SVERIGES GEOLOGISKA UNDERSÖKNING
Rapporter och meddelanden nr 27

Intruded and relict groundwater of marine origin

SWIM 81

Proceedings of
Seventh Salt Water Intrusion Meeting
Uppsala, Sweden, 14—17 September 1981

Uppsala 1981

SVERIGES GEOLOGISKA UNDERSÖKNING

Rapporter och meddelanden nr 27

INTERNATIONAL HYDROLOGICAL PROGRAMME

SEVENTH SALT WATER INTRUSION MEETING

Uppsala, Sweden

September 14-17, 1981

PROCEEDINGS

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Earlier Salt Water Intrusion Meetings:

1968, Hannover, Federal Republic of Germany

1970 May, Vogelenzang, The Netherlands

1972 June, Copenhagen, Denmark

1974 August, Ghent, Belgium

1977 April, Medmenham, England

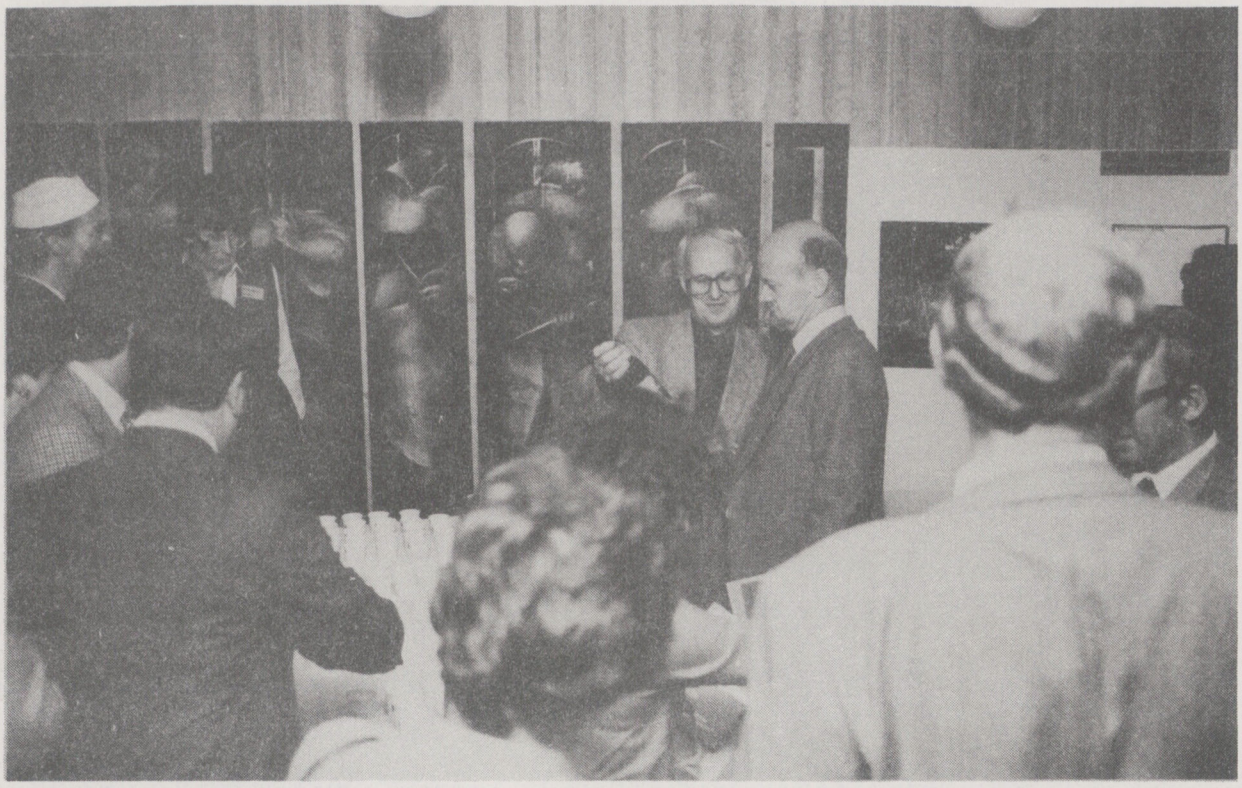
1979 October, Hannover, Federal Republic of Germany

Participants gathering in the main lobby of SGU

Undivided attention in the auditorium

Socializing after a day of lectures and discussions

More round-table get-togetherness



Examining the bouquet and taste of more or less saline groundwaters

Picnic on old viking grounds during excursion

PREFACE

Starting in 1967, at intervals of 2-3 years, experts from countries sharing salt water intrusion problems, have come together to exchange information. The meetings have been informal in character and have been organized within the framework of the IHD/IHP. The number of participating countries has increased through the years.

The Seventh Meeting in the series was convened by the Swedish Committee for Hydrology of the Natural Science Research Council (NFR), which is also National IHP Committee,and the Geological Survey of Sweden (SGU). It was supported also by the VBB AB. The organizing Committee consisted of Mrs Malin Falkenmark (NFR), Mrs Helgi Lindewald (SGU) and Mr Bo Leander (VBB). The meeting was held 14-17 September 1981 in Uppsala, Sweden, and was devoted to problems regarding intruded and relict groundwater of marine origin.

At the meeting which was attended by 50 participants from 10 countries, 22 papers were presented at four sessions, devoted to descriptive contributions, field methods, modelling, and application, respectively. On September 17 a field excursion was arranged to study the water supplies of the towns of Uppsala and Vaxholm, and to visit Verka Representative Basin, where relict salt water is successively brought into circulation in response to land uplift.

This volume contains the papers presented at the meeting together with a short summary of the final discussion. The editorial work was carried out by Miss Iréne Johansson, Committee for Hydrology.

November 1981

Bo Leander
Chairman of the
Organizing Committee

Malin Falkenmark
Executive Secretary
Committee for Hydrology

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PROGRAMME

MONDAY, SEPTEMBER 14

10.00-13.00	Registration at *SGU*, Villavägen 18, Uppsala	
13.00	Opening Ceremony Chairman of Committee for Hydrology: *Gunnar Lindh* Director General of SGU: *Gunnar Ribrant* Chairman of Municipal Council: *Brita Nordlander*	
13.30	Swedish problems with intruded and fossil groundwater of marine origin	*Nordberg*
14.00	*Coffee break*	

S E S S I O N 1 - Descriptive contributions

Chairman: *Mr. Bo Leander*

14.30	Saline groundwater in Sweden	*Lindewald*
	Some wells with high content of chloride in central Sweden	*Engqvist*
	Fossil groundwater of marine origin in the Uppsala area	*Sidenvall*
15.45	*Pause*	
16.00	Sea water intrusion in drilled wells	*Bergman, Sund*
	Effects on groundwater composition of sequential flooding of aquifers with saline and fresh water	*Agerstrand, Hansson, Jacks*
	Spatial salinity variations in the Netherlands	*Boswinkel*
	Saline contamination phenomena in the karstic and fissured carbonatic aquifer of the Salentine Penisula (Southern of Italy) and their evolution	*Cothecchia, Tadolini, Tulipano*
19.30	Reception, City of Uppsala	

TUESDAY, SEPTEMBER 15

S E S S I O N 1 - Descriptive contributions (cont.)

09.00	Salt water intrusion into coastal aquifers of Gdansk region	*Kozerski*

S E S S I O N 2 - Field methods

Chairman: *Prof. W. de Breuck*

09.25	Testing VLF - resistivity measurements in order to locate salty groundwater	*Müllern, Eriksson*
	Hydrochemical observations in the subsoil of the western part of the Netherlands	*Pomper*
10.15	*Pause*	
10.30	Salt water in connection with groundwater sampling in deep boreholes	*Gidlund*
	Sea water encroachment in the Llobregat and Besós areas, near Barcelona	*Custodio*
11.20	*Coffee break*	

S E S S I O N 3 - Modelling

Chairman: *Dr. Z. Haman*

12.10	Fresh and salt groundwater in the Dutch coastal area in relation to geomorphological evolution	*de Vries*
	Application of vortex distributions in modelling the storage of fresh water in saline aquifers	*Peters*
13.00	*Lunch*	
14.15	*Sightseeing tour* Start from the main entrance of the Cathedral	
18.00	Demonstration of *Hydrological Division, University of Uppsala* and *Geological Survey of Sweden*	
	Social evening at *SGU*	

WEDNESDAY, SEPTEMBER 16

S E S S I O N 3 - Modelling (cont.)

09.00	The simultaneous flow of fresh and salt groundwater in horizontally extended aquifers	*van Duyn*
	Assessment of the flow regime of interfingering fresh and salinar groundwaters in the sedimentary cover of a salt dome	*Delisle, Schelkes*

	The subterranean flow of fresh and salt water underneath the Western Belgian beach	*Lebbe*
10.15	*Coffee break*	
10.45	Analysis of the possible shapes of the fresh water-salt water interface in a semi-confined aquifer with radialsymmetric boundary conditions	*van Dam*
	Some considerations on the form of the fresh water-salt water interface in the case of more aquifers separated by semi-previous layers	*ten Hoorn*
	The numerical calculation of salt water/ fresh water interfaces using the stream function	*van den Akker*
12.00	*Lunch*	

S E S S I O N 4 - Applications. Terminology

Chairman: *Prof. J.C. van Dam*

13.00	Artifical removal of intruded saline water in a deep aquifer	*Schuurmans* *van den Akker*
13.25	Terminology	
14.00	*Coffee break*	

S E S S I O N 5 - Conclusions

Chairman: *Prof. J.C. van Dam*

14.30	General discussion. Synthesis and conclusions
16.00	Closing of the meeting

THURSDAY, SEPTEMBER 17 - ONE-DAY EXCURSION

Excursion by bus through the beautiful scenery southeast of Uppsala, well known since the age of the Vikings.

Excursion leader: *Mr. T. Wikner*

Visits:

1. Water supply of *Uppsala*
   Guide: *Mr. J. Sidenvall*
2. Fossil salt water problems in *Verka Representative Basin*
   Guide: *Dr. L.Y. Nilsson*
3. Water supply of *Vaxholm* (town in the *Stockholm* archipelago)
   Guide: *Mr. P.O. Engvall*

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INTRODUCTORY SESSION - SHORT SUMMARY

by M. Falkenmark[1]

Professor Lindh, Chairman of the National IHP-committee, in opening the seminar, recalled the 14 year development of the SWIM-meetings, which have continuously attracted the interest of an expanding number of countries and of an increasing number of participants, as the problem of saline groundwater has increased in importance in Europe.

Mr. Ribrant, Director General of the Geological Survey of Sweden (SGU), co-organizer of the Meeting, expressed satisfaction of localizing the meeting to SGU in Uppsala, and gave a short overall presentation of SGU and its activities.

Mrs. Nordlander, Mayor of Uppsala, recalled that Uppsala was situated in an area earlier drowned by the water masses of the Baltic Sea and its predecessors. Even during Viking times, the town was connected with the Baltic and the Viking ships could sail well beyond Uppsala. Today due to the continued land upheaval, the sailing channel has been reduced to a rivulet in a valley.

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1 NFR's Committee for Hydrology, Stockholm, Sweden

PROBLEMS IN SWEDEN WITH INTRUDED AND FOSSIL GROUNDWATER OF MARINE ORIGIN

by Lars Nordberg[1]

INTRODUCTION

It is a significant occasion today when the Geological Survey of Sweden is hosting an important, international meeting concerning saline groundwaters. The significance for the Survey lies in the fact that this arrangement marks the present responsibility in the field of hydrogeology for this institute. It is a responsibility that has been enlarged considerably during the last few years. Let me give you some figures for illustration. In 1967, about 2.6 % of the total financial turnover of the Survey applied to hydrogeology. It amounted to a mere 0.4 million SEK. In 1978 the share had grown to 14 %, corresponding to 17 million SEK, and this year, 1981, preliminary figures point at 17 %, which is 31 million SEK. A major part of the growth is at present tied to comprehensive undertakings for investigating possible sites for future storage of radioactive wastes which, by the way, has put the work of the Survey in the middle of environmental and political debate. Also that is a rather new experience to us.

Today we gather to discuss less controversial matters, I hope, but matters of great importance. I have not had the opportunity of attending any previous Salt Water Intrusion Meetings, but it has been reported to me that these meetings convey quite a lot of substantial information for exchange between the participants and also quite a lot of pleasant and informal get-togetherness. I hope that our meeting here in Uppsala will not be an exception from that fine tradition.

Now, let me turn to the subject for this communication: Problems in Sweden with intruded and fossil groundwater of marine origin. It has to be said from the very start that our problems in Sweden are of minor significance - at least in comparison with corresponding problems in many of your countries.

I will give you a general overview of reasons for saline groundwater in Sweden, and I will dwell shortly on some specific cases as examples.

REASONS FOR SALINE GROUNDWATERS

Three main reasons for groundwaters with high salinity can be stated - and with high salinity I mean water with a salt content exceeding 1 g/l.

1) Fossil waters or influence from fossil waters of marine origin.

Sweden has very few large aquifers in sedimentary bedrock. In some of them, saline groundwater is occasionally encountered. The genesis of the salinity is not always quite clear. But in a few cases with water from very deep boreholes it is concluded that the salts derive from the marine environment in which the sediments were laid down.That is true, e.g., for southern Scania where Triassic rocks at a depth of 1200 - 1900 m yield groundwater with, in one case 98 and in another case 154 g/l chloride. An enrichment of salt in the formations has

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1 Geological Survey of Sweden, Uppsala, Sweden

taken place since the time of sediment accumulation. Several similar occurrences are reported from the same area. These very deep-seated saline groundwaters have a high temperature and may be utilized for extraction of heat. The high salinity, though, introduces some special problems as installations are vulnerable to salt, i.e. clogging and corrosion problems may arise. As these deep waters never have been considered for water supply, no other problems are at present connected with their utilization. As a conclusion about fossil groundwaters in old sedimentary bedrock in Sweden, it is reasonable to say that they are reported from Cambro-Silurian and Mesozoic bedrock, that they often cannot be distinguished from later intruded water, and that they, as a matter of fact, pose rather small problems.

A more recent type of fossil groundwater - or water influenced by fossil groundwater - is found in areas that were inundated by sea-water after the latest glaciation in Scandinavia. During the deglaciation, at one stage, the saline Yoldia sea covered the coastal parts of Baltic Sweden and with a wide strait across southern central Sweden connected the Atlantic with the Baltic, involving the lake Väner basin. The duration of this stage, though, was rather short - a few hundred years occurring around 8000 BC - and the salinity is believed to have been much lower than in present ocean water. The next stage of saline water in the Baltic, starting with the Litorina transgressions around 5500 BC is still in existence. Due to the land upheaval large areas previously below the Yoldia and Litorina Seas are now land consisting of sediments that were deposited in saline water. Due to recharge of aquifers by atmospheric water, the groundwater circulation has removed the saline water and replaced it with fresh water. But in localized areas, mainly where circulation is slow because of low hydraulic conductivity or because of isolated positions in relation to groundwater flow, saline waters remain. This means that they are never found in well-drained esker systems or similar formations although in eskers in the Stockholm region they occur in some deep, rather isolated basins. They can frequently be encountered in fine-grained sediments, or below such sediments, and preferably in low-lying areas around large lakes. This type of fossil salt water more often occurs in the crystalline bedrock than in the overburden where it has been fully replaced by fresh water. It can also be stated that the salinity usually increases with depth, giving an indication of decreasing dilution and replacement with increasing depth.

As a matter of fact, a map of Sweden depicting reported occurrences of boreholes yielding groundwater in which the chloride content exceeds 300 mg/l coincides very well with a map of areas that were inundated by postglacial seas. Such saline groundwater can be found in localized areas in the crystalline basement and in Paleozoic sedimentary rocks, and also in the unconsolidated Quaternary deposits. Areas where this type of saline groundwater is known to occur more than sporadically include:

a) the sedimentary bedrock, mainly chalk and sandstone, in western Scania, e.g. around Ängelholm and between Malmö and Landskrona; and in loose deposits in the same area.

b) the sandstone of northern Öland

c) the sandstone, limestone and marls on Gotland

d) the sandstone in Östergötland, Västergötland and Närke

e) the clay plains of Västergötland south of lake Vänern

f) the sandy Laholm flats on the west coast

A special investigation of the composition of the groundwaters in one of the mentioned areas, the clay plains of Västergötland, published in 1974, reveals that areas with a chloride content exceeding 300 mg/l coincide with areas having the greatest thickness of loose deposits (20-30 m), and particularly when located in closed basins. Of 870 wells in various geological environments, not a single one in glaciofluvial deposits yielded saline groundwater. Such water was restricted to the crystalline bedrock and frictional material beneath clay.

Besides in the mentioned six areas, saline groundwaters are reported from highly scattered locations from below the Holocene marine limit. It should be kept in mind, though, that only approx. 800 wells in the whole of Sweden have been identified as yielding saline groundwater - out of more than 0.5 million existing wells. For this reason the problems have limited implication. But needless to say, for the domestic well-owner it can be of serious concern, yes it may even be the limiting factor for his livelihood.

However small, it is quite clear that fossil groundwaters from the period of the latest deglaciation causes much more disturbance for the daily well-being of people than similar saline groundwaters from Paleozoic and Mesozoic aquifers.

2) Intruded saline waters.

Near the coastal areas of Sweden many wells yield saline groundwater. No thorough investigations have been performed so far to determine whether these waters are truly fossil or if they derive from recently intruded sea-water. For the private well-owner it makes no difference. But for the management of municipal and industrial water supply systems it may be of decisive importance. Fossil waters and intruded waters have to be attacked in different manners in order to mitigate the side-effects of continued extraction.

Now, let us look at areas where salt water intrusion is known to exit. Sweden has a long coast line, including tens of thousands of islands in the archipelagos. The areas are populated by permanent inhabitants and, temporarily, by a great number of vacationists having summer homes along the coasts. Resorts and camping places are also abundant in the same areas. This concentration of people introduces a seasonally heavy demand for water. In most cases groundwater is utilized - both from dug and drilled wells - and this is where the problem comes in. Many wells draw water from small freshwater recharge areas, particularly when located on small islands. In the summer the period of heaviest demand coincides with the period of little or no replenishment from precipitation water. Therefore the zone of influence may increase to outside the coastline and saline water may intrude into the utilized aquifer. In many places, therefore, people have to restrict their water withdrawal considerably in order to avoid saline water for consumption. In areas with dense settlements many conflicts - some with judicial implications - arise from lack of knowledge or lack of cooperation concerning water extraction. The insufficient knowledge of the nature and origin of saline groundwater in coastal areas also leaves the planning authorities with serious hesitation for granting new building permits. A better understanding of the behavior of groundwater in these coastal areas is therefore necessary. Areas where intruded sea-water cause problems for well-owners are particularly the Bohuslän archipelago on the west coast, the Stockholm archipelago on the east coast and the island of Gotland. These three areas all attract great numbers of summer residents, putting a heavy stress on the environment. The occurrence of, or the risk for, withdrawing saline groundwater is in some places a restraint for further expansion of habitations - both permanent and temporary. So the question of salt water

intrusion is truly an economic one. As too little is known about the relation between saline and fresh groundwater and about the mechanism of recharge and pumping in these areas, it is easy to recommend basic research geared at solving practical problems. The result of such research may also put tools in the hands of planners for planning the use of the questioned coastal areas in the most optimal way.

