwater in Mishima lava conducted by this method. It is clear that the scattered rays with the energy as low as 30 ~ 50 Kev was increased remarkably by the groundwater in the caves of lave.[15)]

In addition, the existence of groundwater veines can be grasped even by n-n logging in this case. However, there exists a defect that the increase ratio of counting decreases due to the in the hole.

The n-n logging is used for the measurement of porosity as well as $\gamma - \gamma$ logging. Epithermal neutron is used for its measurement because the logging is influenced by Cl in the coastal saline water zone. The thermal neutron detector is surrounded by the Cd plate 3 mm in thickness and the n-n logging is measured by the epithermal neutron.

The radioactivity logging is absolutely essential for the hydrogeological survey at present.

6. Improvement of deep wells

Water-collection from superposed layers is made by setting several steps of strainer for deep wells in Japan. The measuring apparatus of pumped water of the deep well in each aquifer was proposed to investigate various problems of water resources being occurred recently and to conduct the proper design of the deep well. The principle of the measurement is as follows: Ten scintillation detectors and RI in the lower and as the point source are inserted in the pipe of the deep well discharging the aquifer. Then, the flow velocity is obtained in the upper layer of each aquifer (Fig. 10). The discharge measured by triangular weir and the pumped water calculated using this meter accorded with an error of 5%.

7. Survey of dam leakage by RI and leakage treatment

Irrigation water is being kept using reservoirs from the olden day in Japan. At present their number is assumed to be about 50,000. In general, such dams are situated in valleys and water is reservoired by damming up the river. In Japan, valleys generally develop along faults and most of them have disturbed and permeable bed rocks.

Japan has so many dams on fractured zones along faults where the natural radioactivity is 1.3 to 2.0 times as high as background and the fractured zones form the leakage route.

Such bed rocks with increased radioactivity were detected in several places such as Birusawa Dam in Yamagata Pref., Yasugawa Dam in Shiga Pref., etc.17-19) (Fig. 11)

The nuclear density and moisture gages were used for the case of impossibility of water storage due to the remarkable leakage from the dam or for the survey during repairs of the dam body.20) The percolation mechanism of leakage would be more adequately investigated if this method is used simultaneously together with the RI tracer method. If the leakage route becomes clear by RI, the leakage treatment works would be possible based on the traced route.

Leakage control of earth dams is mainly made by forming water prevention walls by grouting in the bed rock. The injection capacity for grout milk can be assumed by the following method: The porosity of layers is obtained by nuclear density gage. Then, the coefficient of permeability is obtained by water pouring test. Lastly, the relationship between the permeability test and the effective range of grout injection introduced by H. Jade is used.

The needed amount of grount cement was obtained by this method at Tateshina Reservoir, Nagano Pref. and was compared with the practical amounts of injection of cement. As the results, both amounts were accorded with the error of 2%.

8. Survey of acid water with RI

In Japan, the areas where agricultural water is damaged by naturally gushing acidic hot springs, waste water from mines and factories, etc. reach about 37,000 ha. The percolation method is realized by introducing it to the hillside, letting it flow into the percolation well excavated to the depth of 10 m or several tens of meters, and to neutralize it by contact with rocks or soils. In this method, the acidity is neutralized also by the dilution phenomenon of groundwater to some extent.

In the planning of the treatment works for acid water, the existence of connection of groundwater veins and the flow velocity of the groundwater controlling neutralization should be previously determined.

This method was used for the survey of acid water at Zao Spa, Nishiazuma Mine in Yamagata Pref., Higashiazuma Mine in Fukushima Pref., etc.

Nuclides mainly used were ^{32}P and ^{131}I and the distance of detection reached 600 m to 1,800 m.22)

9. *Measurement of water amount in the river for irrigation use with RI*

The discharge in the upstream part of rivers where the irrigation water is obtained is often measured at the valley where the stream is turbulent. Therefore, it would be impossible to apply the conventional by hydraulic formulas proposed by specialists as they are. The RI method has been noticed for the measurement in such a case and it is being used for small rivers for irrigation use in the field of agricultural water utilization. For this purpose, ^{24}Na, ^{131}I and ^{118}Au are mainly used. Three scintillation probes are set in the central part of rivers and near both banks and they are self-recorded.[23)]

The discharge is obtained by the flow velocity measured by means of the pulse injection of RI and the dilution ratio obtained by means of continuous injection. The RI method was compared with discharge measurement by using a triangular weir in the model canal. Both results were approximately accorded with an error within 10%. It was clear that the RI method is more accurate than chlorine ion titration method in the discharge calculation.[24)]

Fig. 1. Structure of scintillation counter for car-borne

2. Distribution of natural radioactivity at the east slope of Mt. Fuji and pumped water from deep wells
3. Surface shape of base of Narita group and boundary of groundwater before and after 1954
4. Scintillation counter of groundwater tracing （写真省略）
5. Tracer detection curves in the observation wells
6. Diagram of the radioisotope velocity meter
7. Scattered γ-rays intensity in various beds of alluvium and diluvium
8. Determination of porosity by means of $\gamma - \gamma$ logging
9. $\gamma - \gamma$ spectro-logging in Mishima underground river
10. Measuring method of the yield from aquifer by the use of measuring meter for the vertical flow of deep wells with radioisotopes
11. Intensity distribution of natural γ-rays at Yasugawa Dam