Mention should also be made in this context of large aquifers that could be threatened by sea-water instrusion. The two major aquifers in close proximity to the sea are both situated in Scania: the poorly consolidated Cretaceous sands of the Kristianstad plain and the sands and gravels of the Alnarp valley. Let's have a quick look at the latter one. It consists of a deep trough in the limestone of SW Scania, trending in a NW-SE direction, approx. 6-10 km wide and 60 km long and filled at the bottom with permeable sands and gravel overlain by till and clay till. For about 150 years the groundwater in the Alnarp valley has been tapped, first by artesian flow and later by pumping. At present it plays an important role for the municipal water supply of several communities. The total withdrawal from the valley reached 18 million m^3 per year in the late 1950's and in 1971. After that it has decreased to 13 million m^3 in 1980. It has been reported that theoretically more than 30 million m^3 could be extracted annually. But a safe yield considering the risk for salt water intrusion is estimated at 25 million m^3, provided some coastal wells are abandoned. Mr Bo Leander has reported at a previous Salt Water Intrusion Meeting of the Alnarp valley groundwater system and I know that he now can give new information of the modeling and monitoring of this aquifer in order to operate the extraction at a minimum risk level. With this one large-scale enterprise for combatting salt water intrusion, Sweden is reminded of problems that are quite common in many other countries.
Now, finally, let's turn very shortly to the third reason for groundwaters with high salinity.

3) Influence from pollution

Increased salt content in groundwater can derive from local pollution from leaky storage of chemicals or improper handling of them and from leakage from manure or waste depositories. The effects on groundwater is generally of local nature but may in some cases be harmful for pumped wells. Another source of pollution, or, influence if you like, is atmospheric deposition of chlorides and other constituents from the sea or from chimneys. Such deposition is ultimately influencing the groundwater, resulting in increasing salt content. Monitoring stations in Sweden indicate an influence of this type, but not yet leading to saline groundwaters.

This very general overview of problems with intruded and fossil groundwater of marine origin in Sweden served as a mere introduction to more specific matters which will be dealt with later during this meeting. I hope, although very incomplete and sketchy, this communication has given you a framework for further discussions.

SALINE GROUNDWATER IN SWEDEN

by Helgi Lindewald[1]

ABSTRACT

About 780 wells in Sweden have to date been identified as salt-water wells, with chloride (Cl^-) concentrations from 300 to 7000 mg per liter. Characteristic is their location to elevations below the Holocene marine limit. Many of them occur in a 200 km wide zone through the middle Swedish lowlands. From some of the wells, near the coasts, the high chloride content is supposed to be caused by intrusion of recent seawater. About 280 of the salt-water wells are drilled in Paleozoic bedrock on the islands of Öland and Gotland. Most of the salt-water wells are located on the mainland, particularly along lake shores or in clay-covered valleys. Common depths of wells in crystalline bedrock range between 50 and 110 m. Probably the saline groundwater derives from fossil seawater being trapped during and after the latest deglaciation. The future research will clarify the age and the areal occurrence of the saline groundwater.

INTRODUCTION

About 80 wells, being one per cent of the water wells drilled each year in Sweden, are reported to be salt-water wells. According to the well owners there is also an unknown number of wells where the water changes from fresh to salt-water during the withdrawal.

A general need for a compilation and a key map of these wells has existed for long. During a selective collection about 780 wells (sometimes test holes from groundwater investigations) have to date been identified as saltwater wells. The selection has been based either on analysis results showing chloride content >300 mg/l or the remark "salt-water" in the records.

The main part of the data originates from the drillers well records and from the groundwater investigation reports. These records and reports are collected and deposited at the National Well Record Section of the Geological Survey of Sweden (SGU). Some of the oldest records are sometimes incomplete, lacking other information than the remarks "salt-water" and the place and the depth of the well.

The main part of the salt-water wells can be found in four areas (see Fig 1). These areas coincide with the densely populated areas in Sweden. The additional wells occur in certain places in the middle Swedish lowlands and in coastal areas.

It must be mentioned that the collection of salt-water wells does not include drillholes with really high salinity which are drilled for oil prospecting.

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[1]Geological Survey of Sweden, Uppala, Sweden

THE OCCURRENCE OF SALINE GROUNDWATER IN SWEDEN

Saline groundwater occurs in different aquifers in Sweden. It can be found in sedimentary as well as in crystalline bedrock and also in sand and gravel deposits. In the 780 salt-water wells the chloride content ranges from 300 to 7000 mg/l.

SALT-WATER WELLS IN THE SEDIMENT DEPOSITS

In the sediments the salt-water wells occur mostly in flat peneplain areas, where the salt-water bearing sand and gravel layers are covered with thick layers of clay as in Scania, Halland and the Skara plain (see Fig 2). Surprisingly enough, salt-water can even be found in some parts of eskers (eskers are the largest groundwater supplies in Middle Sweden). The salt-water yielding parts of eskers can be found as small islands or peninsulas in the lakes Mälaren and Hjälmaren. The saline groundwater occurs in the bottom layers of the eskers. That kind of eskers are situated outside of the towns Köping, Kvicksund, Enköping, Ekerö and Strängnäs (see Fig 2). In the consulting company AIB's report on groundwater investigations for the Enköping water supply there is a description of the salinity in two of the test holes.

Table 1. The salinity in the esker of Enköping

Test hole nr 1 + 5.25 m a.s.l.		*Test hole nr 2 + 1.15 m a.s.l.*	
Depth in m	*Chloride mg/l*	*Depth in m*	*Chloride mg/l*
18	*35*	*10-11*	*20*
24-25	*1600*	*16-17*	*1500*
30-31	*3000*	*22-23*	*2700*
35-36	*4100*	*28*	*4300*

SALT-WATER WELLS IN THE CRYSTALLINE BEDROCK

The crystalline bedrock areas with salt-water wells are often covered with clay, e.g. the flat coastal areas of the lakes and the valleys of Uppland. The average depth of these wells varies between 50 and 110 m. The deepest salt-water well (245 m) is drilled at Harpsund with a chloride content of 360 mg/l.

In the archipelagos outside of Stockholm and Gothenburg many salt-water wells exist. Even from the islands in the three lakes Mälaren, Hjälmaren and Vänern salt-water wells are reported.

The highest chloride content and the depths for salt-water wells in crystalline bedrock are shown in Fig 3a.

SALT-WATER WELLS IN THE SEDIMENTARY BEDROCK

There are only few areas of sedimentary bedrock in middle and southern Sweden (see Fig 3b). But in all the areas salt-water

wells occur. About 280 of the salt-water wells are drilled in Paleozoic bedrock on the islands of Öland and Gotland. In Scania the chloride content is not as high as on the islands of Öland and Gotland. The highest chloride content (7000 mg/l) has been observed on Gotland. Also the highest percentage of salt-water wells exist on Gotland. More than 25 per cent of the water wells have a chloride content >300 mg/l.

A GEOMORPHOLOGIC VIEW OF THE SALT-WATER BEARING AREAS

In the middle Swedish lowlands the gneisses and granites of Precambrian age dominate. Also small areas with remnants of Paleozoic sedimentary rocks occur. In the south of Sweden bedrock of Mesozoic age can be found. Paleozoic bedrock constitutes the islands of Öland and Gotland. The sedimentary bedrock consists of limestones, shales, mudstones and sandstones.

In the crystalline gneisses and granites the groundwater occurs in fractures and faults. The fractures can be open or more or less filled with different kinds of sediments. To understand the occurrence of saline groundwater in the middle Swedish lowlands it is useful to know the general tectonic framework of these areas. Great faultzones run through the middle Swedish crystalline bedrock. The bedrock has been lifted up or depressed along the fault lines, supposedly many times, which has caused a separation of the lakes Mälaren and Vänern into the present basins. It is supposed that after the deglaciation there has been different movements in the bedrock caused by changes of the pressure. By such a movement the over-lying water may start to move too, and may be trapped in the bedrock.

The level of lake Mälaren is today only 0.3 m above the Baltic Sea. Until 30-40 years ago there existed sometimes a reversed flow of brackish water from the Baltic Sea to the deepest basin of lake Mälaren.

SUPPOSED ORIGIN OF THE HIGH CHLORIDE CONTENT

It is generally supposed in Sweden that the high chloride contents in middle and east Sweden originate in most cases from the earlier salt-water periods of the Baltic Sea, as the salt-water wells are characterized by their location to elevations below the Holocene marine limit (see Fig 4). On the Swedish west coast earlier trangressions from the Atlantic Ocean may have caused the salination. In coastal areas, especially in the granitic regions, it is possible that recent sea water intrusions have been caused by overdraft of the wells.

During the latest glaciation all of Sweden was covered by inland ice. The deglaciation reached southernmost Sweden about 12.000-10.000 year B.C. After the deglaciation three salt-water periods have existed in the Baltic basin, the general succession of which are shown below.

Table 2. The general succession of Baltic stages after the deglaciation.

Period *Years B.C.*	*Name of lake/sea*	*Character of water*
12.000-11.000)-8.000	*Baltic Ice Lake*	*fresh**
8.000-7.000	*Yoldia Sea*	*brackish*
7.000-5.000	*Ancylus Lake*	*fresh*
5.000-1.000	*Litorina Sea*	*salt*
1.000-	*Baltic Sea*	*brackish*

*(*One Swedish scientist, B. Kullenberg, had the opinion 1954 that the bottom water of the Baltic Ice Lake must have been salt around 12.000 B.C.).*

The present chloride content in the Baltic sea varies from 2.000 to 3.500 mg/l depending on the place where the water sample is taken. The lowest values are obtained near the outlets of the rivers.

The chloride content in the fresh groundwater has been measured to be <60 mg/l in the west and south coast areas. The values are decreasing eastwards. By Lake Mälaren the main values vary between 8 and 25 mg/l.

THE EXTENSION OF THE SALT-WATER BEARING AREAS

It is impossible today to tell the extension of salt groundwater areas. The results from some of the test pumpings show increasing chloride contents, especially if the chloride values already from the first analysis are higher than usual for the area. It is not uncommon that the chloride content increases from 85 mg/l to >300 mg/l after one week's pumping and to >1000 mg/l after two weeks' pumping. Also the hardness increases together with the chloride. Available data seem to allow for introducing a correlation factor for the increase of chloride and hardness during pumping in different areas.

By personal communication form J. De Geer (SGU) a municipal groundwater well in Östervåla, about 40 km NW of Uppsala, is reported to have had a chloride content of 200-400 mg/l during more than ten years' withdrawal. During these years about one millon m^3 of water have been abstracted from the well. South of Lake Vänern in the western part of Sweden, there are two springs where the chloride content exceeds 1000 mg/l.

INVERTED CONDITIONS

In two cases fresh water is reported below saline groundwater.

In the valley of Göta river, in the village of Älvängen (40 km N of Gothenburg), there is a groundwater well in sediments. It has the following stratigraphy.

Table 3a. Inverted conditions in sediments

Depth in m	*Geology*	*Character of water*
0 - 50	*clay*	
50 - 50.5	*sand*	*saltwater*
50.5 - 61.5	*clay*	
61.5 - 62	*gravel*	*fresh water*
62 -	*bedrock*	

Perhaps this inversion has to do with some earlier landslide in the valley.
The other place is on the island of Gotland. Drilling for the water supply of the city of Visby, the chloride content at different levels in the limestone was recorded. The drilling started at +68 m a.s.l. and stopped at +2 m a.s.l.

Table 3b. Inverted conditions in sedimentary bedrock

Depth in m	*Chloride content mg/l*
30	*39*
50	*18*
60	*1460*
66	*859*

Futher research is necessary in order to find the correct explanation to these chloride values.

THE CURRENT COUNTER-MEASURES AGAINST SALTWATER

Different methods to eliminate the salt-water in a drilled well can be adopted. Sometimes the drillers make a new shallower bore-hole. Many of the drillers seal the hole with sand/concrete or only concrete to a level above the deepest water bearing fracture. Some of the drillers insert a plug of wood or a packer of rubber in the hole to isolate the deepest fracture.

CONCLUSIONS

The compilation of the 780 salt-water wells proves that the occurrence of saline groundwater is more frequent than was known before. It also points out the necessity for futher research on saline groundwater in Sweden. It is desirable for future drillings that the occurrence of saline groundwater and the fresh/salt water interface can be predicted. The determination of these waters calls for an understanding of their origin and history.

This study shows that the records, although incomplete, and the analysis results having chloride values of 40-50 mg/l above the normal values for the area must be compiled in the future. It also shows that it is necessary to have a continuous collection of data and that they must be evaluated and updated at regular intervals.

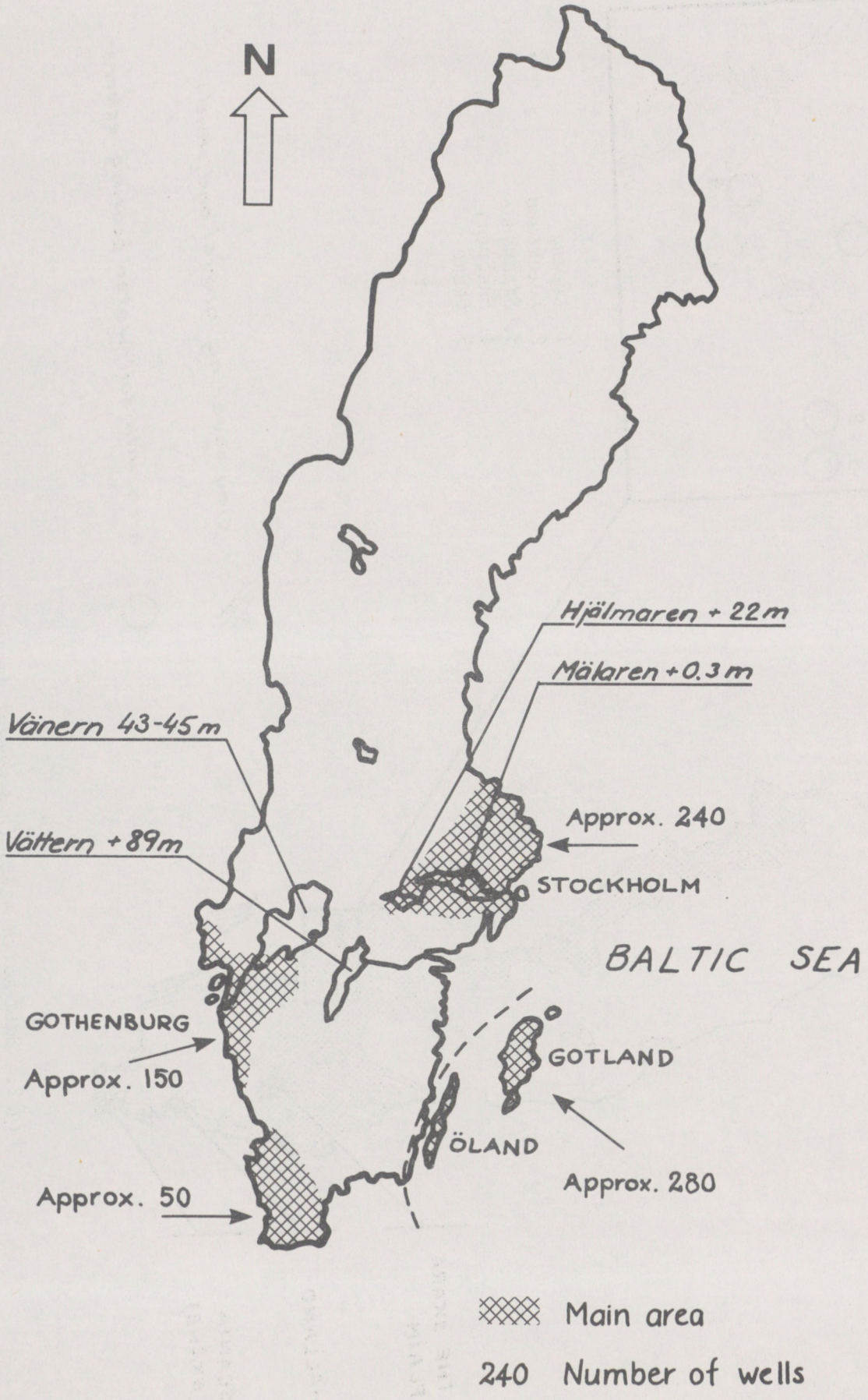


Fig. 1. The main location of salt-water wells in Sweden

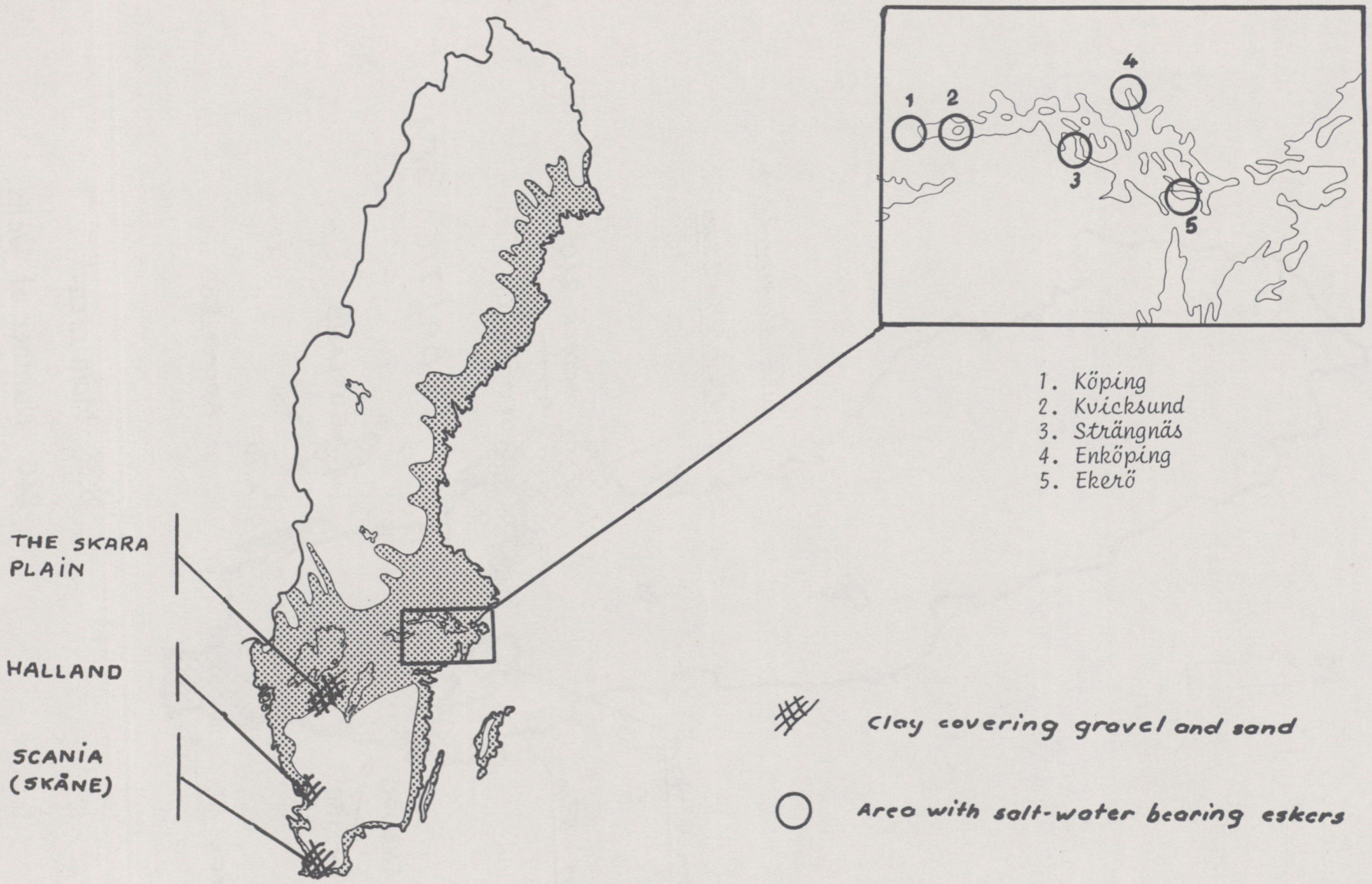


Fig. 2. Salt-water wells in sediment deposits (see also Fig. 4 b.)