FIG. 1 DIAGRAM OF SCINTILLATION COUNTER FOR CAR - BORNE

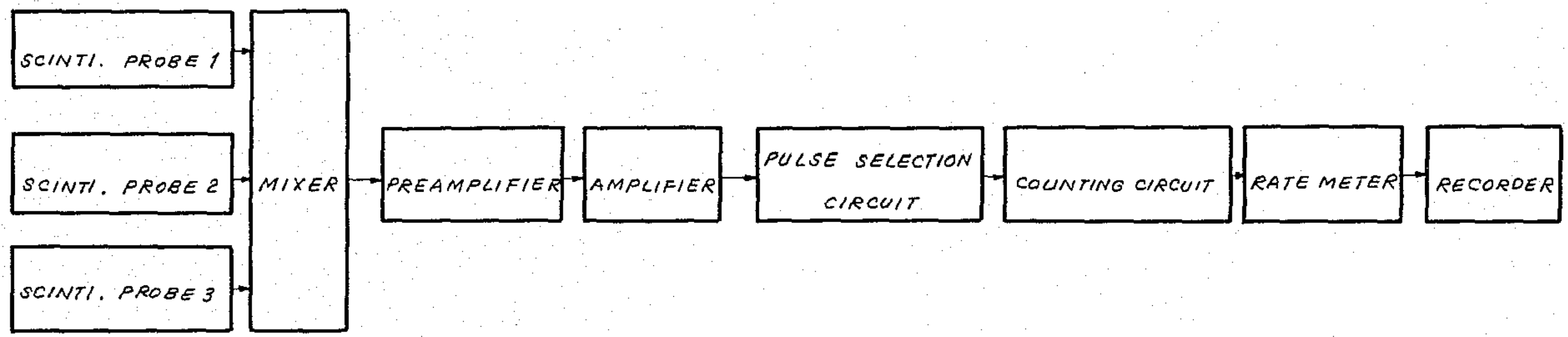


FIG. 2 DISTRIBUTION OF NATURAL RADIOACTIVITY AT THE EAST SLOPE OF Mt. FUJI AND PUMPED WATER FROM DEEP WELLS

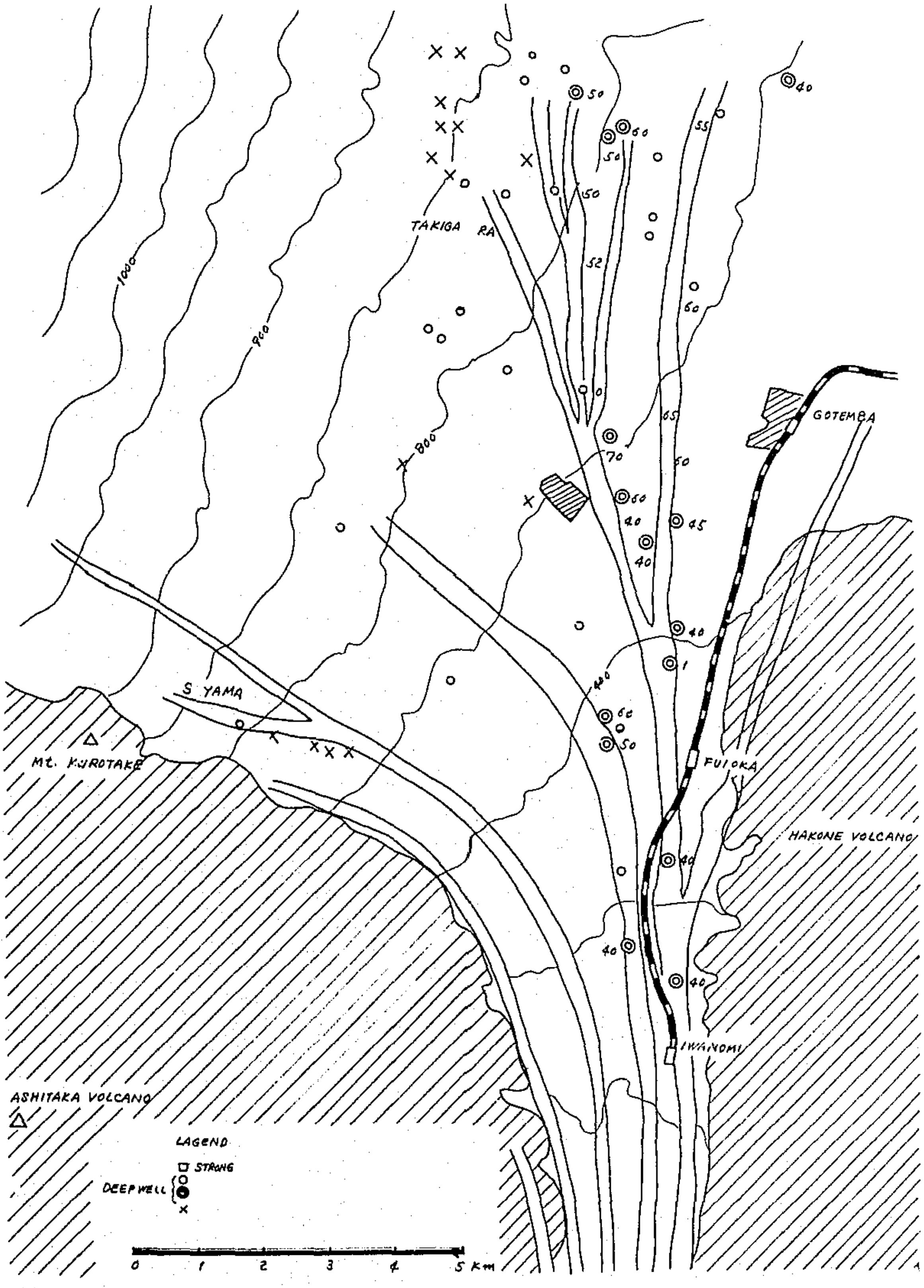


FIG. 3 SURFACE SHAPE OF BASE OF NARITA GROUP AND BOUNDARY OF GROUND WATER BCFORE AND AFTER 1954

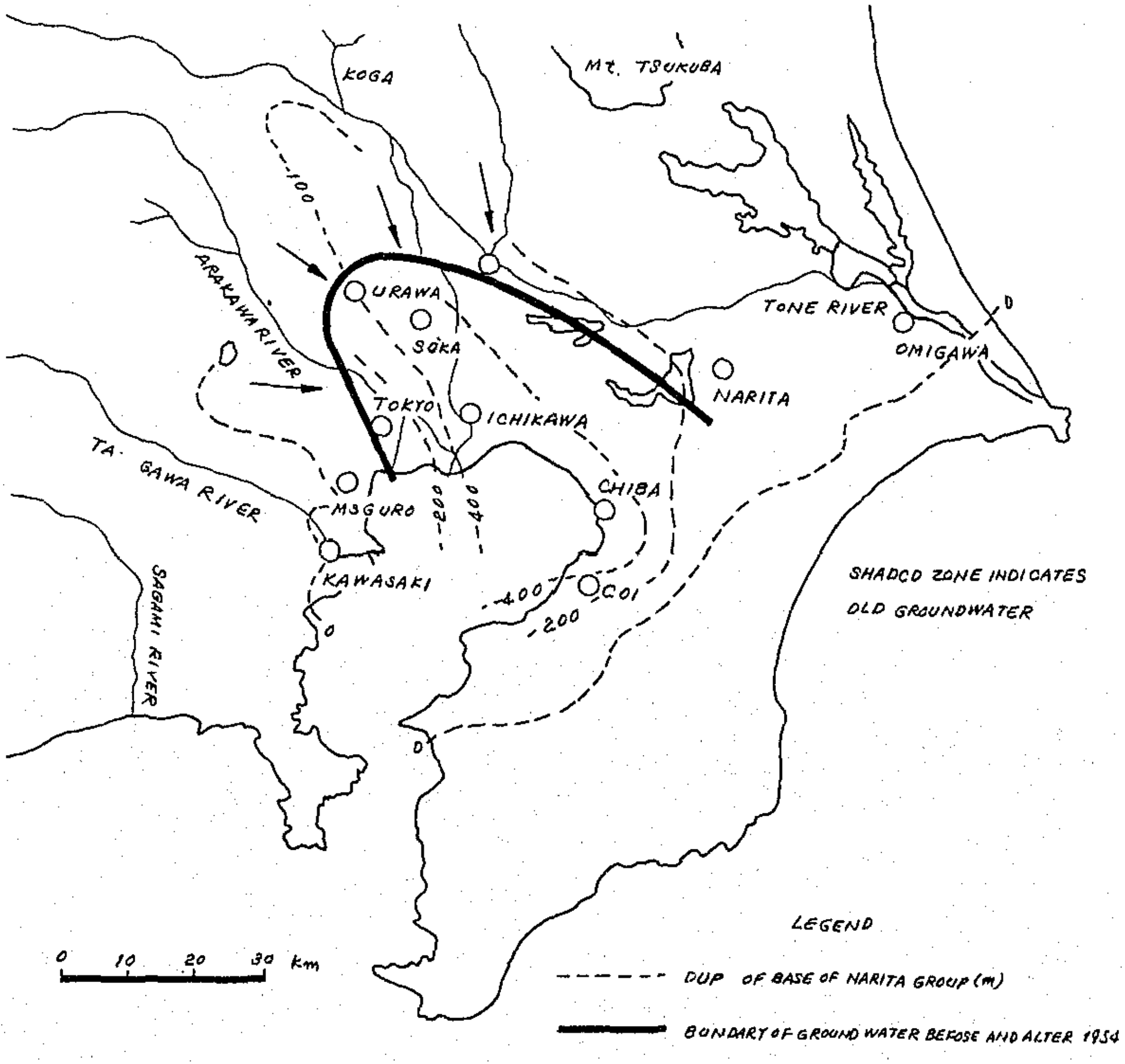


Fig. 4　Scintillation counter of groundwater tracing

（写真省略）

FIG. 5 TRACER DETECTION CURVES IN OBSERVATION WELLS

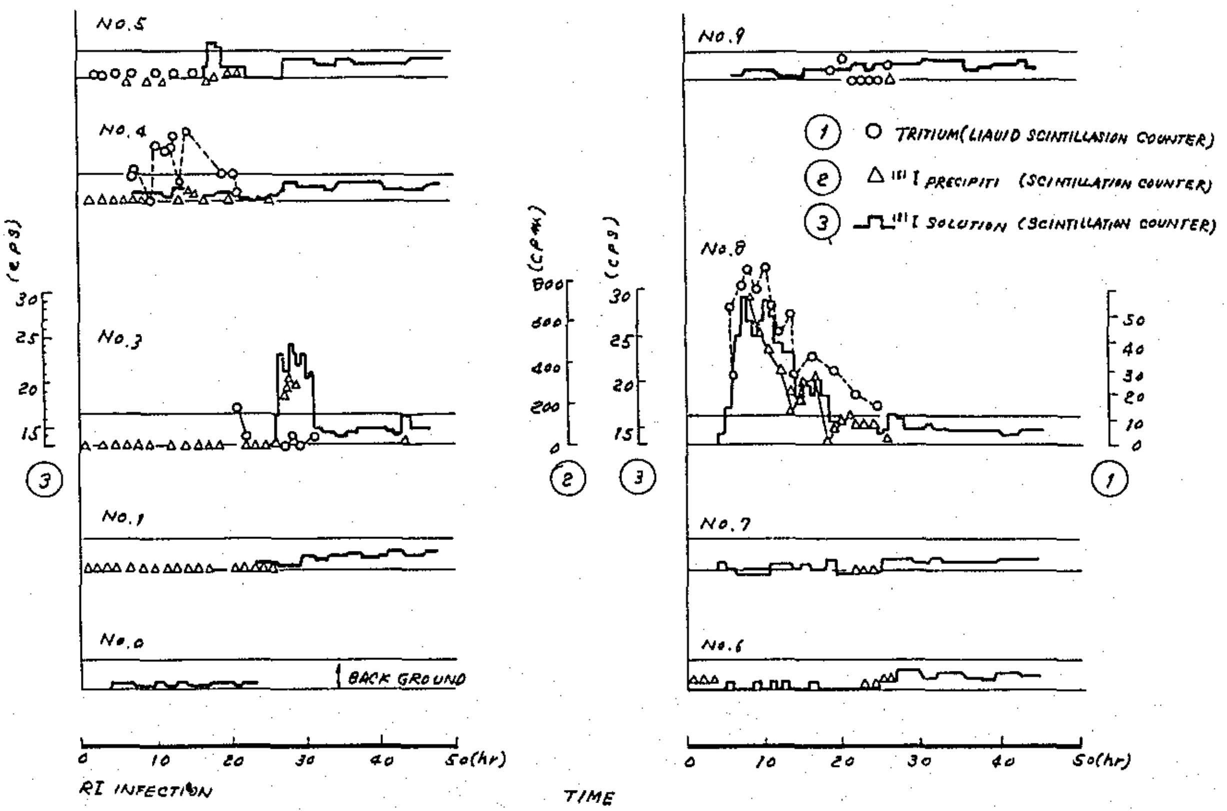
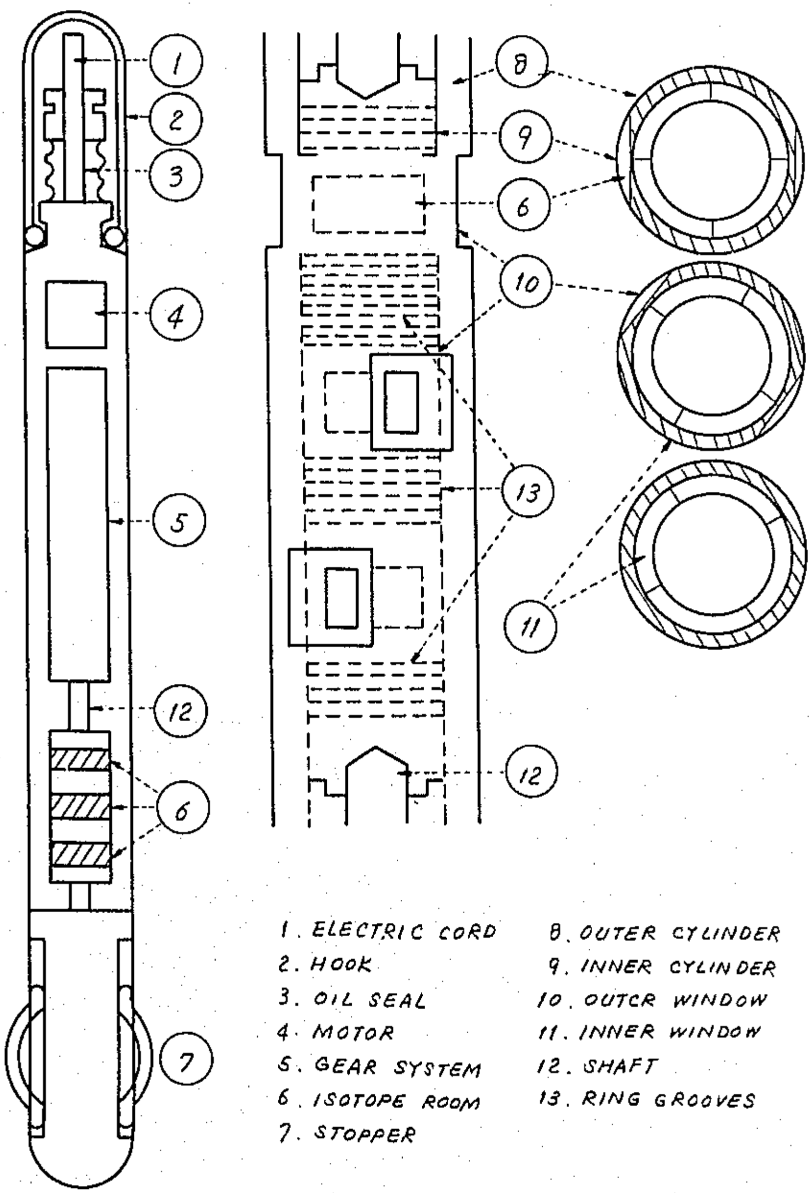


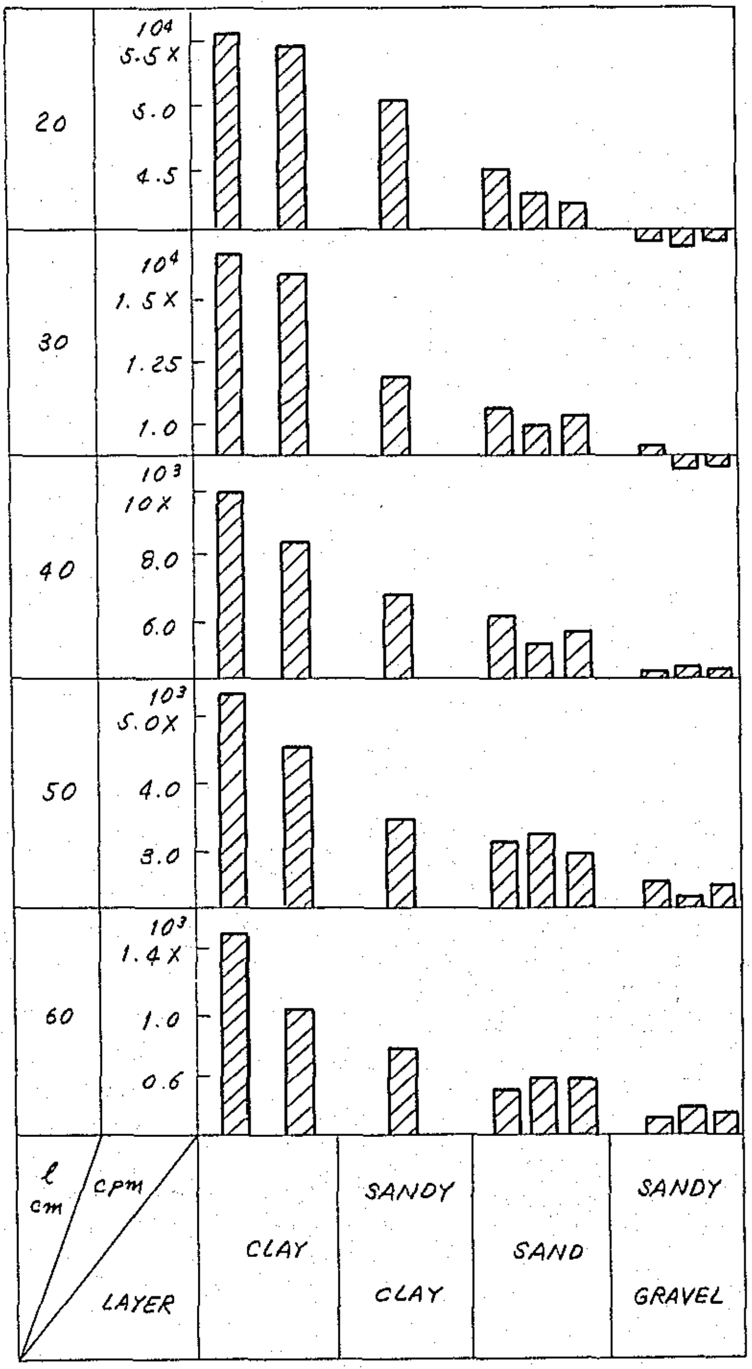
FIG. 6

DIAGRAM OF THE RADIOISOTOPE VELOCITY METER



1. ELECTRIC CORD
2. HOOK
3. OIL SEAL
4. MOTOR
5. GEAR SYSTEM
6. ISOTOPE ROOM
7. STOPPER
8. OUTER CYLINDER
9. INNER CYLINDER
10. OUTCR WINDOW
11. INNER WINDOW
12. SHAFT
13. RING GROOVES

FIG. 7 SCATTERED γ-RAYS INTENSITY IN VARIOUS BEDS OF ALLUVIUM AND DILUVIUM



WITH IRON CASING, 50mm IN DIAMETER.

WITH GROUND WATER IN THE HOLE; γ-RAY SOURCE: ^{60}Co 1 mci,

DETECTOR: SCINTILLATION COUNTER.

ℓ : DISTANCE FROM SOURCE TO DETECTOR.