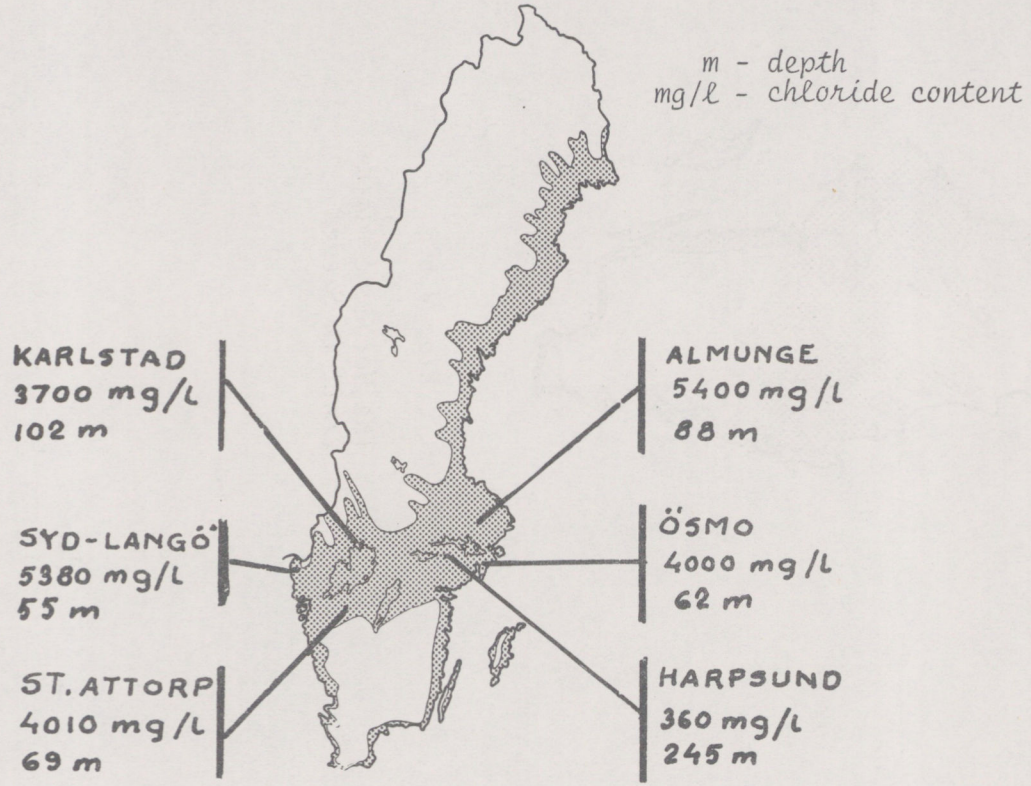


Fig. 3 a.

Major salt-water wells in crystalline bedrock (see also Fig 4 b)

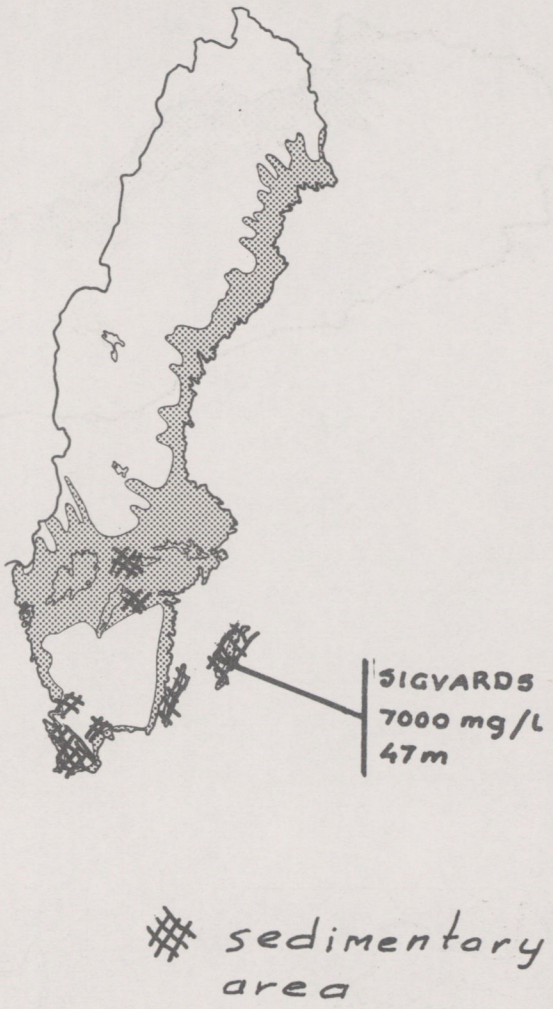
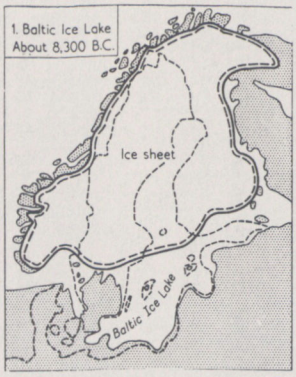
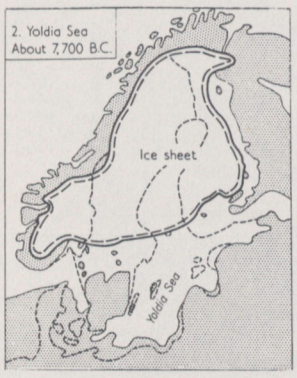
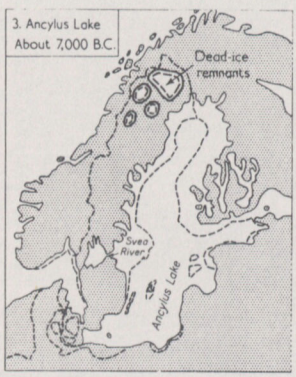


Fig. 3 b.

Salt-water wells in sedimentary bedrock







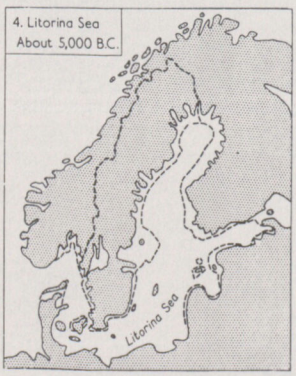


Fig. 4 a. Main stages of the Baltic Sea. (J. Lundqvist, 1965, The Quaternary of Sweden).

Fig. 4 b.

The Holocene marine limit in Sweden

SOME WELLS WITH HIGH CONTENT OF CHLORIDE IN CENTRAL SWEDEN

by P Engqvist [1]

ABSTRACT

In connection with the regular hydrogeological mapping of Sweden some wells with high chloride content have been found. The high chloride content is probably a result of marine trangressions in late-Quaternary time. The chemical composition of the water from these wells and their location is presented. To avoid drilling of salt water wells an attempt is made to present maps which predict areas where salt water may be found.

INTRODUCTION

Ground water with "high" concentration of chloride has earlier been reported from central Sweden by among others Engqvist (1969). In connection with the regular hydrogeological mapping of Sweden by the Geological Survey of Sweden more wells with high chloride content have been found. In this paper water containing more than 300 mg/l Cl, that is the taste limit, is called salt water.

As salt water wells are regarded as failed drillings by the drillers, they are not so delighted to report them. Consequently it is very seldom possible to receive proper samples and data of this type of wells. It is often enough for the drillers to note that the water has a taste of salt. These wells are often abandoned and in many cases sealed. When the well is sealed or without pumping equipment it is of course very difficult to take any samples. Even if samples are available, the analyses are in many cases incomplete especially regarding the positive ions. Sodium and potassium are missing and calcium and magnesium are reported as total hardness.

THE SOURCE OF HIGH CHLORIDE IN GROUNDWATER IN CENTRAL SWEDEN

Since the latest glaciation a large part of Sweden has been invaded by the Sea from time to time. During these periods the groundwater was in reality seawater. With the land uplift which is still proceeding in all parts of Sweden with the exception of the southernmost parts, the salt water was displaced by precipitation water which gradually flushed away the salt water. This process was successful in the topographic high, that is in recharge areas. The rapidity of the outflush of salt is of course also depending on the ground permeability and the gradient of the ground water level and the different densities of fresh and salt water. In the flat discharge areas where the more or less impermeable glacial and postglacial clays are situated, salt water was entrapped. The clay layers obstruct the salt water to get away. Not only these clays act in this way but also the Cambro-Ordovician shales.

HISTORY OF THE BALTIC

The upheavel of the land has caused that we today have ancient shorelines on high levels. The oldest shorelines are those of the Baltic Ice Lake with a maximum height of about 154 m a.s.l. When the landice disappeared from the southern parts of Sweden, the meltwater from the ice flowed into the Baltic Ice Lake. The outlet of the lake was through the Öresund and the Great Belts, at that time rivers. The water was fresh but heavily loaded. The sedimentation of the fine material resulted in varved clay.

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1 SGU Geological Survey of Sweden, Uppsala, Sweden

When the ice front receded from the northern slopes of Mount Billingen, the water level of the icedammed lake fell to that of the ocean and the former fresh water was replaced by a brackish ocean bay, the Yoldia Sea. West of Mount Billingen the tapping of the Baltic Ice Lake can be recognized in the sediments.

The sounds in middle Sweden became at first deeper and wider allowing an appreciable amount of salt water to flow eastward. But the land uplift continued, and after some time the sounds emerged from the Sea. The Baltic became once again a lake, the Ancylus Lake.

In the beginning the outlet of the Ancylus Lake was in Middle Sweden, about 105 m a.s.l. As a consequence of the land upheavel in Middle Sweden the outlets dried and the Ancylus Lake found a new outlet through the Öresund. The meltwater from the ice here and elsewhere made the level of the ocean rise. Again the Baltic became more salty. This phase is called the Litorina Sea. In Middle Sweden the highest shorelines of the Litorina Sea are situated between 45 and 55 m a.s.l. It is supposed that the salt content of the Litorina Sea in the Stockholm areas was 1.6 % to compare with 0.5 - 0.6 % of the Baltic today. The reason for that is that the sounds of the southern Baltic at that time were larger and deeper than today.

DESCRIPTION OF INVESTIGATED AREAS

From six areas nine analyses, labeled A to H, have been chosen to illustrate wells with a high chloride content, see fig. 1, 2 and 4. Analysis 1 represents recent sea water.

Östergötland

The sedimentary rocks of Östergötland have been subject to hydrogeological mapping (Engqvist 1981). About 600 water samples exclusively in drilled wells were taken. About 4 % of the investigated wells contained salt water. The highest chloride content reported was about 2 000 mg/l Cl. The main stratigraphy is shown in table 1. The thickest sediment layers are found in the north of the sedimentary area. In the south the layers are thinner and show a typical denudation border. Water with high chloride content is found both in the sandstone, analysis A, and in the crystalline rock, analysis B. The fact that salt water is found outside the sedimentary area in the crystalline rock may indicate that salt water in the sandstone is not connate water which earlier has been suggested, Thorslund 1951/1952. In the western part of the sedimentary area salt water is found in wells drilled to about 30 m a.s.l. In the central part wells with salt water are deeper than 30 m b.s.l.

Table 1. Generalized Stratigraphy of Sedimentary Rocks in Östergötland.

Quaternary deposits
Silurian shales with intercalations of limestone
Ordovician limestone, shale and mudstone
Ordovician limestone
Cambrian alum shale
Cambrian shale
Cambrian sandstone
Precambrian crystalline rock

Närke

In Närke the sedimentary rocks are of the same origin as in Östergötland. About 15 wells containing salt water have been reported. They are found both in the

sandstone and in the crystalline rock. The highest chloride content reported is 4 200 mg/l Cl (Engqvist 1976). Analysis E represents salt water wells in Cambrian sandstone in Närke. Two wells in Närke driven in waterbearing Quaternary deposits below the clay were artesian and with a chloride content of about 1 440 mg/l Cl (Engqvist 1974).

Uppland and Västmanland

The wells in Uppland, analysis C, and Västmanland, analysis D, are both drilled in crystalline rock. Salt water wells may be found in all low-lying areas, especially in the vicinity of the great lakes.

Västergötland

In Västergötland about 500 wells and watersamples have been examined and analysed by Lindberg et al (1972). The highest chloride content reported in this investigation is 1 820 mg/l Cl. About 6 % of the wells have a chloride content more than 300 mg/l. The highest chloride content is found in wells in the flat areas where the clay is thick. Deep-drilled wells near rivers and lakes are often salt too. Analysis F is taken from a well in the crystalline rock. Analyses E and G derive from wells in Quaternary deposits below clay layer. Analysis E is not typical for this region. The calcium content in relation to other constituents is too high.

DISCUSSION

In Östergötland analysis A represents wells in sandstone. Altitude of land surface of this well is about 110 m a.s.l. It is therefore possible that the Baltic Ice Lake and the Yoldia Sea may have influenced upon the composition of the water of this well. Well B is situated 39 m a.s.l. Here the Ancylus Lake and the Litorina Sea might have influenced too. High calcium values and low contents of magnesium, sulphate and bicarbonate are typical for these wells.

To demonstrate the changes in chemical composition of the fresh water diluting the salt water in Östergötland a number of wells in different aquifers at different depth have been chosen. The distribution of the chemical constituents in relation to the bottom level of these wells is shown in figure 3.

When the salt water of the Yoldia Sea penetrated the shale layers, sodium and potassium were adsorbed in clay minerals and the water became richer in calcium. After the land has emerged from the sea fresh limestone water rich in calcium replaced the salt water. Today there is a softening of the water in the shale layers as can be seen in figure 3.

The sandstone water is low in sulphate presumably because of sulphate reduction. The soft water diminishes the condition of forming calciumfluoride which results in an increasing content of fluoride.

The wells C, D and E could have been influenced by all phases of the Baltic with the exception of the Baltic Ice Lake. These part of the country were at that time covered by ice. In these wells magnesium, sulphate and bicarbonate are more dominating than in the foregoing wells.

In Västergötland, analyses G and H show low content of calcium and higher percentage of bicarbonate. This is strange as the fresh water wells in this area are often very hard. A composition of the same type as analysis F would be expected. See figure 4.

MAPS PREDICTING SALT WATER

At the Geological Survey of Sweden maps are now produced showing where salt water may be found. Consulting them, the drillers can avoid drilling useless salt water wells at high costs. Areas where "high content of chloride are not unusual in the groundwater" are marked with orange-coloured lines on small-scale maps (Engqvist 1981,a) On large-scaled maps other symbols and colours can be used (Engqvist 1981,b). The fact that the analyses and the information about the wells are stored in a computer has facilitated the production of this type of maps.

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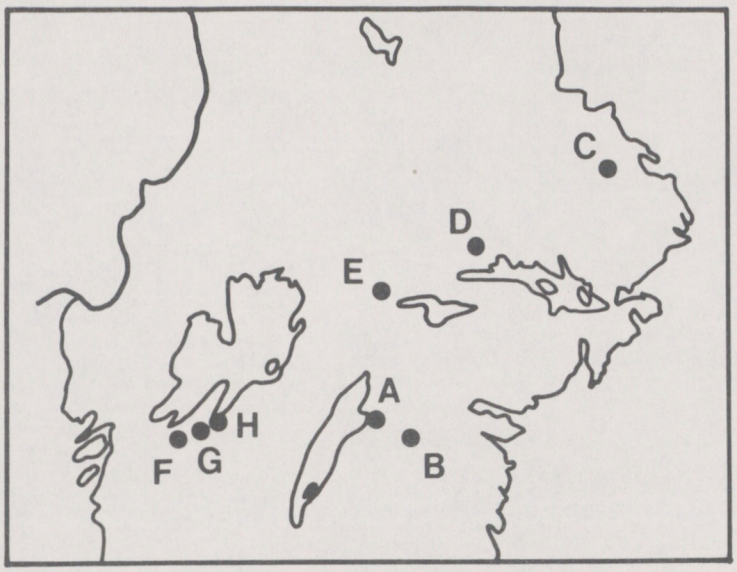

Fig. 1. Map showing sampling sites.

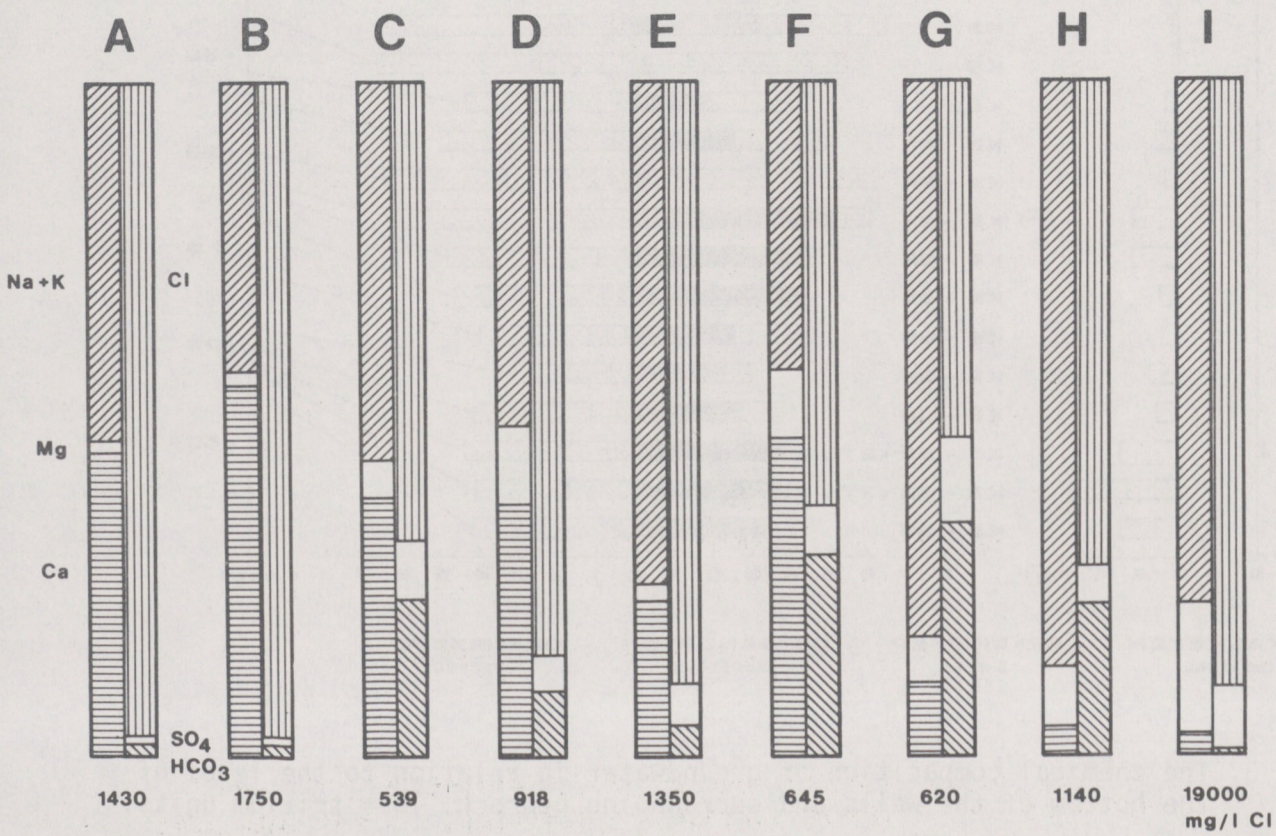

Fig. 2. Chemical analyses from different wells and the Sea.

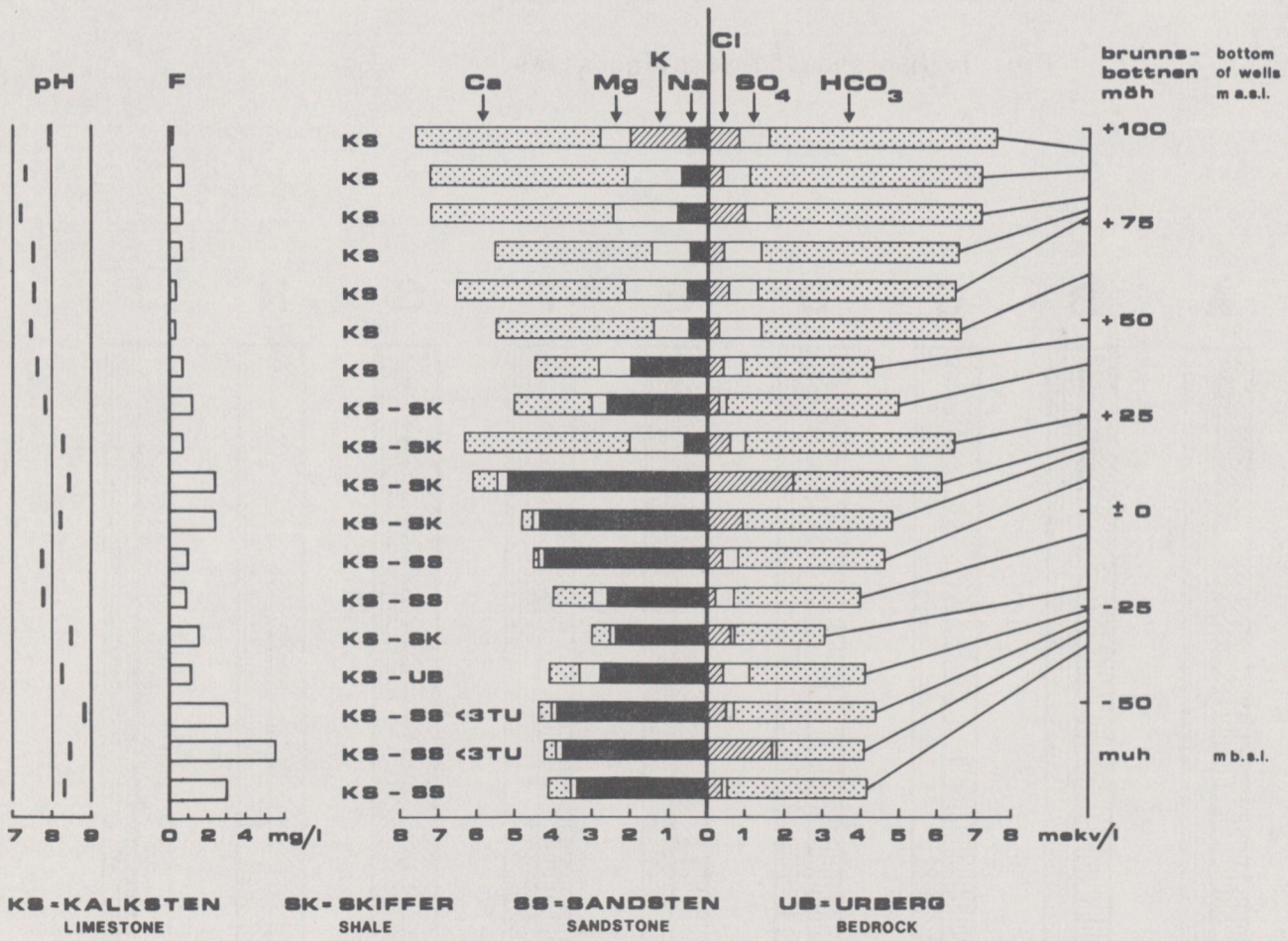

Fig. 3. The chemical composition of groundwater in relation to the level of the bottom of the wells and surrounding bedrock. TU = tritium units.

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DATUM 2.10.1981

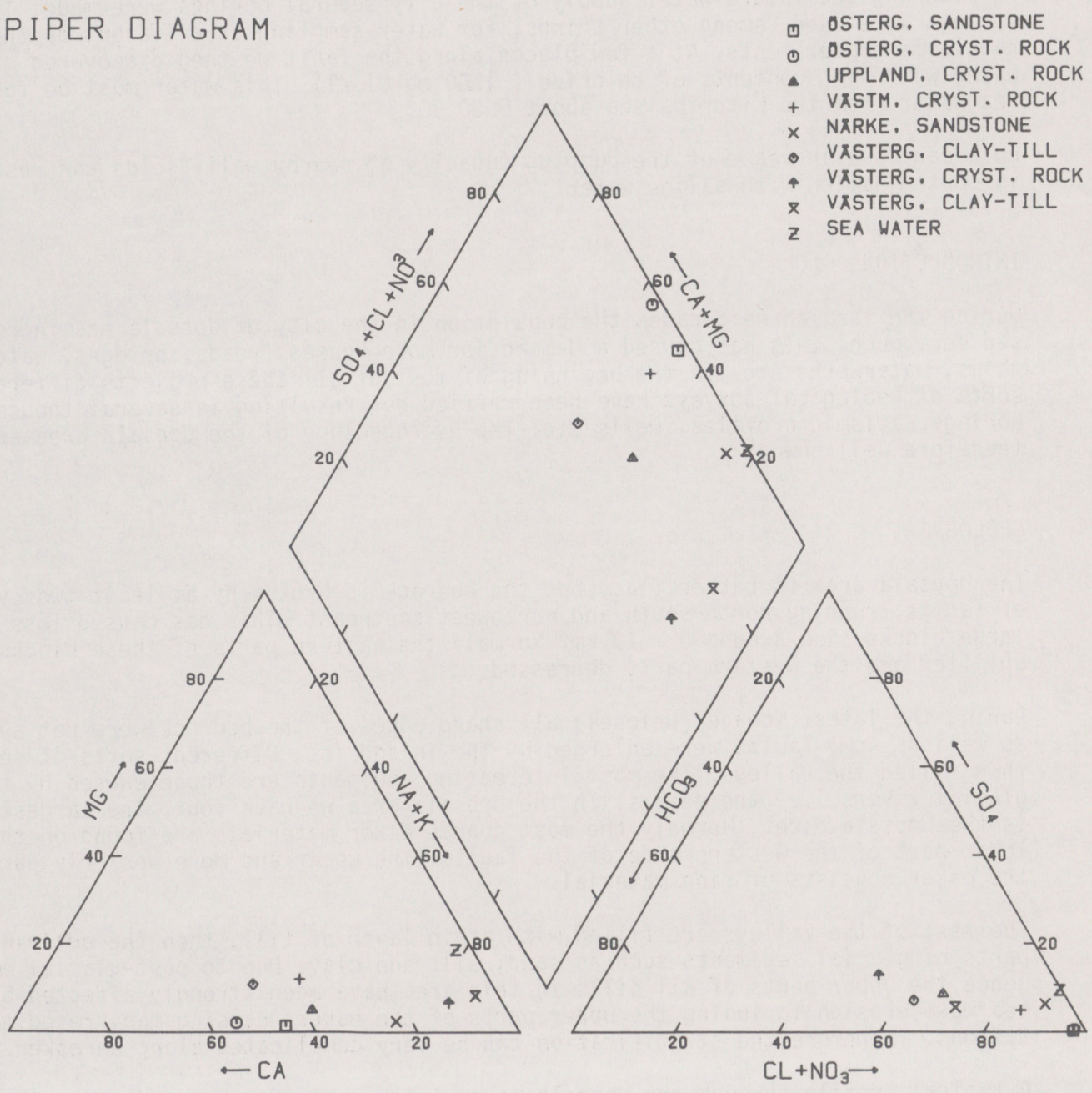

Fig. 4. Piper-diagram

FOSSIL GROUND WATER OF MARINE ORIGIN IN THE UPPSALA AREA, SWEDEN

By Dr. Jan Sidenvall[1]

ABSTRACT

The water supply for the city of Uppsala, Sweden, is based on a number of well fields and artificial recharge plants situated in the Uppsala esker. The esker has been deposited on the western slope of a large fault running through the city.

In planning the future water supply of the city several borings were made. These borings were used, among other things, for water sampling at different depths during some tracer tests. At a few places along the fault we then discovered ground water with high contents of chloride (1500 mg Cl^-/l). This water must be fossil sea water from the Litorina sea about 5000 BC.

An incautious increase of the pumping capacity at nearby well fields can result in contamination with saline water.

INTRODUCTION

During the last three decades the population in the city of Uppsala has increased very much. This has caused a demand for more houses, roads, bridges, watermains, waterworks etc. In the beginning of most of all these projects different sorts of geological surveys have been carried out resulting in several thousand borings, seismic profiles, wells etc. The hydrogeology of the Uppsala area are therefore wellknown.

GEOLOGY

The Uppsala area is rather flat, but the bedrock is broken by at least two systems of faults, running north-south and northwest-southeast. This has caused lots of large blocks size around 5 x 10 km. Normaly the eastern parts of these blocks are uplifted and the western parts depressed.

During the latest iceage (Weichel) all sharp edges of the bedrock were put away as well as some faults were enlarged by the inland ice. Different sorts of sediments then filled the valleys. The most interesting sediments are those caused by the glacial rivers i.e. the eskers. In the Uppsala area we have four. The largest one is the Uppsala esker. Normaly the most coarse esker materials are found on the lower part of the western side of the faults. The upper and more westerly parts of the esker consists of fine material.

The rest of the valleys are filled with athin layer of till, then the outlying parts of glacial sediments such as sand, silt and clay. Due to post-glacial emergence the upper parts of all hills in this area have been strongly affected by sea wave erosion including the upper parts of the eskers causing many resedimentations. Therefore the stratification can be very complicated along an esker.

A typical profile through the Uppsala esker is shown on figure no. 1. We can here see that we have esker materials from -40 to +40 meters. The fine sediments such as clay have a thickness of about 20 meter.

Just below the central parts of the city two large faults intersect. Here there are more than 150 meters vertically from the top of the esker to the bottom. Up to 130 metres of fine sediments (clayey materials) have also been found.

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1 Uppsala kommuns gatukontor, Uppsala

GROUNDWATER

The eskers in the Uppsala area can be compared with large drains in the landscape. Of course the Uppsala esker is the most important one. It is in the lower, coarse parts of the esker we have the groundwater flow. The upper parts are normaly above the groundwater level.

The main part of the groundwater in the Uppsala area is infiltrated in the till areas surrounding the large valleys. Then the groundwater will flow through cracks and faults in the bedrock as well as in the till towards the lower parts of the valleys and there we have the draining eskers. Amazingly little water are recharged directly into the eskers. On the other hand, a mathematical model has shown us that we have a vertical leakage through confining clay layers down to water bearing till or eskermaterial.

At the terminal of the groundwater flow in the Uppsala esker the flow would be about 450 l/s if nothing was pumped away or disappered through springs. Now it is not possible to pump up the whole natural flow. That would cause low groundwater levels which will start settling of buildings, mains, streets etc. Other risks are pollution and salt water intrusion.

To prevent such damages we have suggested certain upper and lower groundwater levels along the Uppsala esker. Between the levels may the groundwater levels alternate during a year. Is the level too low we have to recharge more or decrease the pumping rate at the nearest well gallery.

Figure no 2 shows these levels as well as the maximum and minimum groundwater levels noted during the last 35 years. The diagram also shows where we have our ground water well galleries and recharge plants.

THE WATER SUPPLY OF UPPSALA

The watersupply system have an old history. In the 1640-ies wooden mains were laid between a spring on the riverside (S:t Erikskällan) and the Castle. This watersystem lasted less than one century.

The modern water distribution system is a little more than one century old. Today it consists of four large groundwater well galleries, Sunnersta, Stadsträdgården, Galgbacken and Storvad. There are three waterworks. From there water are pumped out into one distribution system with two large water towers. There are more than 500 km of pipes in the system. The dimensions of the waterworks are such that one at a time can stop for months without any watershortage in the city. There are also two large and one small recharge plants (Tunåsen, Vallskog and Rosta). We pump water (up to 1 100 l/s) from the river Fyris to the recharge plants where the pretreated water percolate into the Uppsala esker. Thanks to this technique it is today possible to distribute an average of at least 900 l/s. The demand today is 600 l/s. Only 50 % of that sum is natural groundwater.

SALT WATER

During all different surveys such as testpumping, test recharging, tracer tests etc we found anomalies in the groundwater chemistry along the esker. The watersamples were taken from several testwells. In every well samples were taken at different depths. The deepest wells are about 80 metres.

All along the large faults we found salt groundwater in the silt sediments. The dephts down to this salt water varied between 10 and 50 metres. Normally the contents of chloride is above 500 mg/l. The conductivity varies between 150 and 300 ms/m. The Ca/Mg ratio is also such that it resembles that of true sea water. The only possible origin of this salt water is from seawater some time. The latest time this area was buried in salt water was during the Litorina age about 5 000 years BC. These fine sediments will for hundreds of years slowly pollute fresh groundwater by vertical and horizontal drainage.

Well below these salt water sediments there are normaly aquifers with fresh groundwater. At other places especially where the faults are unusually deep the aquifers have salt water. Such place is for example at Skediga. This place is situated about 1 500 metres upstream the well gallery at Storvad. At a depth of about 70 metres the chloride content is continously about 1 500 mg/l. The conductivity is about 450 ms/m. The total hardness is above 60 german degrees. If we now pump too much at the Storvad well gallery we might get this salt water polluting our drinking water. We then have to decrease the pumping rate or increase our recharging. Therefore this very pollution is not especially serious.

In the Uppsala area there are a few dozen similar cases. Sometime it is only a few metres down to the salt water aquifers. The most common cases with salt water are of course the bedrock wells.

To the east of Uppsala there is a village, Gunsta. At this place we have had a chloride content between 100 and 150 mg/l for several years. During the decades there has not been any decrease or increase. During years with little snow or rain the salt contents is higher. There are also variations during the year with the highest contents in May and December. In this case the salt water has been found in the fine sediments and in the bedrock.

For two villages the cloride contents were so high that new water supplies had to be found.