FIG. 8 DETERMINATION OF POROSITY BY MEANS 7-7 LOGGING

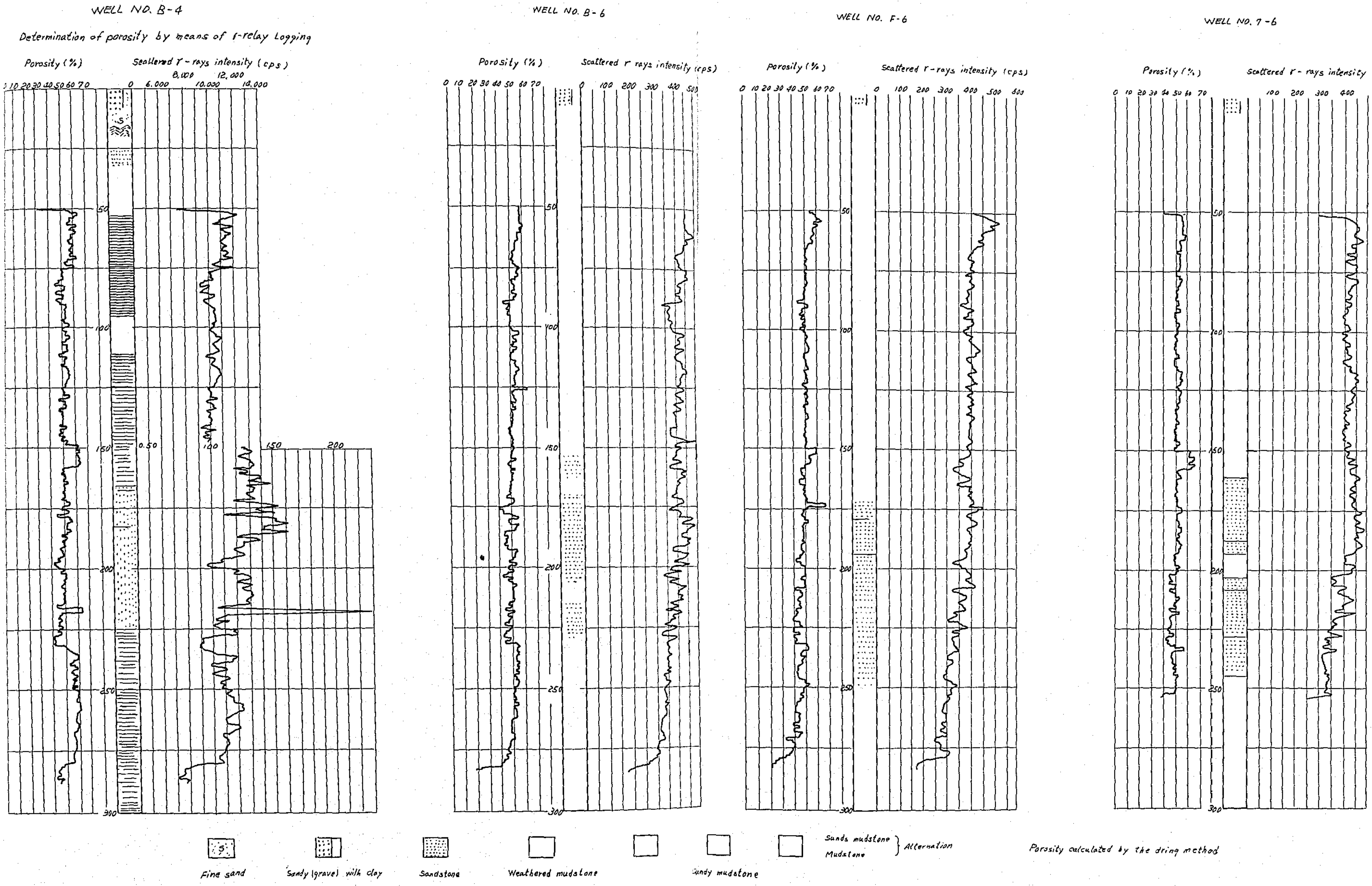


FIG. 9 γ – γ SPECTRO-LOGGING IN MISHIMA UNDERGROUND RIVER

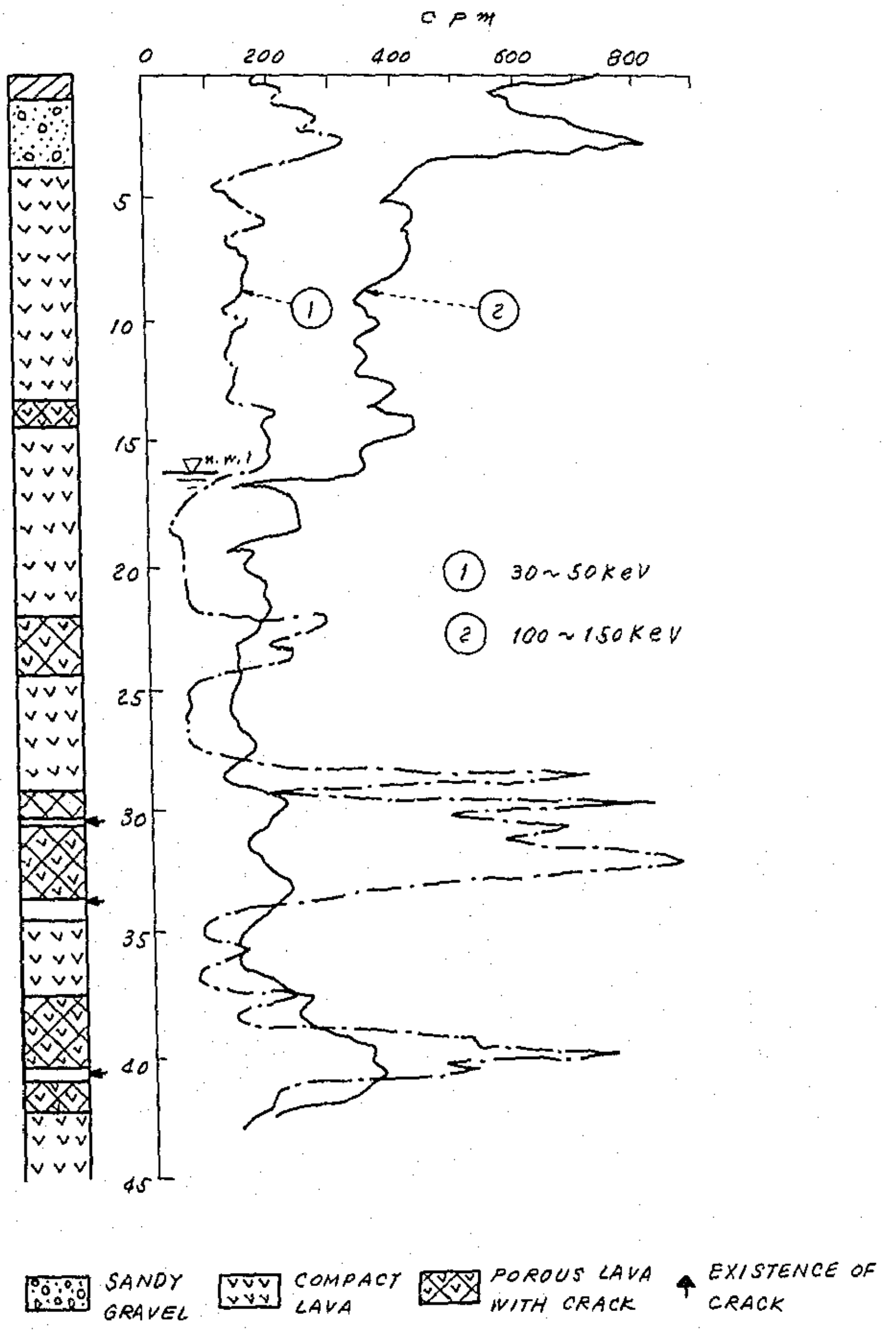
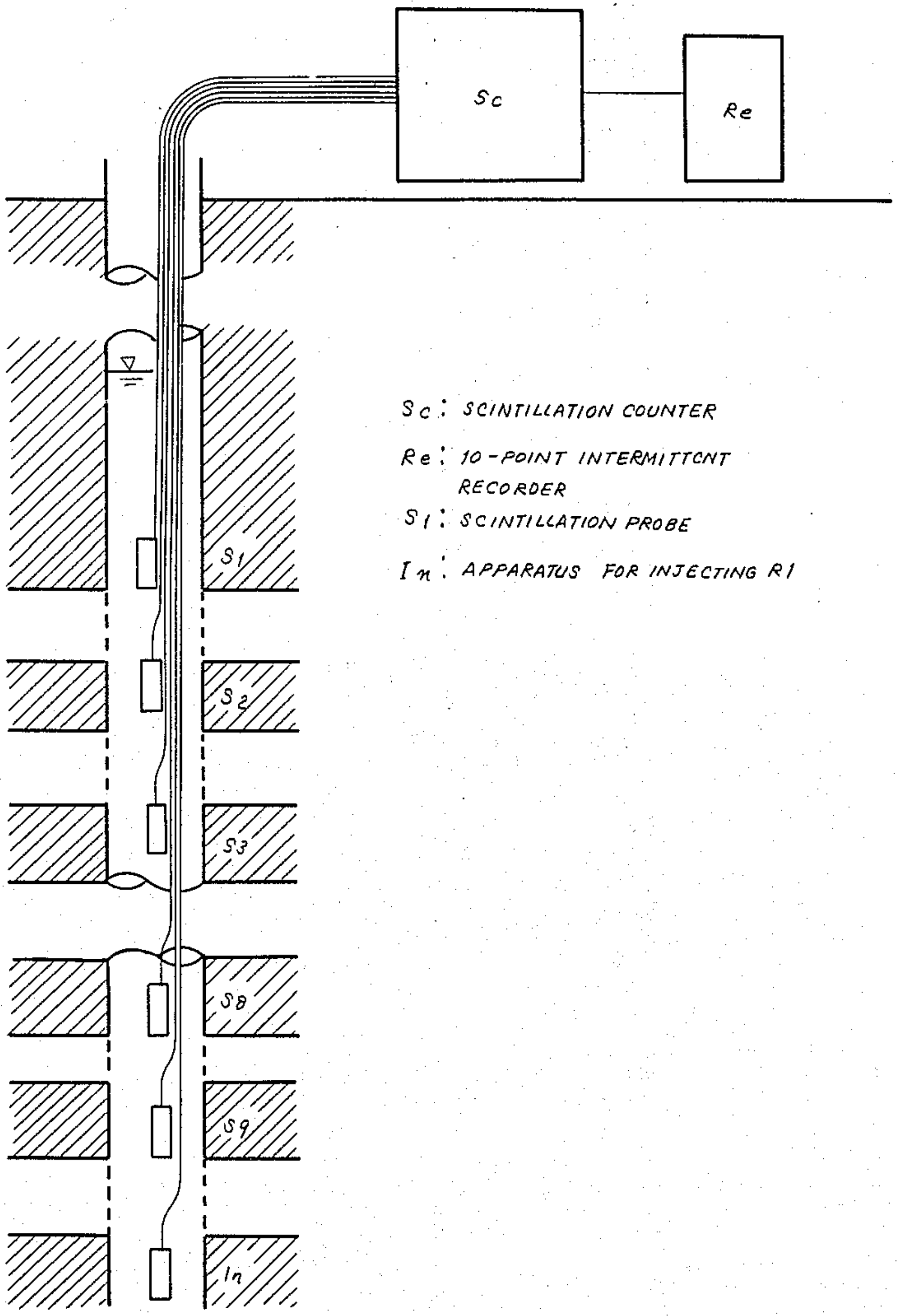


FIG. 10

MEASURING METHOD OF THE YIELD FROM AQUIFER BY THE USE OF RADIOISOTOPES



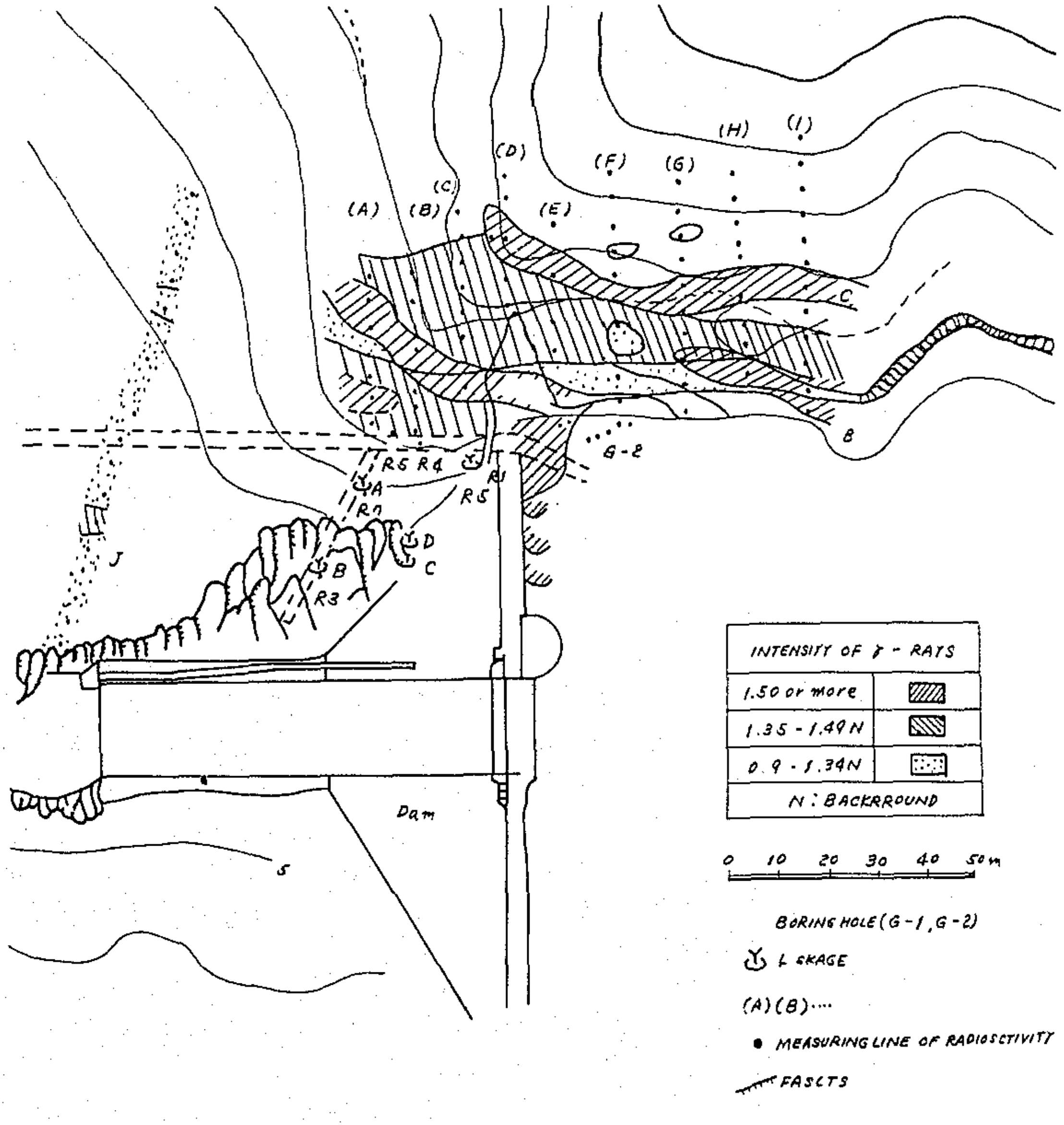
S_c : SCINTILLATION COUNTER

R_e : 10-POINT INTERMITTENT RECORDER

S_i : SCINTILLATION PROBE

I_n : APPARATUS FOR INJECTING RI

FIG. 11 INTENSITY DISTRIBUTION NATURAL γ - RAYS AT YASUGAWA DAM



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GEOCHEMISTRY

OUTLINE ON THE GEOCHEMISTRY OF GROUNDWATERS

– especially on the geochemistry of fossil waters –

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Japan is located on the island arc and occupies the eastern part off the Asian Continent. Japan occupies the latitude from about 24°N to 46°N, and she has the moderate climate and much amount of rainfall (about 1,000 ~ 2,000 mm a year). The topographic conditions of Japan are very steep, and we have only 30% of agricultural plain throughout the country. The stream waters run through the mountainous areas rapidly, therefore, the shallower underground waters circulate in a relatively short duration compared with the groundwaters in the continents.

On the earth, there are many kinds of water such as meteoric water, surface water, groundwater, hotspring water, sea water, etc., but at this time, I would like to explain the fossil water. The groundwater of metal mines and magmatic water are excluded.

What Is the Fossil Waters

Generally speaking, the old water which has been sealed in the sedimentary formations is called fossil water or connate water. At first, I would like to explain briefly the usage of technical terms concerning the concealed waters.

Levorsen, A.I. (1954) classified the oilfield water into three groups, namely meteoric water, connate water and mixed water. Omura, I. (1934) used in his book the technical terms of fossil water, connate water and residual water as the same meanings. Kartsev, A.A., et al. (1954) classified the oilfield water into bound water (capillary, pendular) and free water, but they did not use the technical terms concerning the genesis of groundwater in the sedimentary formations. Rankama, K. and Sahama, T.G. (1950) divided the groundwater into meteoric water and connate water, and Takahashi, J. (1922) used in his book the terms of fossil water and connate water (The term of connate water was first used by Lane, A.C. in 1908). De Sitter, L.U. (1947) classified the water in the lithosphere into meteoric water, connate water and juvenile water. White, D.E. (1957, 1960) used the terms of magmatic water, connate water and metamorphic water, furthermore, he used the terms of connate water and volcanic water for the waters of deep origin.

Anyway, judging from above mentioned facts, I can get the following conclusion.

"Fossil water" is almost the same meaning as connate water, and when the sedimentary formations reserve the groundwater with almost the same geological age as of the formations then we can call the groundwater fossil water.

Classification of Waters by Chlorine Ion

In the sedimentary formations deposited under marine condition we can expect the existence of fossil water of marine origin, on the contrary, in the sedimentary formations of fresh water environment we can expect the existence of fossil water of non-marine origin.

The water in the sedimentary formations (especially in the surface sediments of the sea) shows very complicated chemical compositions caused by the various kinds of chemical reactions between the interstitial water and sediments. Chlorinity shows no significant variation during the deposition, and sea water contains about 20g Cl^- per kg. Therefore, some researchers divide the waters by the content of Cl^-, and here I show Gesser's classification.

1.	Fresh water	$Cl^- < 0.1$ g/l
2.	Low concentration saline water	$Cl^- = 0.1 - 1$ g/l
3.	Medium concentration saline water	$Cl^- = 1 - 10$ g/l
4.	High concentration saline water	$Cl^- = 10 - 17$ g/l
5.	Sea water	$Cl^- > 17$ g/l

Water with 1 – 17 g/l Cl^- (3 and 4 of the above classification) is called "Brackish water".

Some oilfield waters from foreign countries contain eight times as much as in Cl^- content compared with that of the sea water. Therefore, above listed classification can be adopted only on the mixed waters of fresh water and sea water.

It is presumable that almost all groundwaters in the sedimentary areas have been derived originally from the mixture of fresh water (mineral concentration is very low; meteoric origin; nearly equal to distilled water) and sea water. Chemical reactions in the sedimentary rocks will be explained later, and at this time, I have to list up the chemical composition of sea water. (unit is mg/l of sea water)

Na – 10,730, K – 390, Mg – 1,310, Ca – 420, Cl – 19,460,
Br – 67, SO_4 – 2,690, HCO_3 – 132, HBO_2 – 19, I – 0.05.

Weight ratios of chemical components are as follows.

Ca/Cl – 2.16×10^{-2}, Ca/Mg – 0.32, Mg/Cl – 6.73×10^{-1},
Na/Cl – 0.55, K/Cl – 2.0×10^{-2}, Br/Cl – 3.9×10^{-3},
I/Cl – 0.003×10^{-3}

Some examples of oilfield waters show extremely high mineral concentration, and as the reference to consider this fact, the solubility of three salts is shown below.

NaCl – at 0°C 357 g/l of H_2O or 229 g/l Cl^-
at 100°C 391 g/l of H_2O
KCl – at 20°C 347 g/l
at 100°C 567 g/l
$CaCl_2$ – at 0°C 595 g/l
at 100°C 1,590 g/l

Some oilfield waters contain about 200 g/l Cl^-, and this value is nearly equal to that of NaCl saturation in water. In Japan, there are only three or four examples of oilfield waters which are containing minerals more than sea water, but in case of foreign oilfields it is not difficult to notice the values of mineral concentration from one to three times larger than that of sea water.

Kartsev, A.A., et al. (1954) listed up the following values as the characters of ground-water from the oilfields.

mg/1,000 g water % equivalent	total ions – 200 to 300,000 100	Cl^- – 100 to 180,000 1 to 50
Br – 0 to 2,000 0 to 49	I – 0 to 120 0 to 0.01	HCO_3 – 0 to 18,000 0 to 49
$C_nH_mO_2$ * – 0 to 5,000 0 to 49	HS – 0 to 200 0 to 0.2	SO_4 – 0 to 4,500 0 to 15
Na – 100 to 90,000 30 to 50	K – 5 to 1,400 0.05 to 1	NH_4 – 0 to 450 0 to 0.3
Ca – 4 to 20,000 0.1 to 30	Mg – 0 to 6,000 0 to 10	

* organic anions (chiefly naphthenate ion)

Examples of Chemical Composition of Groundwaters from Oilfields, Coalfields and Gasfields

Sampling of Groundwaters

The majority of coalfield waters are collected in the underground coal mines.

In the underground coal mines, some chemical exchanges on the groundwater are induced by the artificial ventilation, and natural gases dissolved in water also escape from groundwaters. If there is a boring well with natural water flowing, it is the best sampling site for geochemical survey of groundwaters.

Fig. 1 shows an example of the chemical variation of formation water in the coal mine in relation to the sampling localities. This study has been done at the Syoro Coal Mine, Hokkaido, by the members of the Geological Survey of Japan. In this figure, ◎ shows the bore hole from which the original formation waters are flowing out. During the flowing down of the water along the fissures in the Harutori Coal Bearing Formation, the chemical composition of the original formation water changes rapidly. ◎ shows the site of prospecting well for coal seams, and this well reaches to the ceiling of gallery. Therefore, the chemical composition of the original formation water is pH-8.2, Cl = 2,520 mg/l, HBO_2 = 32.7 mg/l, I = 3.9 mg/l. However, at the site 1.5 m apart from ◎, groundwater shows extremely high pH value (pH = 12.08). This is not the natural value, because at this site, several years ago cement gratuing had been done for the protection against the breakdown of the gallery. The Cl^- contents are changing from 2,520 mg/l to 2,800 mg/l, according to the relative positions of sampling site and the boring well, and this variation may be caused by the artificial ventilation. I^- contents are decreasing rapidly from the boring site. HBO_2, also decreases according to the distance from the well, and it shows the values from 32.7 to 11.8 mg/l.

Above stated facts are clearly showing the chemical exchange in the underground coal mine.

The directions of chemical changes of groundwater in the coal mine gallery are supposed as follows.

pH – Approaches to the value of RpH. Water containing much amount of carbon dioxide and having acidic character changes to the basic side. In case of oxidation of FeS_2 (pyrite), groundwater shows low pH values (pH = 3 to 1).

Cl – In case of small amount of water production, Cl^- concentration increases by the evaporation of mine water.

HCO_3 – Chemical variation is relatively small.

Free CO_2 – In case of large amount of free CO_2, the CO_2 content decreases. Sometimes, CO_2 increases by the oxidation of organic matter in the gallery.

Fe – Decreases with the reaction of oxidation, $Fe^{2+} -> Fe^{3+}$. But, sometimes Fe increases by the reaction of oxidation of pyrite.

SO_4 – Increases by the oxidation of pyrite.

I^-, HBO_2, Ca^{2+} – Decrease.

Ca^{2+}/Mg^{2+}, P – Decrease.

NH_4^+ – At the first stage, decreases. If the groundwaters contact with air and sedimentary rocks with organic matter the NH_4^+ content increases by the slight oxidation.

Examples of Chemical Analyses for Groundwaters

At this time, I do not explain the analytical methods for fossil water.

Table 1 is showing the analytical data of groundwaters from gasfields, oilfields and coalfields. These analyses have been done at the Geological Survey of Japan except for three data.

In this table, groundwaters of Nos.5 and 6 are obtained from fresh water deposits of Pleistocene and Pliocene ages. Both of these waters contain natural gas of CH_4 type. The Cl^- content is very low (below 10 mg/l).