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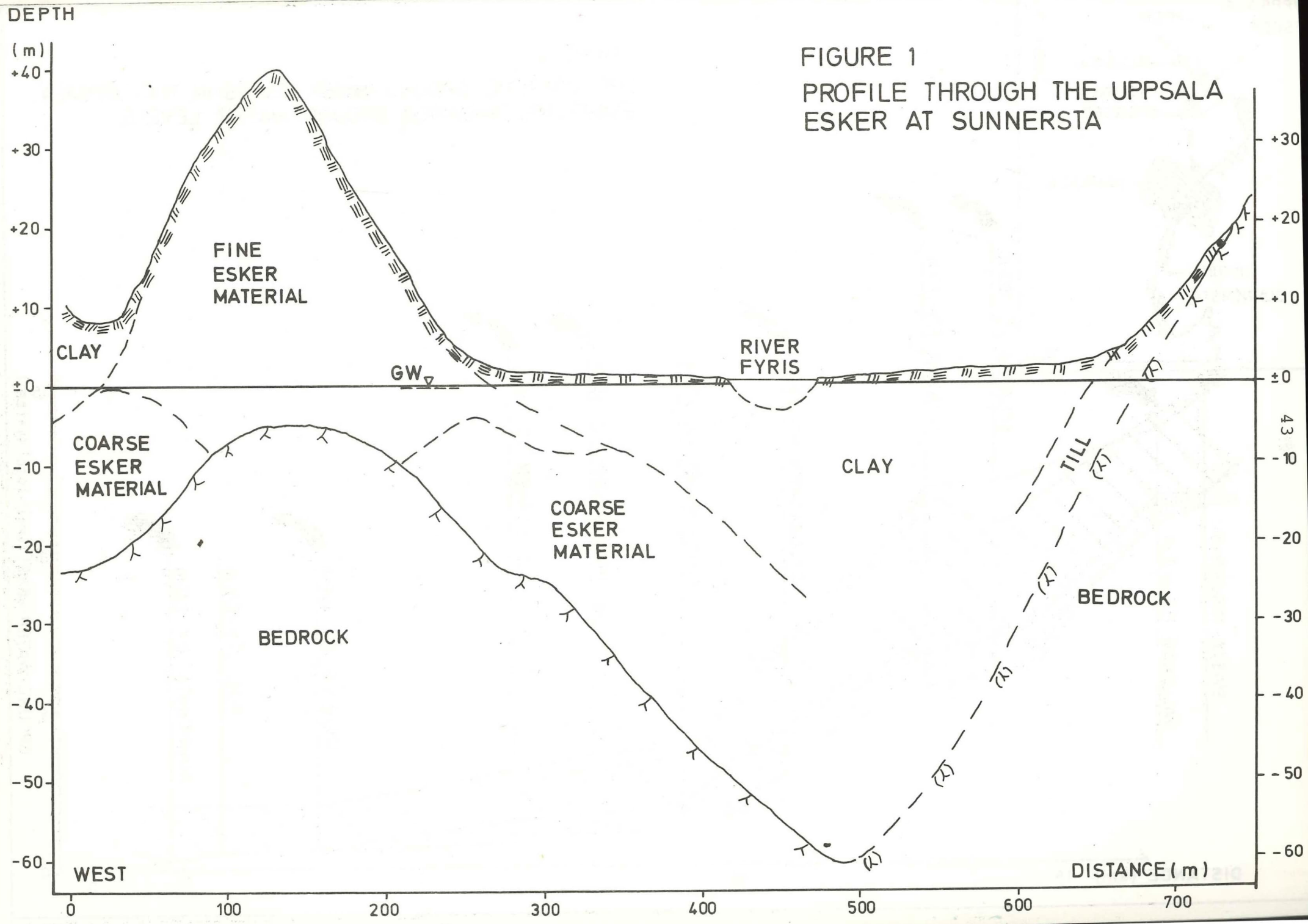
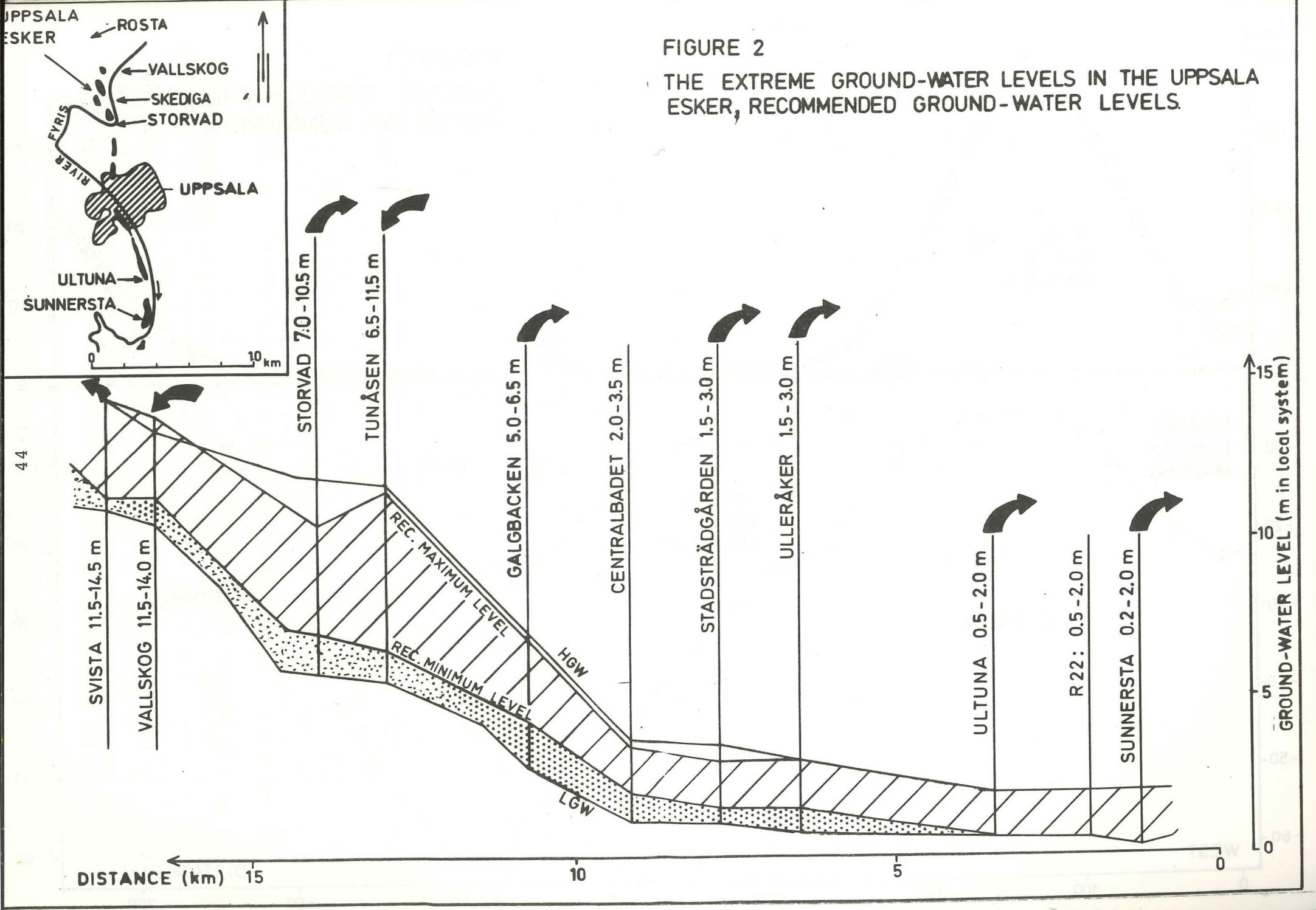

DEPTH
(m)
+40
+30
+20
+10
±0
-10
-20
-30
-40
-50
-60
FIGURE 1
PROFILE THROUGH THE UPPSALA ESKER AT SUNNERSTA
FINE ESKER MATERIAL
CLAY
GW
RIVER FYRIS
COARSE ESKER MATERIAL
BEDROCK
COARSE ESKER MATERIAL
CLAY
TILL
BEDROCK
WEST
DISTANCE (m)
0
100
200
300
400
500
600
700
43

FIGURE 2

THE EXTREME GROUND-WATER LEVELS IN THE UPPSALA ESKER, RECOMMENDED GROUND-WATER LEVELS.



SEA WATER INTRUSION IN DRILLED WELLS

by B Sund and G Bergman[1]

ABSTRACT

Under natural conditions the water table of a coastal aquifer is sloping toward the sea which implies that the hydrostatic pressure of the fresh-water is higher than that of the sea water. Owing to this and the existing hydrodynamic balance, there is an interface between the fresh and salt water that makes it possible to draw fresh-water in coastal areas.

This paper intends to show that the quality of a fresh-water aquifer bordering to the sea can be contaminated in a short period of time by intrusion of sea water when subjected to uncontrolled withdrawal thus rendering it unfit as potable water. Such contamination can take place when discharge and pumping exceed the total recharge.

INTRODUCTION

The advancing development of land for recreational and summer dwellings on islands of our archipelagoes and coastal areas gave rise to an increasing need of potable water. Such water is usually pumped up from wells drilled into the bedrock. Groundwater in the bedrock is though a limited natural resource which is usually recharges by rainfall each springtime and autumn. At several locations of our coasts an excessive overdraft is presently destroying this natural resource. In case of overdraft the pressure balance between the fresh and sea water is disturbed making it possible for the sea water to penetrate the wells. If this penetration is not noticed in time, the sea water will spread from well to well ruining the entire fresh-water aquifer. Once the seawater invaded the system, the possibilities to obtain potable water from the aquifer is completely obliterated since salts remain in the fissures for a very long time (40-50 years), notwithstanding that the previously existing balance between fresh-water and sea water had successfully been restored. In order to be able to appreciate the risks for sea water intrusion, the geological setting of the area must be studied including the existing fissure pattern of the bedrock, the possibilities for recharging the aquifer, and the magnitude of the total pumping need.

The Swedish Water and Air Pollution Research Institute (IVL) approved in early 1980 a research grant for a comprehensive study of the above outlined problems. The research object was both devised and executed by the authors. Certain preliminary studies indicated that parts of a recreational dwelling area on Vindö in Stockholm County might face problems of sea water intrusion, and therefore the main investigation was centered on that area.

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[1]IVL, Stockholm, Sweden (Swedish Water and Air Pollution Research Institute)

SOUTHEAST VINDÖ, AN AREA WITH SEA WATER INTRUSION IN PROGRESS IN THE ARCHIPELAGO OF STOCKHOLM

Programme of the Investigation

To determine how far the sea water intrusion progressed in the area the following examinations were undertaken:

I Hydrogeological mapping

a) Location of the watershed

b) Fissures, type and pattern

c) Ground water flow paths

II Examination of wells drilled in the bedrock

a) Location, depth, ground water table

b) Study of old water analyses concerning chloride content

c) Sampling of water from wells

III Laboratory examinations

a) Analysis of water of old and recently drilled wells.

Site of the Investigation

On southeast Vindö a rather large area, bordered by the sea, had been developed for recreational dwellings in the 1960s.

Potable water for the recreational dwellings is supplied by 13 drilled wells equipped with hand pump . Pumping was in the beginning limited, but as years passed the dwellings had been successively furnished with showers, water closets and saunas which demanded increased quantities of water.

Some property owners have drilled their own wells on their land, while others laid pipes fitted with electric pumps and thus pumped water from the common wells. These changes signify an increasing withdrawal from the same aquifer. A number of the property owners have now complained that water in their wells taste salty.

The Geological Setting of the Site of the Investigation

In the Vindö site the bedrock mainly consists of gneiss, known as Stockholm-gneiss. It is a medium to coarse-grained gneiss with a high mica content, and in which occur coarse-grained granit layers or lenses. In the orientation of the strike of the gneiss beds, approximatively E-W, appear dark, basic green stones dipping steeply, about 80° which means that the outcropping rock

types continue downward to considerable depth. The bedrock had been subjected during the past geological times to tectonic movements (e.g. faulting) which caused the breaking up of the bedrock into several uninterrupted fracture lines along which the newly formed blocks could move independently one from the other. The largest quantity of the groundwater occurs in these weakness zones. For that reason it is interesting to note that at the Vindö site the old zones along which movement had taken place are easily discernible. Today we can identify the largest fault zones in the shape of clay, filled roughs which intersect one the other at nearly right angles. Not so distinct fracture lines can be recognized at various places by their micacaous surfaces.

The dominant facture pattern of the Vindö site is shown in Fig 1. Of this pattern one can deduce that the bedrock is divided into 7-8 clearly delineated hillocks.

The dissected topography of the area makes it easy to distinguish several watersheds. It is most probable that each hillock constitutes a separate hydrological entity.

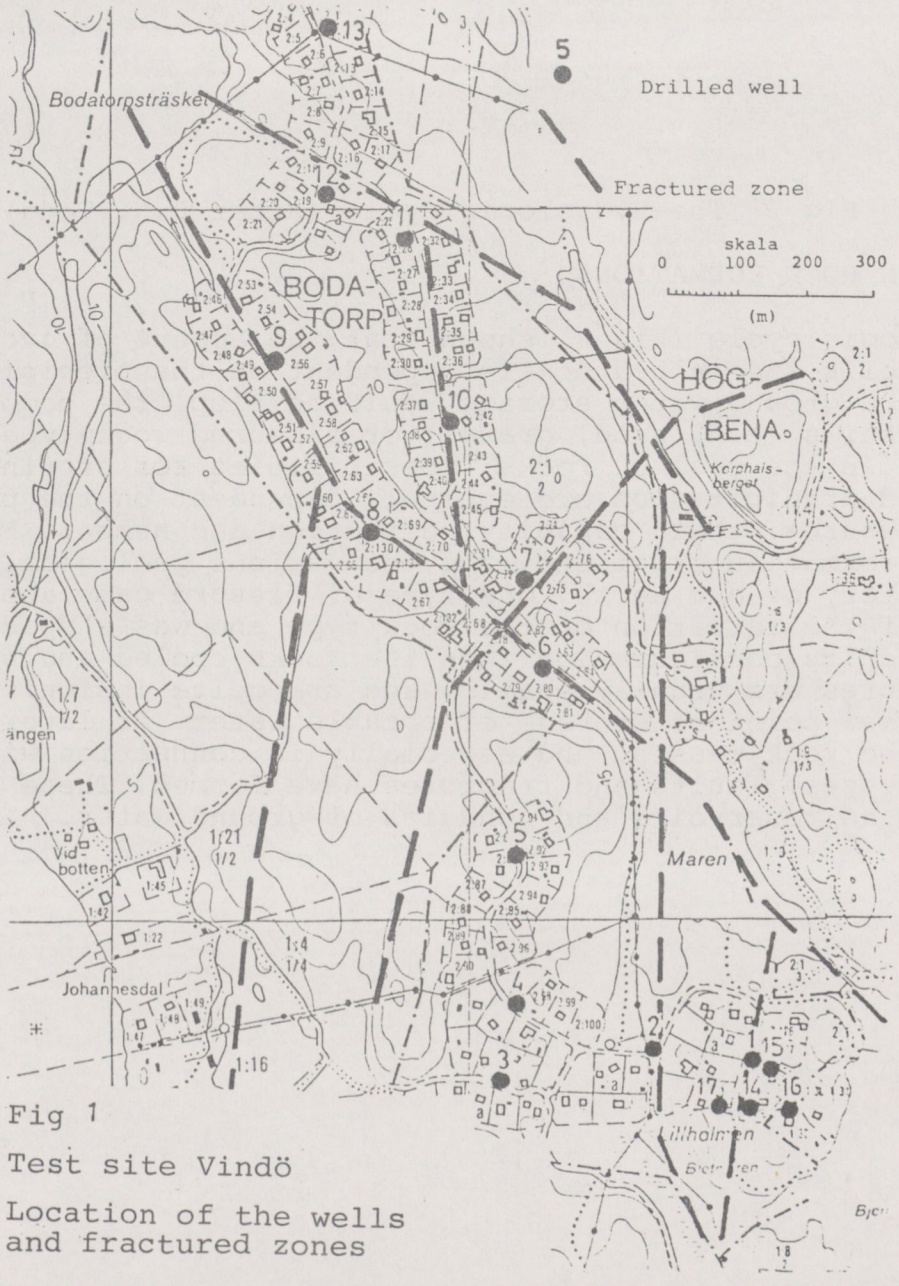

Fig 1

Test site Vindö

Location of the wells and fractured zones

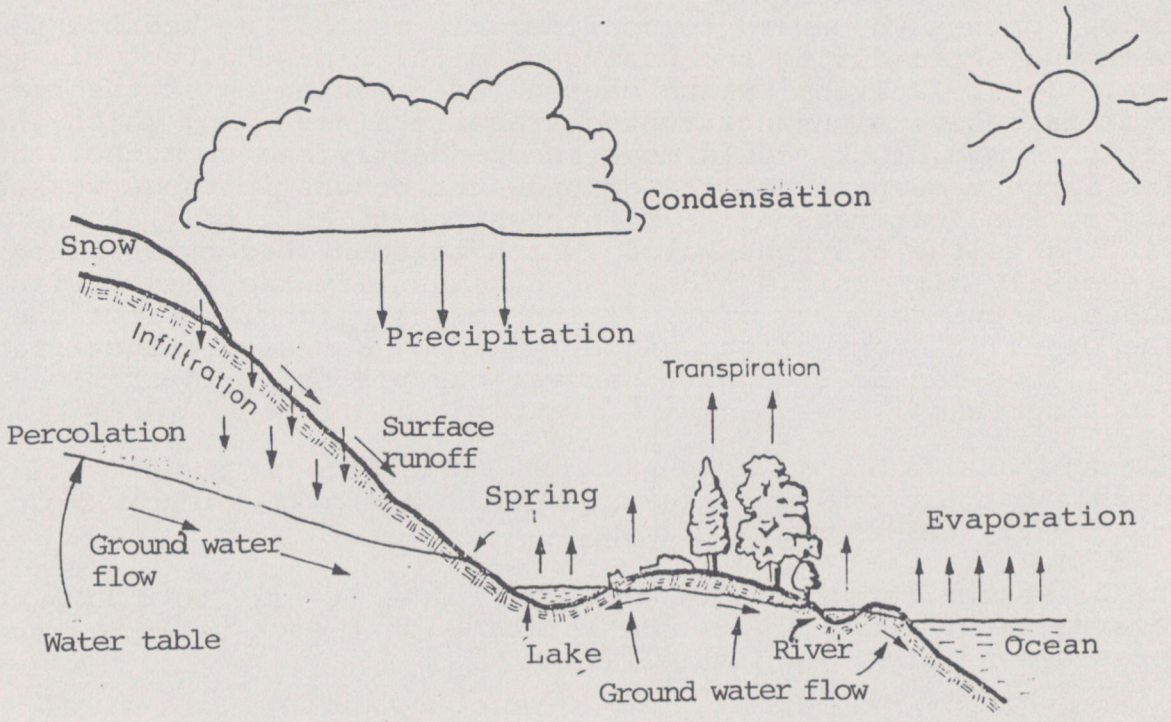

Fig 2 The hydrologic cycle (After D.K. Todd)

AQUIFERS IN ROCK FORMATIONS

The hydrologic cycle (Fig 2) shows that the amount of ground water in soils and in bedrock is dependent on the precipitation. That a ground water storage would develop, the occurrence of porous soils (e.g. sand, gravel) or fissured rock types is mandatory. In the case of rocks, possibilities for storing the infiltrated precipitation are entirely dependent on the presence of open fissures, their capacity, direction and number. Fissures in a rock formation originate from the various geological epochs of the history of our earth. One type of fissure came about at the time of the formation of the rock type and was shaped and it received its configuration when the rocks cooled and shrunk. These fissures are usually shallow and are quite insignificant from the view point of storage of potable water. At later movements of the rock masses, and especially in connection with tectonic processes, faults and fractures have formed. These are of importance as reservoirs and conduits of ground water.

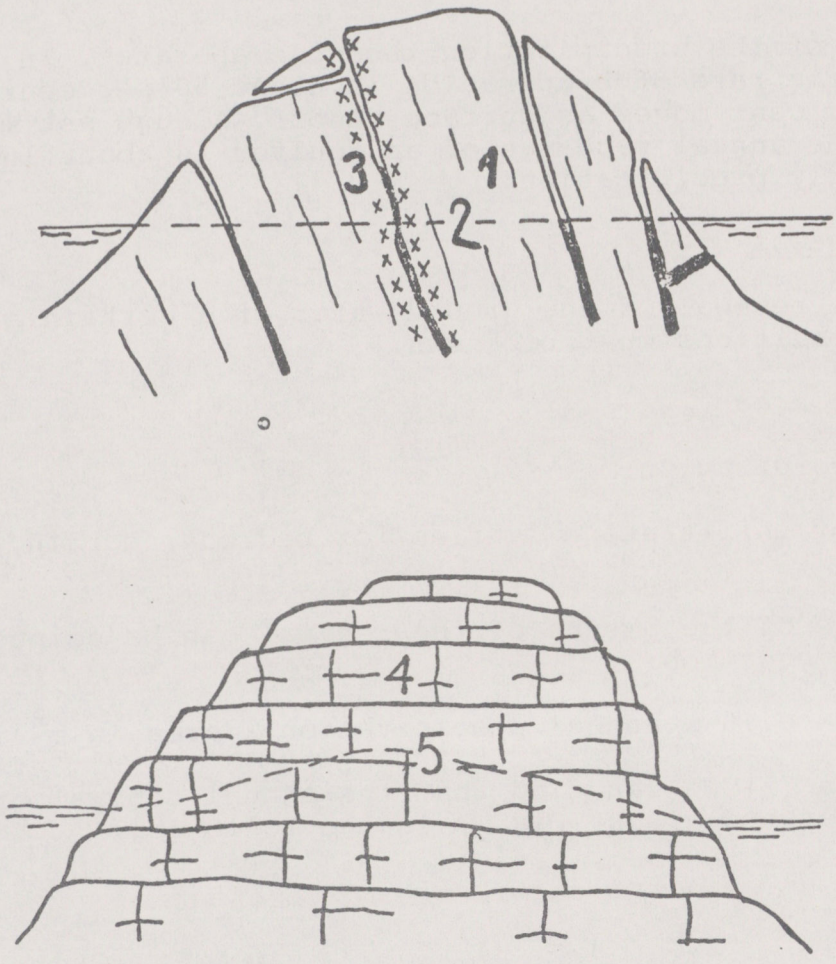

Fig 3 Ground water in rock formation

1 Gneiss
2 Ground water table
3 Ground water level in a separate fissure
4 Granite
5 Ground water table

Comprehension of the tectonic structure of a rock formation is therefore of prime concern to understand the behavior of the ground water. The fissures of a rock formation make up a system of channels connecting the separate ground water reservoirs. Already the fact that the fissure patterns in our most common rock types, granite and gneiss, are rather different, this will affect the evaluation of the volume of a reservoir (Fig 3).

Origin of Ground Water

Almost all ground water originates from precipitation that enters the ground through permeable soils and through fissures in

the rock. The technical terminology for this action is infiltration (Fig 2).

A large portion of the precipitation though evaporates. In the Stockholm area the rate of evaporation is about 50 per cent, and a further 40 per cent moves as surface runoff. A rugh estimate suggests that the annual recharge of an aquifer is about 10 per cent of the yearly precipitation.

Ground Water Renewal

To calculate the renewal of the ground water of a certain area, the following conditions must be known:

a) Extent of the area
b) Watersheds
c) Amount of precipitation
d) Runoff areas
e) Coefficient of infiltration (varies for bedrock, morain, gravel etc)

A coarse estimate of the amount of the renewal can be computed by the following equation:

annual recharge = 0.10 x annual precipitation x area in m^2 (1)

By using equation (1) we can find that the annual renewal of the six separate watershed areas (Fig 4) is the following:

Table 1

Area	1	2	3	4	5	6	Total
Ground water renewal m^3/yr	1800	5740	9230	3430	1485	2320	24,000

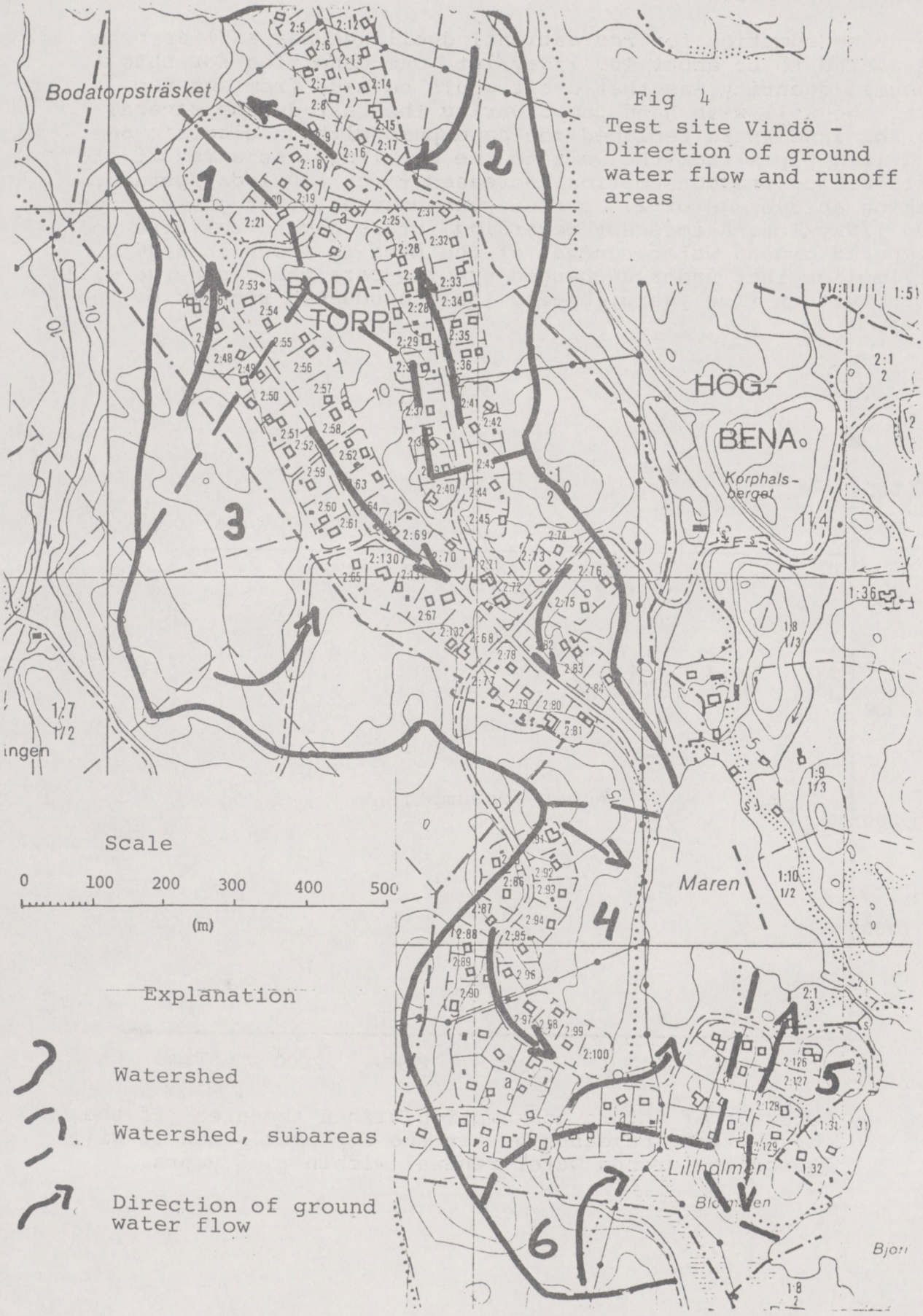

Fig 4
Test site Vindö - Direction of ground water flow and runoff areas

WATER CONSUMPTION

Water consumption for recreational dwellings has earlier been estimated to be about 100 l/day/lot. One condition for this assumed quantity was that draft would be made from one tap, i.e. one well with hand pump. During the last decade several of the landowners replaced the hand pump with an electric one and laid pipes from the well to the dwellings. As a result of this action the consumption increased to ca 150 l/day/person. Taking an average of 2.5 persons per household, it gives about 400 l/day/lot. A forecast regarding future demand relative to computed ground water renewal of the area is shown in Fig 5. It implies that under unchanged pumping rate the discharge will soon, within five years, exceed the groundwater renewal of the area.

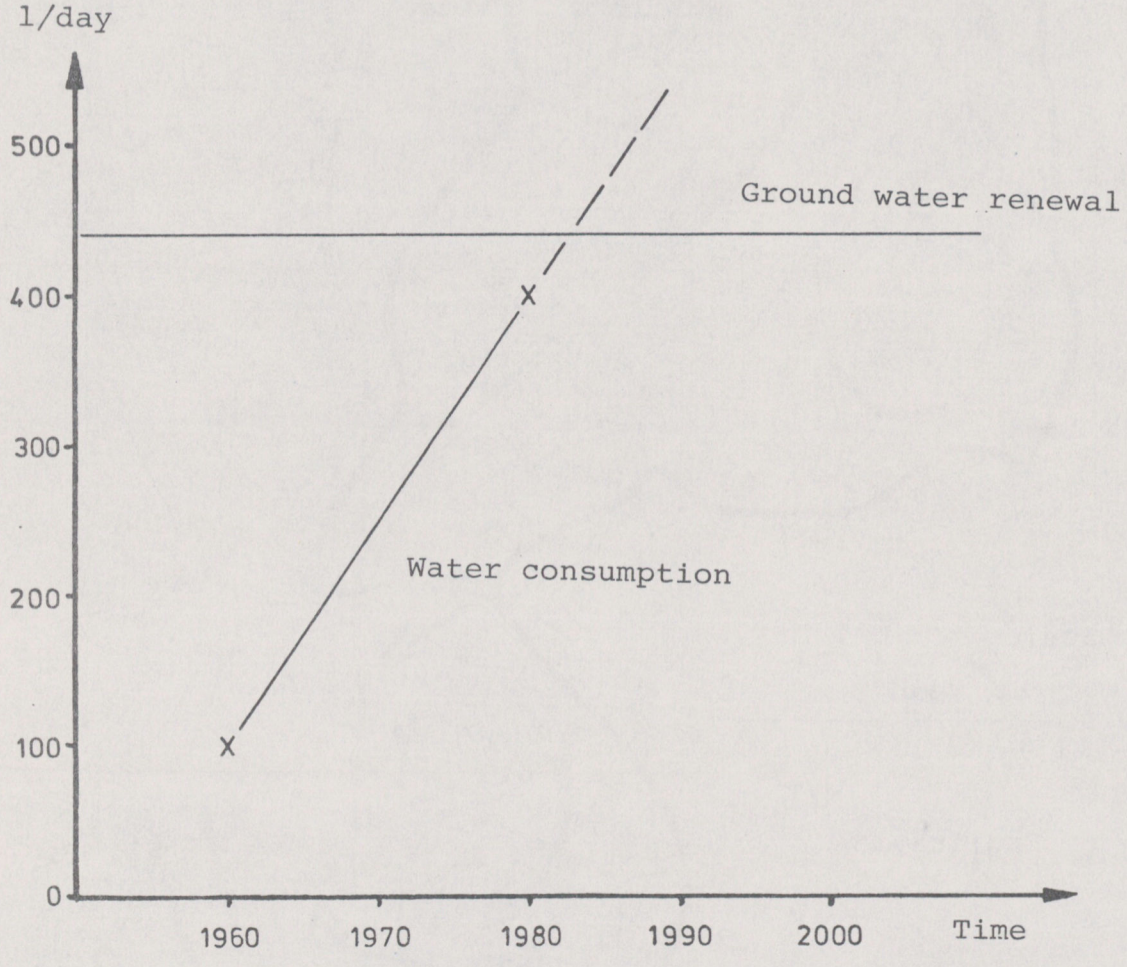

Fig 5 Forecast of future consumption within the area. If the consumption increases as per the forecast, then it will exceed the ground water renewal within 5 years.

WATER ANALYSES

Old Drilled Wells

On Vindö thirteen wells were drilled at the time of the development of the area in the mid 1960s (Fig 1).

Table 2 shows the chloride content of these wells.

Table 2

Year	1965	1965	1964	1965	1965	1964	1964	1964	1967	1967	1965	1965	1967
No	1	2	3	4	5	6	7	8	9	10	11	12	13
Cl^- mg/l	6	130	9	8	9	58	4	24	88	14	14	9	8

Results of the analyses of water samples taken at the time of the drilling of the wells or soon thereafter indicate that the chloride content was not alarmingly high. Any contact with the saline sea water was not revealed at that time with the possible exception of well no 2.

Drilled Wells Analized in 1980

In 1980 seventeen wells were sampled to check the chloride ion content. Since 1967 a few additional wells were drilled (Fig 1). Table 3 shows the result of the analyses.

Table 3	Cl^- mg/l Date of Sampling		
Well No	80-05-13	80-06-18	80-09-09
1		24	28
2	150	190	230
3		26	25
4		12	18
5		11	18
6		62	76
7		6	10
8		21	33
9		84	94
10		47	52
11		10	14
12		26	34
13		48	49
14	280		190
15		580	220
16			790
17		19	47

Chloride content of Lake Maren 1600 - 2200 mg/l
" " " the sea 2000 - 2300 mg/l

The results of the conducted analyses reveal that the chloride content of the old drilled wells became, with a few exceptions, two to four times higher than the initial values 15 years before. Since then new wells were drilled in the area. Of these four lie within the zone of danger for sea water intrusion. In three out of these four wells, high to very high chloride ion content was found. For the sake of comparison the brackish water of Lake Maren and the sea water were also analyzed as to their Cl content.

DISCUSSION OF THE RESULTS OF THE INVESTIGATION

Changes in the Chloride Ion Content

When comparing the analysis figures of Tables 2 and 3, we can notice that wells no 1 and 3 had higher Cl content in 1980 than in the early 1960s. This increase can be attributed to a slowly progressing intrusion of saline sea water into the fissures of the seashore rocks. Wells 10, 12 and 13 situated at a considerable distance from the coastline also have higher Cl content than the initial values. There appear to be several divergent explanations for this trend. Some sources support the theory that mist of condensed sea water vapor can be formed along the coastline drifting inland. This assumption, however, is contradicted by the fact that the water of well no 9 had the highest chloride content already in 1965 and it did not alter since then. Therefore the higher salt content in wells 10, 12 and 13 can be attributed to the mineral composition of the bedrock. The variation in the chloride content of well no 2 can be deduced through the analyses of the water samples taken at spring, summer and fall of 1980 (Fig 6).

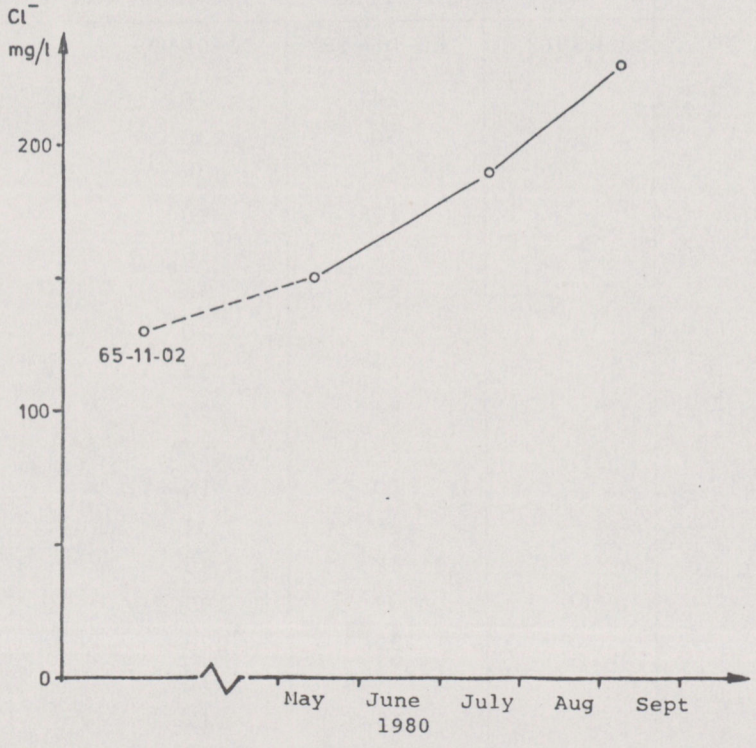

Fig 6 Variation in the chloride content of well no 2 from 1965-11-02 to 1980-09-09

In this well the influence of saline water can be traced to the adjacent Lake Maren, as the chloride content increases with increased pumpage. Fig 1 indicates too that this well must be directly connected by fissures with the coastal rock of Lake Maren. The highest chloride contents were found in the samples from the recent wells no 14, 15 and 16. All these are situated in the same area between well no 1 and the sea. On the other hand well no 17 shows no exceptionally high chloride content depending on the fact that it is located on a different hillock and is connected to other fissure systems. In well no 16 and periodically in no 15 the drinking water quality limit of 300 mg/l is exceeded (Table 3).

SEA WATER INTRUSION

The results of the analyses of the 15 year old drilled wells show that sea water has recently invaded the aquiferous fissures which in the 1960s yet carried fresh-water. Reason for this intrusion is an increased pumping from the common ground water reservoir. This increased withdrawal resulted in a lowered hydrostatic pressure in the fresh-water reservoir, by which the protection against the adjacent sea water had been weakened facilitating a gradual intrusion of sea water. When a well is pumped so excessively that it becomes empty, the piezometric surface around the well will then be lowered. A low-pressure zone is formed that nature tries immediately to equalize. The ground water from the surrounding fissure systems gushes toward the well causing a suction that propagates in the course of time longer and longer away from the well. In effect it means that in the coastal areas the dominating seaward flow of ground water is reversed, causing the sea to penetrate inland aquifers (Fig 8.). In case when the bedrock adjoining the sea has high porosity the fresh-water will slowly be forced back by the sea water and the well will be filled with brackish water. One liter of 2 percent saline water mixed with fresh-water can make seven liters of fresh-water salty rendering it undrinkable. If a well is pumped without restriction the sea water can enter the suction area of a neighbouring well making it unusable. An example for this is the Lillholmen area where five wells are situated close to each other and where the sea water intrusion is spreading from the well nearest the sea to the other wells (Fig 9). The principal risk with uncontrolled ground water pumpage from wells in coastal areas is the propagation of the sea water intrusion from well to well when the wells are so close one the other that the cones of depression intersect each other (Fig 8 and 9).

PROTECTIVE MEASURES

A forecast for the development of recreational dwellings indicates that they are going to increase in numbers and that also their use timewise is going to be extended. From the view point of water consumption it can result in a doubled withdrawal of the groundwater and an additional load on the already highly exploited ground water supplies.

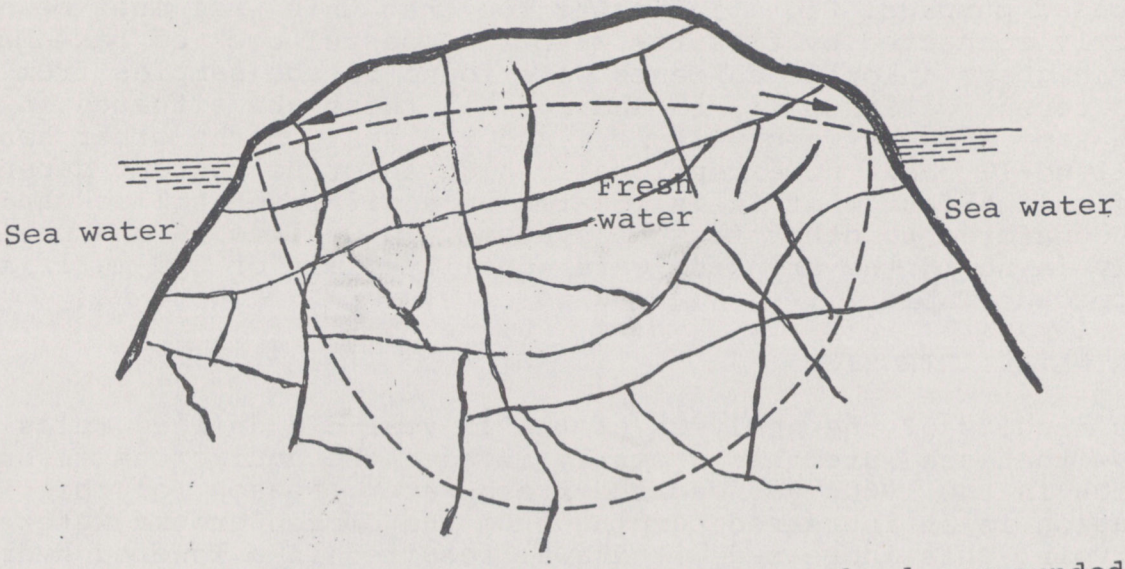

Fig 7 Fresh water reservoir in an island surrounded by sea water

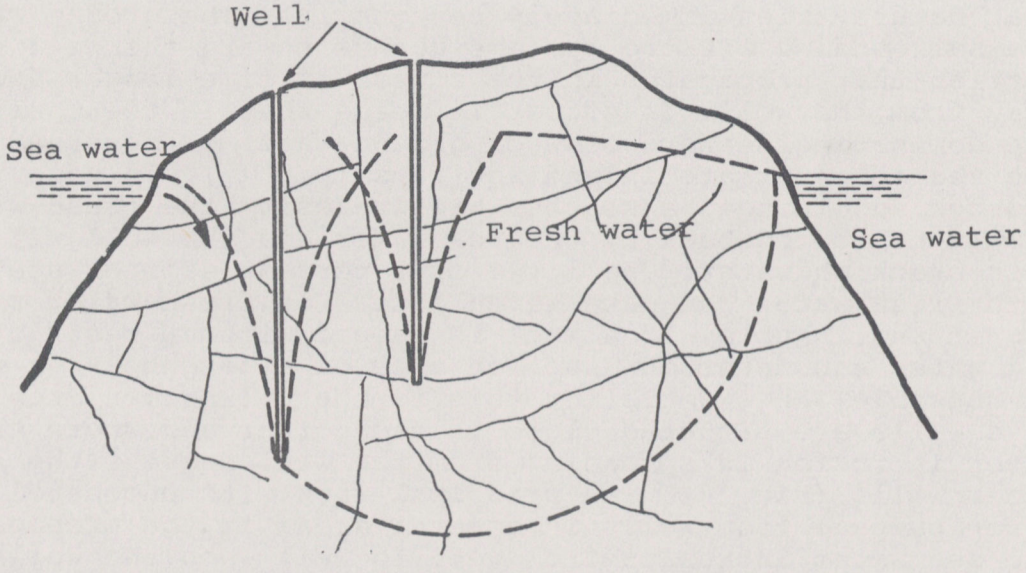

Fig 8 Cones of depression around drilled wells in an island sursounded by sea water. The lowered hydrostatic pressure changes the equilibrum between fresh and sea water facilitating the sea water to enter the fresh-water aquifer and to mix with the fresh-water. One liter of 2 percent saline water can make seven liters of fresh-water salty.