Groundwaters, from natural gas fields, with large concentration of Cl^- have the following chemical characteristics compared with sea water: I^- increases largely; Mg decreases; Ca increases slightly: The values of Ca^{2+}/Mg^{2+} ratio are below one except for No.4 (Takable Gas Field, Shizuoka Prefecture). The groundwater of No.4 has especially high concentration of Ca^{2+}, and its chemical characteristics are resembling to those of the coal mine waters.

All groundwaters from oilfields are collected from the marine sedimentary formations (No.7 to 12). These waters have high concentration of Cl^-, I^- and HCO_3^-, and the chemical characters resemble to the groundwater from natural gas fields. The values of Ca^{2+}/Mg^{2+} ratio are almost the same or slightly larger than those of groundwaters from natural gas fields. I suppose, the groundwater of No.13 (Venezuela) is not the fossil water, because the Cl^- content is very small (only 89 mg/l), the HCO_3 content is large (5,263 mg/l), CO_2 is rich (120 mg/l) and the Ca^{2+} content is smaller than the Mg^{2+} content.

Groundwaters of Nos.14 and 15 are containing 40 g/l Cl^- and 184 mg/l respectively, and both have 250 mg/l SO_4^{2-}, therefore, these waters may be recovered from oil pools in the evaporite formations. On the contrary, in Japan, crude oil has been recovered from saproperic formations.

The geochemistry of coalfield waters is very interesting. Water from the Shinhoronai Coal Mine, Hokkaido, riches in HCO_3^- and I^- contents and its chemical composition is almost the same as that of oilfield and gasfield waters. This water has relatively small value of Ca^{2+}/Mg^{2+} ratio as a coalfield water. According to the data on Table 1, the coal mine waters in Japan have low concentration in SO_4^{2-} content. The existence of SO_4^{2-} is well explained by the invasion of sea water into the sedimentary formations and by the oxidation of sulfides in the strata. Generally, coal bearing formations and their surrounding formations are shallow marine, brackish water and fresh water sediments, and these formations consist mainly of organic sediments, therefore, as a hole, these formations have the reduction capacity. Thus, in general, high concentration of SO_4^{2-} does not occur in the original coalfield waters. As a result of the reduction of SO_4^{2-}, H_2S

increases, and sometimes S is precipitated as iron-sulfides. I^-, Br^- and NH_4^+ contents in the coalfield waters are fewer than those of the oilfield and gasfield waters, and coalfield waters show the $Ca^{2+} \gg Mg^{2+}$ relation and also they generally show the value of Ca^{2+}/Mg^{2+} ratio over five. The coalfield waters recovered from the lower formations of coal seams at the Taiheiyo and Syoro Coal Mines have the high Ca^{2+} contents like groundwater from Takabe Natural Gas Field (No.4 in Table 1), and they have the values of Ca^{2+}/Mg^{2+} ratio from 60 to 110.

These waters in the lower formations are clearly classified from those in the upper formations, therefore, it is possible to decide the geologic horizons of water reservoirs using the chemical composition of waters. At the Syoro Coal Mine, saline groundwater with 6 g/l Cl^- is flowing out from fresh water deposits, consequently it is possible to presume that the groundwaters had migrated secondary into these formations. In this case, we also call these groundwaters "Fossil Water".

Japan is located on the volcanic belt, therefore, there are many CO_2 gas seepages throughout Japan. In case of coal mines, Ube Coal Mine, Yamaguchi Prefecture, is famous for its large production of carbon-dioxide. No.31 of Table 1, shows the analytical data of coal mine water from Ube. Inspite of the high concentration of HCO_3^- from 14 to 17 g/l in coal mine water at the Ube Mine, water shows pH value lower than 7 because of the richness in the CO_2 content. The Cl^- content is as small as 100 mg/l, SO_4^{2-} content is also as small as 2 mg/l, NH_4^+ is 0.6 mg/l: Therefore, these waters may have their present chemical characteristics by the chemical reactions between groundwater, carbon dioxide agas and sedimentary rocks. In this case, the strong chemical attack against the sedimentary rocks by carbon dioxide gas might occur.

The coal mine waters from the Oyubari Coal Mine, can divided into two groups, one is the high concentration group of Cl^- (1.47 g/l) and the other is the low concentration group of Cl^- (306 and 42 mg/l). Both groups have high contents of HCO_3^- and Na^+ (260 ~ 2,150 mg/l). These waters may be derived by the sufficient chemical reaction between the sedimentary rocks and groundwater with relatively low concentration of Cl^-. Therefore, these waters show the chemical characteristics of water types of $Na\text{-}HCO_3$ or $Na\text{-}Cl\text{-}HCO_3$.

There are some differences between the chemical compositions of oilfield waters and coalfield waters, as stated above.

Kartsev, A.A., et al. (1954) stated in their book as follows: The useful chemical components to distinguish the oilfield waters and the coalfield waters are I^- and NH_4^+. I^- is more abundant in the oilfield waters than in the coalfield waters, and the element exists as sodium iodide, furthermore I^- might be derived from seaweeds. The solubility of iodine compounds in oilfields is larger than that of coalfields, and also the content of iodine in coal is small. NH_4^+ is abundant in oilfield waters, while the nitrogen compounds in the oilfields are more unstable than those in the coalfields.

Chemical Change or Metamorphism of Sea Water and Fresh Water in Underground

Judging from the relationship between the chemical compositions and the geological ages of groundwaters (see Table 1), we can decide the general direction of metamorphism of groundwater caused by the interaction of water and sedimentary rocks. It is important to research the direction of metamorphism for the identification of the fossil water or for the recognition of the history of groundwaters.

According to the following researchers, at the first or initial stage of the deposition under the aquatic mass, the interstitial solutions of sediments increase the biogenic elements, such as alkalinity (nearly equal to HCO_3^-), NH_4^+-N and P, further increase K^+, SiO_2, and decrease Mg^{2+} (as a result of takeup by chlorite and possibly incipinent dolomitization). The names of the researchers are Emery, K.O. (1952), Chilinger, G.V. (1958), Motojima, K., et al. (1958), Siever, R., et al. (1968), Brooks, R.R., et al. (1968).

Chave, K.E. (1960) stated the metamorphic direction of sea water in the sedimentary rocks as follows.

	Cl	K/Cl	Ca/Cl	Mg/Cl	Sr/Cl	SO_4/Cl	HCO_3 *	Br/Cl	I/Cl
increase			0		0			0	0
irregular	0								
decrease		0		0		0	0		

* In case of no influence of meteoric invasion.

The above listed data were obtained from the sedimentary regions from Pliocene to Ordovician in age. HCO_3^- is produced by the decomposition of organic matters, and the rate of this decomposition may be increased by the supply of oxigen, therefore, in the foreing countries, sometimes HCO_3^- and CO_2 have been used as the indicator components to identify the meteoric invasion into the underground. According to this theory, the majority of the oilfield and gasfield waters in Japan have been influenced by the meteoric invasion, and so-called "fossil waters" obtained from Japan are not correspond perfectly to those of foreing countries. Concerning to this point, I will explain later.

According to White, D.E. (1960) almost all deep oilfield brines from the Tertiary formations are rich in Na and Cl, and he stated also the type of the deep oilfield waters from the pre-Tertiary strata is Na-Ca-Cl type throughout the world. The classification of these two types has been done for the sake of convenience by Ca/Na, and it is clear that such chemical reactions as increase the values of Ca/Cl and Ca/Na exist in the underground.

White also presumed the existence of "Metamorphic Water". After the connate water with high Cl^- content has been driven out from the sedimentary rocks, the bound water in the water-bearing minerals (clay minerals etc.) will be pushed out into the porous layers and this water is called "Metamorphic Water". Generally, this water is $NaHCO_3$-Boron type. The characteristics of this water are rich in HCO_3, B and Na components, and the water has relatively high value of I/Cl ratio, on the contrary it has extremely low values of Li/Na and K/Na ratios compared with the values of volcanic waters.

Alekseev, F.A., et al. (1958) stated that highly metamorphosed groundwaters show high value of (Cl - Na)/Mg ratio and they called the values "Coefficient of metamorphism". On the other hand, groundwater of Na-CO_3 type has high value of Na/Cl ratio. They called this ratio "Degree of freshing". When the meteoric water invaded the underground and mixed with the groundwater of NaCl type, the decomposition of organic matter occur, and consequently the contents of CO_2, HCO_3 and CO_3 increase, after the increase of these CO_2 groups, the groups attack the sedimentary rocks and accordingly Na^+ is driven out from the rocks until its content comes to be in a state of equilibrium to the contents of HCO_3^- and CO_3^{2-}. The groundwater becomes gradually Na-CO_3 type and it shows high value of Na/Cl ratio. This is the reason why HCO_3^- has been used as an indicator component of the invasion of meteoric water.

According to De Sitter, L.U. (1947), the metamorphism of sea water in the sedimentary rocks can be divided into two groups owning to the geological ages of the sedimentary rocks.

(a) From Recent to Tertiary: Mg^{2+} decreases grestly as a result of takeup by carbonates and sulfate minerals. $K^+ + Na^+$ increase slightly. Ca^{2+} decreases slightly as a result of takeup by carbonate and sulfate minerals.

(b) From Tertiary to Devonian: K + Na and Ca increase.

(c) In case of the addition of meteoric water: Concentration of minerals in water decreases. SO_4^{2-} and HCO_3^- increase. The value of Na/Ca + Mg ratio decreases.

Therefore, when Ca, Mg and Na contents are plotted on the triangular net, in case of (a) the points are distributed in the area of Mg decrease and Na increase, and in case of (b) the points are distributed in the area of Na decrease and Ca increase.

The chemical changes of the sealed fresh water in the sedimentary rocks of fresh water deposits, may proceed along the direction of the increase of Na, Ca and HCO_3, and ultimately, this chemical reaction proceeds toward the water type of Na-HCO_3, but in this case, boron content may be very few compared with the content of "Metamorphic Water".

Quality of Groundwater and the Status of Existence of Groundwater

To study the fossil waters, it is very important to clarify the relationship between the quality of groundwater and the present status of existence of groundwater. To obtain this relationship, we can use the factors such as (1) quality and quantity of dissolved gases in groundwater (Nalijunaya, 1961), (2) helium, argon, neon, nitrogen and other inert gases (Aleksev, F.A., et al., 1958), (3) oxidation and reduction potentials of groundwaters (Germanov, A.I., et al., 1959). Anyway, this problem can be solved, using the distribution of many kinds of elements and the chemical characteristics of them.

In regard to the qualities of groundwaters, at first, I pay attention to the geochemical research for oilfield waters at the Apsheron Peninsula, Azerbaijan S.S.R., U.S.S.R., studied by Chilinger in 1959. He classified the oilfield waters into four types, and each type may corresponds to the following environments.

(1) $Na\text{-}SO_4$ type (Sodium bicarbonate type): $(\gamma Na - \gamma Cl)/\gamma SO_4 < 1$: where γ is percent equivalent. The groundwater of this type had not been shut off from the atmosphere, and this water does not show the preferable reserve condition of crude oil in the underground, and also this type of groundwater can find in the shallower part of the underground.

(2) $NaHCO_3$ type (Sodium bicarbonate type): $(\gamma Na - \gamma Cl)/\gamma SO_4 > 1$: The groundwater of this type contains much amount of bicarbonate salts and NaCl, but SO_4 is very few, H_2S exists in the water, naphthenic acids also exist in the water. The majority of the oilfield waters, gasfield waters and coalfield waters are belonging to this type. This water has had some communications between the atmosphere.

(3) Mg-Cl type (Magnesium chloride type): $(\gamma Cl - \gamma Na)/\gamma Mg < 1$: The water of this type is a special one in nature and it has not yet been found out in Japan.

(4) $Ca - Cl_2$ type (Calcium chloride type): $(\gamma Cl - \gamma Na)/\gamma Mg > 1$: The groundwater of this type has been regarded as the end member of metamorphism of groundwater in the closed underground. This is the typical type of the fossil water. I have already stated this point, and we can find the groundwater of this type from various localities of coalfields and oilfields in Japan. This water has been obtained as a result of strong mineralization in the underground. The existence of the water of this type without I^- and naphtenic acid can be used as an indicator to show the condition of good shutoff from the atmosphere.

In Japan, when the groundwaters show the qualities as shown in (2) and/or (4) and have high concentration of Cl^-, generally we call them "Fossil Water".

I suppose that in case of the "Fossil Water" of fresh water deposits, the chemical composition of this water may be nearly equal to that of type (2).

Figure 2 is showing the examples of the relationship among the chemical types of groundwaters, gas water ratio and geologic conditions.

How to Recognize the Invasion of Meteoric Water into the Underground

When the meteoric water invaded the underground, it is possible to recognize this phenomenon using the following facts.

(1) Dissolved oxygen in the water – The meteoric water contacting with air has the dissolved oxygen of 6.36 ml/l at 20°C, on the contrary, as a rule, fossil water does not contain the dissolved oxygen. Using these facts the invasion of the meteoric water can be recognized.

(2) Dissolved argon and dissolved nitrogen – The atmospheric air contains 0.93% of argon, and corresponding to this value, the meteoric water at 20°C contains 0.30 ml/l of dissolved argon. This meteoric water flows into the underground and changes *the quantity of dissolved argon, corresponding to the surrounding* conditions. Further, it is possible to use the combined relationship between argon and helium, and sometimes the dissolved nitrogen has been used for the determination of the influence of the meteoric water against the underground.

(3) SO_4 – Meteoric water contains SO_4^{2-}, but in general, the fossil waters from the saproperic oilfields do not contain SO_4^{2-}.

(4) Cl – The fossil water with large amount of Cl^- exists in the underground and the meteoric water invades this water mass, then the fact of the invasion can be noticed by the dilution of Cl^- content of groundwater. According to this method, in Japan, the invasion of the meteoric water has been noticed to the depth of about 200m below the groundsurface at the oilfield and coalfield regions.

(5) CO_2, HCO_3 and H_2S – When the water containing the dissolved oxigen contacts with the organic matters in the underground, the organic matters may be oxidized by oxygen then CO_2, HCO_3 and H_2S may be produced by this reaction. This phenomenon has been observed at the oilfields in the U.S.A. and the U.S.S.R.

(6) NH_4 – The decomposition of organic matters in the sedimentary rocks may be progressed when the meteoric water with dissolved oxygen contacts with the organic matters. NH_4 is the products of this decomposition.

(7) Oxidation reduction potential (Redox potential) – Generally, fossil waters have the negative (minus) values of the redox potential. On the contrary, meteoric water shows positive value. Therefore, meteoric invasion changes the negative value groundwater into the positive one. Germanov, A.I., et al. (1959) noticed the meteoric invasion into the underground to the depth of 2,000 or 3,000m from the groundsurface. Generally, old groundwaters have the negative values of the redox potential, and these values reach up to –0.5 volt.

(8) Water temperature – When the strong invasion of the meteoric water into the underground strata occurs, the temperatures of these strata change compared with other strata. In Hokkaido, some water bearing strata show lower temperatures than their surrounding strata.

(9) pH – Motojima, K., et al. (1962) noticed at the Syoro Coal Mine, Hokkaido, that the weathering of the marine sedimentary rocks results in acidic condition at the near surface parts of the formations.

(10) Na^+ etc. – Generally, the invasion of the meteoric water into the underground changes the value of Na/Cl ratio of the groundwater into the direction of high value.

(11) ^{14}C, ^{3}H and ^{10}Be – These are the methods to use the radioactive elements. At present, the relation between geologic age and absolute age is as follows.

Unit: Million years

Alluvium	0.01
Pleistocene	1
Pliocene	11 ~ 12
Miocene	25
Oligocene	33 ~ 40
Eocene	55 ~ 60
Paleocene	70 ± 2
Cretaceous	135 ± 5
Jurassic	180 ± 5
Trias	270 ± 5
Paleozoic	540 ~ 600 ± 20
Precambrian	4,500 ±

^{14}C has a half life (t) of about 5,500 years, therefore it is possible by ^{14}C method to decide the absolute age (T) till 20,000 ~ 40,000 years. The value of t of ^{3}H is about 12.5 years, therefore T is about 25 years. The t of ^{10}Be is about 300,000 years, and the T is about 600,000 years. All these elements are the products of the cosmic radiation, and they may be caught by the meteoric water in the atmosphere, after the catch the quantity of these elements decreases gradually according to their half lives. Using these elements as the indicators. It is possible to notice the latest influence of the atmosphere against the underground.

There are some possibilities to decide the atmospheric influence to the underground using the stable isotopes, such as ^{2}H (or D), ^{36}Ar and ^{18}O. Concerning a part of this problem, White, D.E. proposed the method of recognition of "where the groundwater came from?"

At present, it is impossible to decide both the absolute ages of the groundwater and of its reservoir rocks, except for the case of very young ages (till about 20,000 ~ 40,000 years).

Conclusion

Almost all sedimentary rocks contain the groundwater. A part of the groundwater is the free water. Sometimes, we can obtain the groundwater through the boring holes even from the very old formations.

The quantity of the movable groundwater is different place by place. However, we have the surveyed result that at the Syoro Coal Mine, Hokkaido, the groundwater with high value of Ca/Mg ratio and 6 g/l of Cl^- was recovered from the underground at the rate of two million tons a year. Therefore, it is true that vast amount of groundwater exists in the regions of the sedimentary rocks. The problems are the quality and quantity of recoverable water. In the lithified sedimentary areas, at the first step of water production, the character of compressibility of water may act as a role of increasing the production

Fig. 1 Figure showing the relation between the chemical composition of groundwater and sampling site at the Syoro Coal Mine, Hokkaido