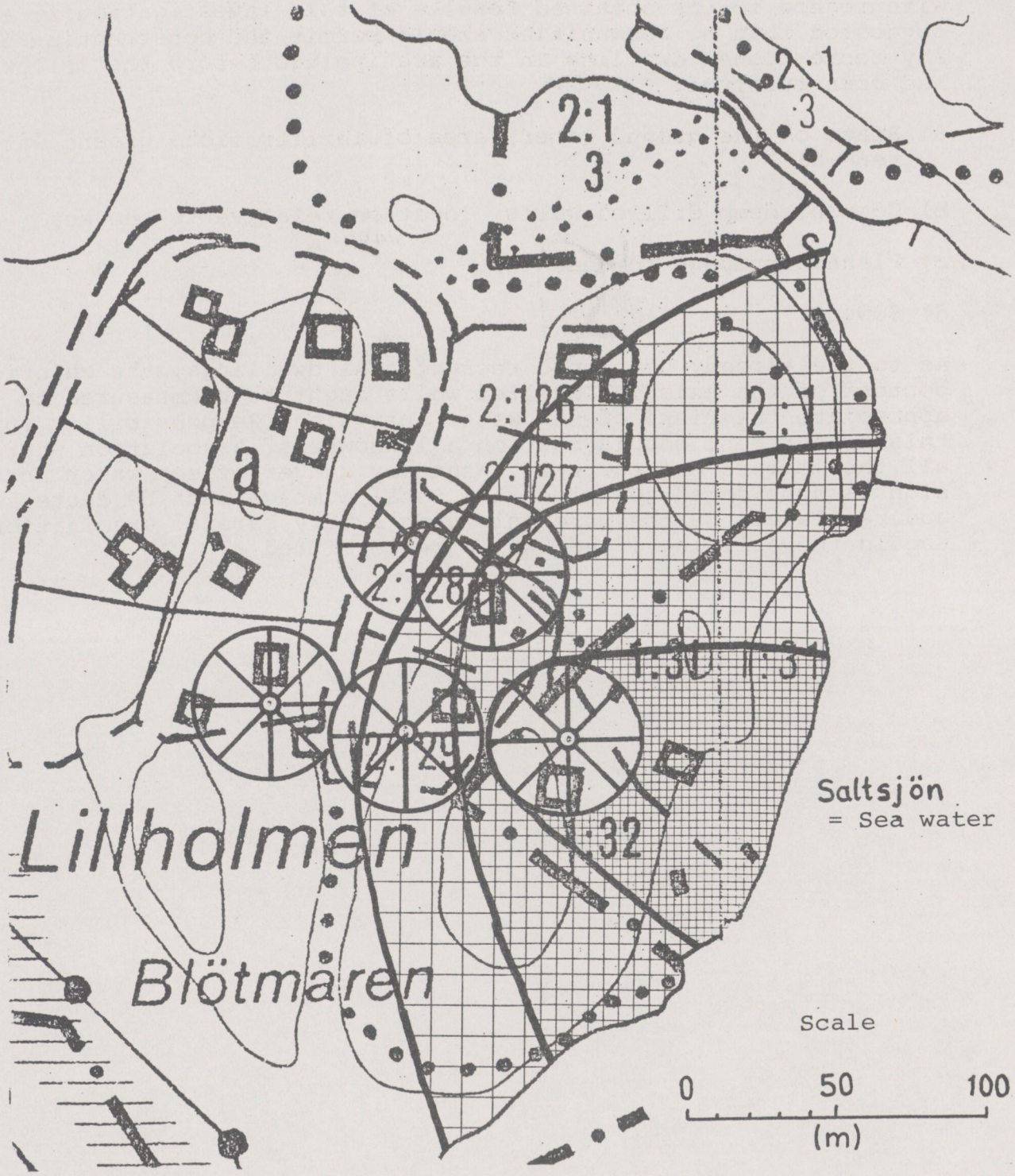

Fig 9 Vindö, sea water intrusion in drilled wells

With regard to the obtained results of this investigation it is suggested that no communities should permit the construction of any recreational dwelling in the archipelago before the following had been resolved:

a) State of the ground water, area of infiltration, ground water renewal

b) Common, deep drilled wells, location relative to the sea

c) Planned pumpage for each lot

d) Sewers.

As to the already developed recreational dwellings, the chloride content of the existing drilled wells ought to be measured to appreciate the risks of sea water intrusion. Responsibility for this measurement should rest on a landowners´ association where all the members are partners. When any danger of sea water intrusion is considered to arise, the total pumpage must be decreased and regulated. Furthermore only upon a very careful examination should the drilling of new wells be permitted.

EFFECT ON GROUND WATER COMPOSITION OF SEQUENTIAL FLOODING OF AQUIFERS WITH FRESH AND SALINE WATER

by Torgny Agerstrand[1], Göran Hansson[1] and Gunnar Jacks[2]

ABSTRACT

Marine transgression and regression in Scandinavia has caused repeated flushing of aquifers with fresh and saline water. The ion exchange properties of the aquifer material, which might be cristalline or sedimentary rock, is likely to have caused the following sequence of events:

sea water intrusion ⟶ hard saline water of Ca-Na-Cl-type

fresh water enters ⟶ fresh water of Na-HCO_3-type

continued flow of fresh water ⟶ fresh water of Ca-HCO_3-type

The Na-HCO_3-water is usually high in fluoride. In hard rocks the ion exchange capacity is limited leading to readily observable changes in composition during prolonged pumping. Sedimentary rocks allow prolonged pumping before changes are obserable. A good example of this is the Na-HCO_3-water extracted at the famous Ramlösa spa in southern Sweden.

INTRODUCTION

The events of sea transgression and regression in Scandinavia during post-glacial times have been described in detail by Lindewald (1981, this symposium). In central Sweden the following phases can be distinguished:

Baltic icelake	
Yoldia sea	9500 years B.P.
Ancylus lake	
Litorina sea	6500 years B.P.
Present Baltic	

In southern Sweden there are sedimentary strata which at depth contain brines of far greater age.

The ground water turn-over rapidly decreases towards depth. Englund & Meyer (1980) has measured turn-over times for springs in loose deposits at the order of weeks to months. In hard rocks ground water ages of thousands of years have been estimated at about 500 m depth by C-14 dating (Hultberg et al 1981). At intermediate depths there are aquifers that have been flushed just a few times during the last few thousand years. These ground waters are often encountered in hydrogeological work.

The ground water turn-over depends on available gradients for flow and the hydraulic conductivity of the aquifer. When saline water is involved the difference in density relative to fresh water is the driving force during sea transgression. The importance of the hydraulic conductivity is illustrated in fig. 1. Chloride in excess of what comes from precipitation is present only in wells with a poor yield. The figures represent wells drilled in hard rock separated only a few hundred meters from each other. The area is situated north of Stockholm.

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2 Royal Institute of Technology, Stockholm, Sweden

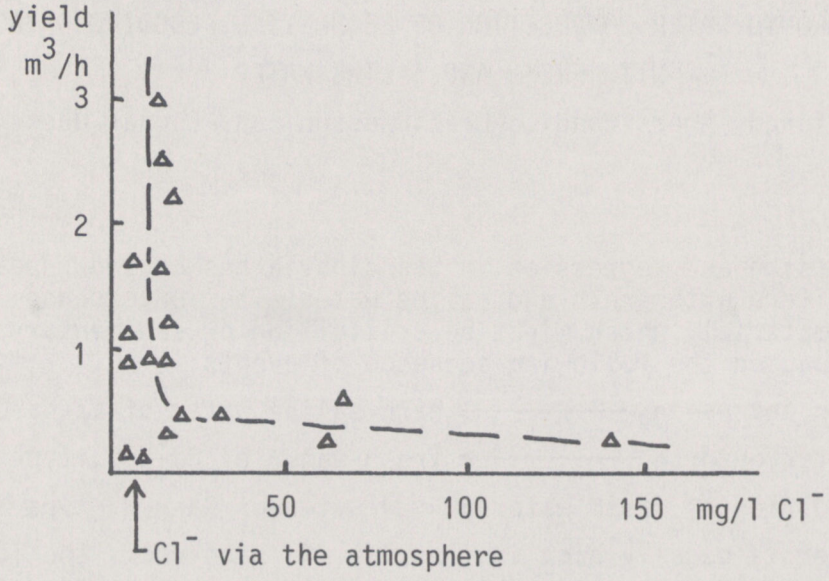

Fig. 1. Chloride content in drilled wells as a function of yield, Långsjön, recreational area north of Stockholm.

GROUND WATER IN HARD ROCK

The Mälar area

In the area affected by marine transgression during postglacial times ground waters with increased salinity are often found (Lindewald 1981, Engqvist 1981). Especially in a belt across central Sweden from Stockholm to Gothenburg where hard cristalline rocks prevail it is rather common. This includes the Mälar region.

The most saline waters are of the Ca-Na-Cl-type sometimes with a salinity exceeding that of the present Baltic. Another distinguished type of water found is of the Na-HCO_3-type. Attention is often drawn to these waters because of their excessive content of fluoride, prone to cause dental fluorosis. The common type of ground water in the region both in loose deposits and in rocks are of the Ca-HCO_3-type.Table 1 gives examples of these types of ground water.

Table 1

locality	EC	pH	Na^+	Ca^{2+}	Cl^-	SO_4^{2-}	HCO_3^-	F^-
	mS/m				mg/l			
Baltic sea	511	8,7	1060	69	2075	156	72	-
Stavsnäs	1210	7,1	1220	1410	5010	300	63	-
Finnsjön[1)]	1850	7,6	1500	1790	5650	324	44	1,1
Baggensnäs	543	8,3	105	16	23	47	257	2,4
Markim 1	827	8,3	186	13	90	55	325	3,5
Markim 2	622	7,4	22	89	31	51	289	0,7

1) Data taken from Hultberg et al (1981)

The saline ground water represented by the analyses from Stavsnäs and Finnsjön is the result of sea water intrusion during a marine transgression. The calcium in exchangeable positions on clay minerals has been replaced by sodium in the sea water, resulting in a very hard ground water of the Ca-Na-Cl-type. The Litorina sea is belived to have had a salinity of 10-15 °/oo (Ekman & Fries 1970). The Finnsjön sample is in the same range, indicating that this ground water might have remained trapped in the rock since the Litorina age. Once the sea water had entered the aquifer by means of a density flow there was no more driving force for flow until the sea withdraw from the area. Ground water formed near surface in drained areas was of the mixed HCO_3-type represented by Markim 2 in the table. The calcium is derived from calcite in the moraines and clays, and also from plagioclases in the rocks (Jacks 1973). At depth this ground water has met with sodiumsaturated clay minerals and lost its calcium through ion exchange. This process has resulted in ground water of the Na-HCO_3-type represented by the wells at Baggensnäs and Markim 1 in the table above. The considerable fluoride content is likely to come from hydroxyminerals like hornblende and micas. CaF_2 is generally the solid restricting the solubility of fluoride in ground water (Handa 1975, Jacks 1973). The low content of calcium allows a considerable amount of flouride in solution.

If the above hypothesis is correct it should be possible to produce the events outlined above by prolonged pumping, thus creating a forced ground water turn-over. So far the whole sequence has not been observed, but parts of it. The saline ground waters usually found in wells with a restricted yield do not encourage efforts to replace the saline water with fresh water by pumping.

The Rakkestad area

Near Rakkestad in southeastern Norway pumping has resulted in a rapidly changing water quality according to the proposed pattern. The area lies just below the highest marine shoreline. Ground waters found in the area are of the mixed HCO_3-type, except for one case. This well was drilled adjacent to a small lake, whose water composition is of the mixed HCO_3-type. The well had a Na-HCO_3-water when the pumping started, but the composition has rapidly shifted towards that of the lake during the process of pumping. The chloride content was slightly raised from the beginning indicating an influence of sea water. Pumping has gradually caused a decrease in chloride content. The changes in composition are illustrated in a trilinear diagram below (fig. 2).

The observed sequence of water compositions caused by a marine transgression followed by a regression can be summerized as follows:

sea water intrusion ⟶ hard saline water of Ca-Na-Cl-type

fresh water enters ⟶ fresh water of Na-HCO_3-type

continued flow of fresh water ⟶ fresh water of Ca-HCO_3-type

The last type may also be a mixed water of Ca-Na-HCO_3-type, depending on the local soil and rock composition.

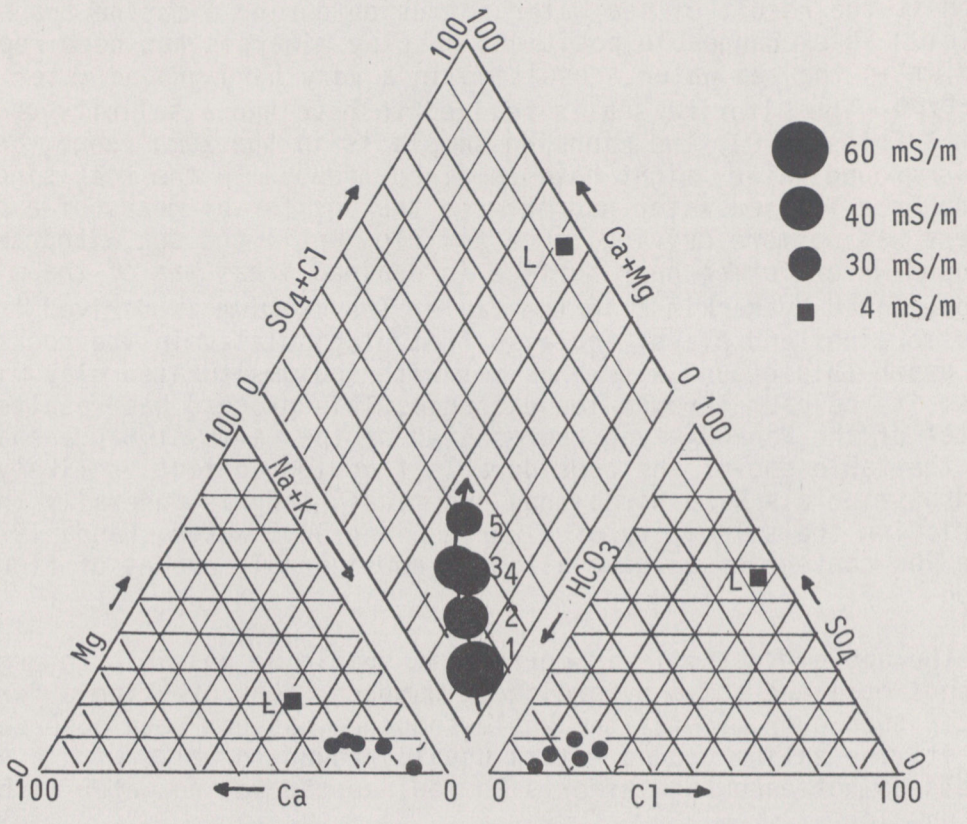

Fig. 2. Trilinear diagram for analyses from Rakkestad. 1-5 are ground water analyses from well adjacent to lake Honningen (L). Cation composition changes during pumping towards the proportion in the lake water.

GROUND WATER IN SEDIMENTARY ROCKS

Similar ground water types are found in sedimentary rocks as in cristalline rocks. A few examples will be given below from different areas.

Table 2

locality	pH	Na^+	Ca^{2+}	Mg^{2+}	Cl^-	HCO_3^-	strata
				mg/l			
Karlsborg I	-	-	30	5	23	190	algonk sandstone
Karlsborg II	-	120	4	0,5	88	160	
Ängagården, Mölleberga	7,4	1175	173	125	2270	415	
Hyby, SGU monit. well[1]	7,7	83	33	39	69	407	quaternary sand
Well 40[2]	7,2	40	178	13	59	432	

1) Nordberg & Persson 1974 2) adjacent to Hyby well

There is a striking difference between the two Karlsborg wells, in that well II has a very low hardness, this in spite of that calcite cemented sandstone is found during drilling. The Na-HCO_3-character is most likely explained by an ion exchange process.

Ängagården, Hyby and well 40 are all found close to each other in the Alnarp valley, a sandfilled trough in the cretaceous chalk in Scania. This aquifer is described by Leander (1970). The sites are far from the sea excluding a recent sea water intrusion. While well 40 represents a normal composition of ground water found in the calcerous sand, the Hyby well has a pronounced Na-HCO_3-character.

The famous Ramlösa spa is also situated in Scania, in silt- and sandstone intercalated by clay. The strata are of rät-lias origin. The Ramlösa ground water is an almost pure Na-HCO_3-water. Fig. 3 shows the result of a test drilling in the Ramlösa area. While the ground water near the surface has a considerable hardness, this gives way to a typical Na-HCO_3-water at depth. Test pumping revealed strata of lower permeability between 80-90 m depth. This aquitard thus is the"roof"of the proper Ramlösa aquifer. In contrast to the aquifers in hard rock this aquifer has an extensive ion exchange capacity, enabling heavy pumping without causing a change in the water composition.