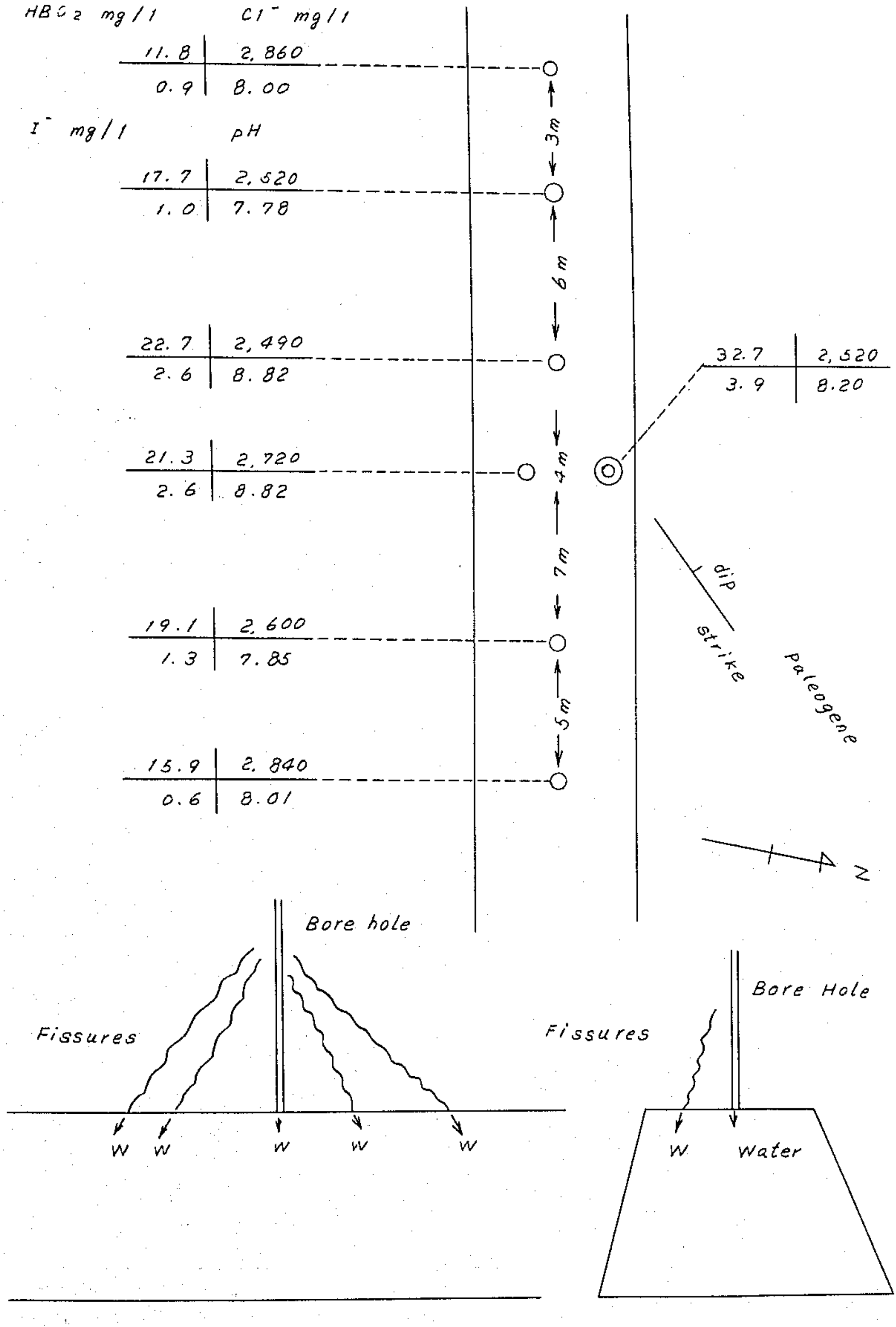

Fig. 2 Type of groundwater versus gas water ratio

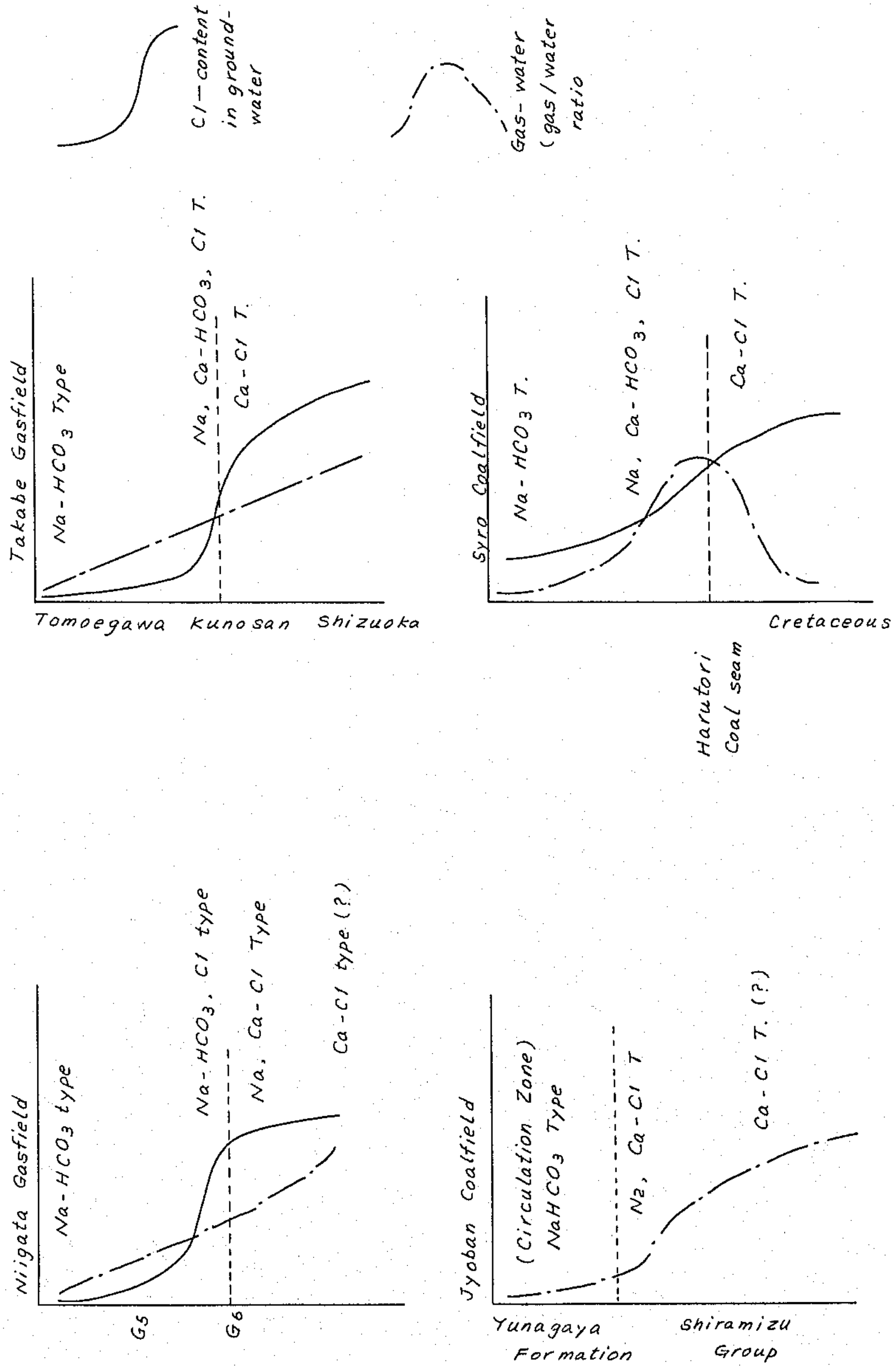

Table Examples of chemical composition of groundwaters

Unit: mg/l and ppm

	Locality	Geology	pH	RpH	HCO_3^-	CO_3^{2-}	f·CO_2	Cl^-	I^-	Br^-	NH_4^+	Ca^{2+}	Mg^{2+}	K^+	Na^+	SO_4^{2-}	(Ca^{2+}/Mg^{2+})	HBO_2
Gasfield water	1. Niigata-city	Qt. (P)	7.3		900	0	86.4	7250	26.1		77.6	345	372	91.2	4000		0.93	
	2. 〃	Ter. (850m)	7.7		1040	0	26.4	15820	33.8	58.4	36.6	209	539	424	8990		0.39	
	3. Yokoshiba, Chiba	Ter.	7.5		1400	0	57.2	19000	90.5	68.2	34.2	194	550	340	10600		0.35	
	4. Takabe, Shizuoka	Ter.	7.7	7.7	47.4	0	7.5	6020			0.34	2167	41				52.8	
	5. Nakano, Yamagata	*Qt.	6.7		413	0	136	6.5			18	30						
	6. Kakuto, Miyazaki	*Ter.	7.4		341	0	35.2	6.7			28.8	17.4	11.8	3.8	49.9	2	1.6	
Oilfield water	7. Isobe 2B, Gumma Pref.	Ter.	6.9	7.9	8570	0	616	12940	5.7	41.7	30.3	203	53.4				3.8	673
	8. Kurokawa R6, Akita	Ter.			1650	0	202	15650	17.1	43.8	36.6	176	256				0.69	
	9. Yabase, Akita p.	III	7.1	7.7	5830	0	141	12090	28	37	27	110	187	427	8901	2	0.59	
	10. Yabase, Akita	V 〃	6.7	7.0	4580	0	273	14840	38	50	59	168	87	644	10198	0	1.93	
	11. Yabase, Akita	VII 〃	7.9	8.1	1790	0	70	4480				22	6.7				3.3	
	12. Yabase, Akita	VIII 〃	7.2	7.6	891	0	18	16100	35	47	103	91	85	204	10136	0	1.1	
	13. Lagunillas, Venezuela Mio.				5263	120		89				10	63	2003			0.16	
	14. East Texas, U.S.A. Up. Cret.					387		40598				1432	335	24653		259	4.3	
	15. Oklahoma, U.S.A.	Ordov.				18		184387				18753	3468	91603		268	5.4	

	Locality	Geology	pH	RpH	HCO_3^-	CO_3^{2-}	f·CO_2	Cl^-	I^-	Br^-	NH_4^+	Ca^{2+}	Mg^{2+}	K^+	Na^+	SO_4^{2-}	(Ca^{2+}/Mg^{2+})	HBO_2
Coalfield water	16. Yubari, Hokkaido	Ter.	7.1	7.7	854	0	34.3	4820	7.1		1.8	1377	32.3			2 >	43	173
	17. Shinhoronai, Hokkaido	Ter.	7.8	8.0	487			11680	44.7		3.7	383	104			2 >	2.7	
	18. Heiwa, Hokkaido	Ter.	7.4	7.7	112	0	20	9470	11.3		3.6	635	77.2			2 >	8.2	
	19. Oyubari, Hokkaido		8.3	8.5	3010	117	0	1470			1.2	9.3	10.9	13.7	2145	2 >		
	20. Oyubari, Hokkaido		7.8	8.4	1510	0	2	306			1.0	7.3	1.7	5.9	754	28		
	21. Oyubari, Hokkaido		7.4	8.4	646	0	11	42				14.6	3.0	2.4	257	25		
	22. Syoro 29B, Hokkaido	Ter.	8.0	8.0	149	0	3	2620	5.8	5.4	0.9	76.4	11.4	8	1320		6.7	
	23. Syoro 4B, Hokkaido	Ter.	8.2	8.4	38	0	9	2520	3.9	20.9	0.7	325	8.7	11	1330		36.3	
	24. Syoro 4B, Hokkaido	Ter.	8.2	8.2	57.1	0	3	2490	2.6		0.9	242	11.7	4.1	1242	3	20.7	
	25. Syoro 4B, Hokkaido		7.4	7.3	28.6	0	9	6270	5.2		1.2	1426	22.9	10.0	2378	3	62.3	
	26. Taiheiyo, Hokkaido	30m	8.28		31.7		0	3030	10.9	9.9	0	913	8.1	5.4	1410	0	113	
	27. Nakaso, Fukushima	Ter.		8.1	203	0	4	840	1.0	6.0	0.1	57.3	3.9	10.5	538	tr	15	
	28. Iwazaki, Fush	Ter.	8.1	8.2	24	0	3	4100	4.0	20.2	1.1	362	9.6	40.5	2270	7	3.6	
	29. Ube, Yamaguchi	Ter.	7.0	7.6	1710	0	90	137			0.0A							
	30. Ube, Yamaguchi	Ter.	7.2	7.3	79	0	7	5740	3.4	12.0	4.5	860	43.9	23.3	2640	5		
	31. Motoyama,	Ter.	6.7	7.3	1400	0	95	100			0.6	47.4	122	17.8	262	2 >		