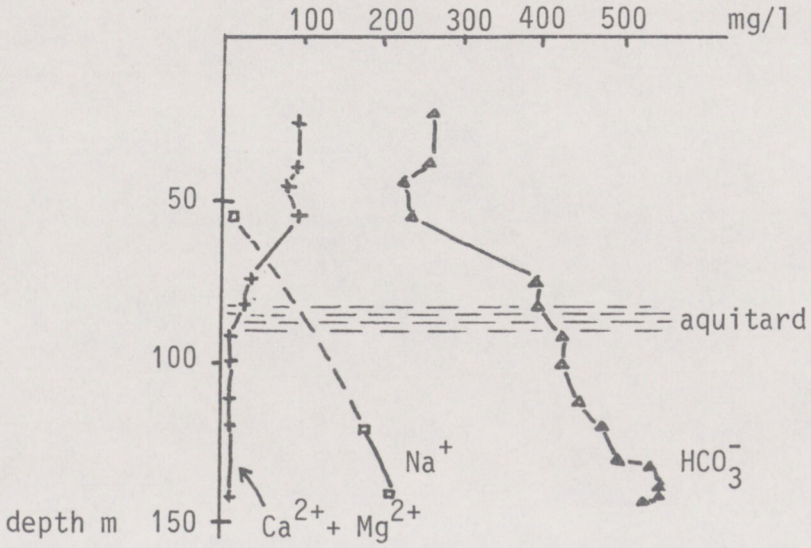

Fig. 3. Water quality profile for test drilling near Ramlösa spa.

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SPATIAL SALINITY VARIATIONS IN THE NETHERLANDS

by J.A. Boswinkel[1]

ABSTRACT

After a short review of the geological history of The Netherlands some aspects of groundwater salinity variations in The Netherlands are presented. Attention is paid to the processes involved in the redistribution of fresh and salt groundwater as well as the detection methods, used to determine the depth and thickness of the transition zone from fresh to salt groundwater.
On the base of the available data, the regional distribution of fresh and salt groundwater is discussed and some characteristic vertical chloride profiles are explained in relation to geology and hydrology.

INTRODUCTION

In the Netherlands, the chloride content is generally used as an overall single parameter for salinity. For that reason investigations related to salinity are generally focussed on the chloride content and its variations. The variation in groundwater chloride content with depth is amongst others related to geological history, lithology and geohydrological processes (natural or induced by human activity).
A qualitative approach of the processes involved can be given about the regional and vertical distribution of fresh and salt groundwater.

GEOLOGICAL CONDITIONS

The origin of the distribution of salt water in the underground can be deduced from geological history (figure 1). As The Netherlands is situated at the edge of the North Sea basin, from the beginning of the Tertiary period large amounts of sediments have been deposited due to tectonic subsidence. During Tertiary and Old Pleistocene ages mostly marine sediments were deposited. After regression of the sea in the beginning of the Pleistocene the depositional zones of both Rhine and Meuse as well as the North German rivers extended in western direction and until about 200.000 year ago sediments have been deposited by these braided rivers. At the end of the Mid-Pleistocene period there was an extension of the Scandinavian ice sheet, which covered also half of the Netherlands (Saalian glaciation). This ice sheet caused both valleys and ridges in the central part of the Netherlands.
In recent periods the sea transgreded several times the western and northern parts of the Netherlands and partly the valleys originated from the Saalian glaciation, depositing marine sediments (Eemian and Holocene transgressions). Also dunes were formed along the North Sea coastline.

PROCESSES MODIFYING FRESH- SALT WATER DISTRIBUTION

Bearing in mind that marine sediments originally have been filled with salt water and continental sediments with fresh water one would expect a distribution of salt and fresh water corresponding to the distribution of marine and continental sediments. In reality, however, there are considerable deviations from this simple concept (figure 1). Some examples of the processes causing these deviations are given below.

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[1] Groundwater Survey TNO
Delft, The Netherlands

The depth of the interface between fresh and salt groundwater may change through displacement, due to difference in density of fresh and salt water. An example is given in figure 2. Rising of the fresh groundwater level results in a lowering of the interface between fresh and salt water. In the relation between fresh water and sea water it means that if the water table of fresh water is 1 metre above the sea level, the interface will be about 40 metres under this seawater level (based upon considerations of hydrostatics). This is the wellknown principle of Ghyben-Herzberg.

In reality there is not always a sharp interface, but often there is a transition zone from fresh (chloride content < 150 ppm) to salt (chloride content > 1000 ppm) water.
Where fresh and salt water meet, there will be a migration of chloride ions, tending to level the concentration difference in chloride content in both types of water. This process is called molecular diffusion.
Moreover, due to flow of groundwater this contact of fresh and salt water is very dynamic. The combined action of molecular diffusion and movement of groundwater is called hydrodynamic dispersion.
In consequence of this dispersion a gradual transition from fresh to salt groundwater is usually observed.

The depth of the transition from fresh to salt groundwater can also be influenced by compaction. Sediments will compact if subject to a load and there will be a decrease of volume. Water will be driven out and displace or mix with the water of surrounding sediments.

DETECTION METHODS

In order to gather information about the distribution of fresh and salt water in the subsurface, not only chemical analysis of groundwater samples can be made, also physical properties of groundwater like its electrical conductivity or resistivity can be used as well.

Chemical analysis of groundwater samples gives an exact picture of the chemical composition of the groundwater at the depth the sample is taken. About the salinity of the groundwater in between the screens in one borehole, or at different locations, interpolation methods have to be used.

A powerful method for obtaining continious information on vertical salinity variations is geophysical well logging.
Geophysical well logging gives amongst others information about the resistivity of the bearing formation. This resistivity can be correlated to the total amount of dissolved solids or, as usually is done in The Netherlands, to the chloride content of the groundwater concerned, provided that accurate lithological information is available or can be inferred from other geophysical logs. For qualitative interpretation a few groundwater samples, taken at selected depths should be analysed for calibration.

A variation on this principle is the permanent electrode system, which does not only provide information in the vertical but also in the time. The permanent electrode system, permanently installed in a borehole, consists of a multi strand special cable, into which a number of electrodes are attached in pairs. Each pair forms an observation point. The electrical formation resistivity at the location of an observation point can be determined. From this resistivity the total salt content and, with some additional chemical analyses for calibration, the chloride content can be deduced.

These "salinity watchers" are used for permanent groundwater salinity monitoring where upcoming of the salt water can be expected.
Extension from the information in the vertical to a three-dimensional pattern is given by geo-electrical sounding.
Geo-electrical sounding provides information about the conductivity of the subsoil. With some additional information the resistivity of the groundwater can be deduced and a estimation of the salintiy can be made. With this method, which is frequently used in The Netherlands, the subsoil has to be schematized. The concept of a sharp interface between fresh and salt groundwater is often used.

ANALYSIS OF OBSERVED PATTERNS

With the available data it is possible to obtain an idea of the depth of the transition zone from fresh to salt groundwater as well as the thickness of the transition zone and the change of chloride content with depth in this zone.

Depth of the transition zone

The most simple approach to consider the transition from fresh to salt groundwater is a sharp interface. Of the available data a map is produced which indicates roughly the isochloride plane of 150 mg per litre (fresh/salt water map, figure 3). The choice of this chloride concentration is related to practical implications as taste and detection methods.

If we assume displacement as the main process, there will be a relation with the piezometric level.
A comparison of the fresh-salt water map with the piezometric level pattern of The Netherlands (figure 4) gives a rough resemblance in pattern. High groundwater levels coincide with a deep lying interface, low groundwater levels with a shallow position of the interface.
Only to get an indication of the order of magnitude of the process of displacement, the simple Ghyben-Herzberg relation is used, although the assumptions in this relation (horizontal flow of fresh groundwater, no flow of salt groundwater, sharp interface) are not always realistic.
In this approach the sea is used as reference level all over The Netherlands and hydrostatic conditions are assumed. Deviations will occur due to dynamic conditions, the distance to the coast, inhomogeneity and because of boundary conditions may change before the processes involved have reached equilibrium.

First we consider the eastern and south-eastern undulating and higher part of the Netherlands, as well as the higher dune area along the Nort Sea coast.
In these higher regions a part of the drainage water is removed by gravity through a system of small brooks and rivers, mostly of natural origin. The remaining part of the water flows as groundwater towards the lower regions, the rivers and the sea.
In the dunes area along the coast the average piezometric level is about 2 m above Mean Sea Level. According to Ghyben-Herzberg the interface will be at a depth of 80 m. In reality the salt water is encountered at a depth varying of 50 to 100 m below MSL. Remarkable is a slight displacement in eastern direction, most probably due to an inward direction of the groundwater flow.
In the province of Drenthe the piezometric level is more than 10 m above MSL. The position of the interface is roughly 150 to 200 m below MSL. Probably due to the very fine lithology of the sediments of Pliocene and Miocene ages below a depth of 200 m, the distance to the coast and dynamic conditions, the Ghyben-Herzberg relation (interface > 400 m) does not hold here.

In Twente the mean piezometric level is 10 to 20 m above MSL. However, the transition from fresh to salt groundwater is encountered at small depth. This is probably due to a shallow position of silty and clayey marine tertiary sediments, with a predominantly surface and shallow subsurface discharge of precipitation and hardly any deeper infiltration.
In the central part of The Netherlands in the region called the Veluwe the mean groundwater level is up to 20 m above MSL. According to Ghyben-Herzberg the interface will be deeper than 600 m below MSL. In reality this position is not known. Here again we find a slight displacement of the fresh water in the direction of flow, especially in north-western direction.
In the Utrecht-hill area the piezometric level of about 4 m above MSL has to result, according to Ghyben-Herzberg in a position of the interface of about 150 to 200 m below MSL. This agrees with the real situation. Here again is a slight displacement in north-west direction.
Finally in the Brabant Massif the piezometric level is about 15 m above MSL. The interface lies deeper than 300 m below sea level, whereas according to Ghyben-Herzberg it should be at about 600 m below MSL.
Summarising, in the higher part of The Netherlands we find mostly a displacement of the interface in a downward direction: freshening of marine sediments.

In the lower parts of The Netherlands, the hydrological regime has been changed thoroughly since 1200 AD by human activity. Before this time the area behind the dunes was drained by a system of broad slowly streaming rivers. Afterwards the big rivers have been endiked and the areas in between became polders. Former lakes partly originated as result of peat mining, were reclaimed. What has been left from a swampy flat area with some natural streams and lakes, is a complicated pattern of smaller and larger polders with different artificial surface water levels and consequently different groundwater levels.
The creation of deep polders has initiated an intensive groundwater flow. Especially in the middle of these polders there is an upcoming of the interface, due to lowering of the fresh water table. As a result near recharging rivers the interface lies deeper than under low lying polders, where the interface sometimes reaches the surface.
Most probably here hydrodynamic dispersion plays an important role too. Moreover, a complicating factor is the recent holocene transgressions of the sea. In sea-covered areas, chloride ions are transported into the subsurface by molecular diffusion.
Therefore, the assumption of a sharp interface is, especially in the lower parts of The Netherlands, not very realistic.
Summarising, in the lower half of the country there is an upward movement of the interface: salinisation of fluviatile sediments. Due to recent interference by man, there will be no equilibrium yet.

Finally on figure 3 is indicated where inversion occurs. Here the general trend of chloride content increase with depth is disturbed.
These zones with a complex fresh-salt water stratification are mainly grouped around the IJssellake. In the eastern part, they will probably be related to lateral fresh water intrusions from the high lying surrounding areas. In the province of North Holland they are most likely related with fossile water, as this part of the Netherlands has been an area with a lot of islands for a long time.

Change of chloride content with depth

In order to get an overall indication of the distribution of fresh and salt groundwater the assumption of a sharp interface was made. To get a more shaded impression the change of chloride content with depth at one location is considered.

Interpretation of geophysical well logs, supplied with chemical analysis of groundwater resulted in a number of detailed vertical chloride profiles. Three types of vertical profiles can be distinguished.
A. Displacement
B. Combined influence of displacement and hydrodynamical dispersion
C. Inversion

Displacement (figure 5)

42B-51 [1]. The vertical profile, shown on figure 5[a] is located in the dune area around Haamstede in the province of Zeeland.
The thickness of the transition zone is rather small; the increase of chloride content with depth from 100 to 10.000 ppm occurs within 20 m. This transition zone is located in marine sediments, as the base of fluviatile sediments is at about 40 m below MSL. According to Ghyben-Herzberg the interface is supposed to be on a depth of 100 m below MSL. This agrees with the depth encountered in reality.

24F-56. This location is in the dune area near Zandvoort.
The part of the profile shown in figure 5[b] deals with fluviatile sediments. The top of marine sediments is at a depth of about 180 metres below MSL. Before the development of the dunes the sea must have intruded the subsurface and salinised the fluviatile sediments. Afterwards, due to a higher fresh water level the process of displacement was generated.
As the increase of chloride content from 100 to 10.000 ppm occurs within 5 m, we deal with a real interface. The depth of this interface at about 100 m below sealevel is conform the Ghyben-Herzberg relation as the piezometric level is about 2 m above MSL.

40B-305. The location of borehole 40B-305 is situated near the town of Arnhem at the periphery of the Veluwe. Here the bottom of fluviatile sediments are at a depth of about 100 m below MSL, so the transition from fresh to salt groundwater as shown on figure 5[c] occurs in marine sediments. According to the relation of Ghyben-Herzberg this transition zone should be on a depth of about 300 metres below MSL. In reality this transition zone is encountered at a depth of about 120 m, amongst others due to the change of lithology in fine sediments. In these sediments hydrodynamical dispersion may be responsible for the slight increase of chloride content.

Combined influence of displacement and hydrodynamical dispersion (figure 6)

49F-240. In this example, situated in the western part of the province of Brabant, at a depth of about 190 m a considerable increase of chloride content (figure 6[a]) is observed. It looks like displacement in the coarse sediments. According to Ghyben-Herzberg the interface should be on a depth of 150 - 200 m below MSL.
Remarkable is the change in increase of chloride content with depth. After the

[1] Borehole number 51 on topographic map 42B.

sudden increase in the coarse sediments, there is only a slow increase in the fine grained silty sediments. In the latter the process of hydrodynamical dispersion is probably dominating.

44A-369. This profile is located in the dense river system near Dordrecht. As the mean piezometric level is at about MSL there should be no influence of Ghyben-Herzberg and the transition from fresh to salt water should be on the depth of the transition to marine sediments i.c. about 70 m. As shown on figure 6[b] we find a slow increase at a greater depth. This can be explained by the influence of the infiltrating river system near this location.

31B-111. In the polder area near Aalsmeer where the profile, shown in figure 6[c], is located, the Ghyben-Herzberg relation in reverse direction can be expected. If the fresh water table is lowered, the interface between fresh and salt water will rise. As we have here a piezometric level below MSL, there will be an upcoming of the salt water. Due to hydrodynamical dispersion a slow increase with depth is observed. The top of marine sediments is at about 160 m below MSL.

39H-119. Figure 6[d] shows an example of very gradual increase of chloride content with depth. The thickness of the zone covering the increase of chloride content with depth from 100 to 10.000 ppm is almost 150 metres. In these fine silty marine sediments (top of marine sediments is at about 70 m below MSL) most probably the hydrodynamical dispersion is dominating.

Inversion

21C-113. The vertical chloride profile, shown on figure 7[a] is located near the mouth of the IJssel in the IJssellake. Up to the clay layer at a depth of about 130 m there is a rather constant chloride content. In top there is a slight increase which may be ascribed to the salt water in the Zuiderzee, as the IJssellake was called before the dam was built (1932). Under the clay layer of fluviatile origin (depth of marine sediments is at about 190 m), there is water of very low chloride content. This has to be related to lateral flow of fresh groundwater from the high lying areas south and east of this location.

26D-42. Another example of inversion is given on figure 7[b]. Only the deeper part is shown; above there is an increase of chloride content to several thousands of ppm chloride. The higher content of chloride concentration in the upper part is related to diffusion from seawater, covering the fluviatile sediments in recent times. The top of marine sediments is lying at a depth of about 240 m below MSL. The top is freshened due to lateral flow of fresh groundwater originated from the Veluwe.

25A-943. This profile (figure 7[c]) is located in western North Holland near the North Sea Canal, at the transition from dune area to polder area.
The influence of the polder area is expressed in an upcoming of the salt water into the fluviatile sediments; the top of pleistocene marine sediments is at a depth of about 250 m below MSL. In the upper part displacement is dominating. There is a clear interface at a depth of about 100 m below MSL; salt water originated from recent transgression periods is displaced, due to higher fresh groundwater levels in newly formed dunes. The clay layer at 60 - 80 m depth, disturbs this process. Moreover, perhaps the process of compaction plays a role here. Fresh water is pressed out of this fluviatile clay layer, freshening the underlying sediments, which have been slowly saltened already.

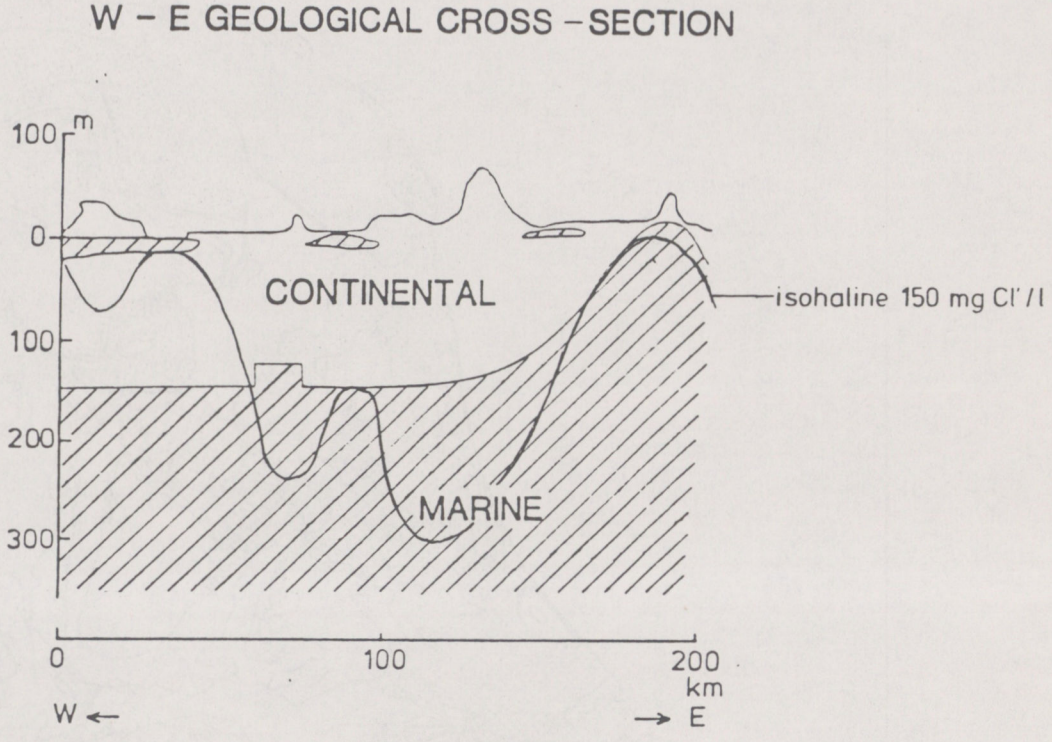
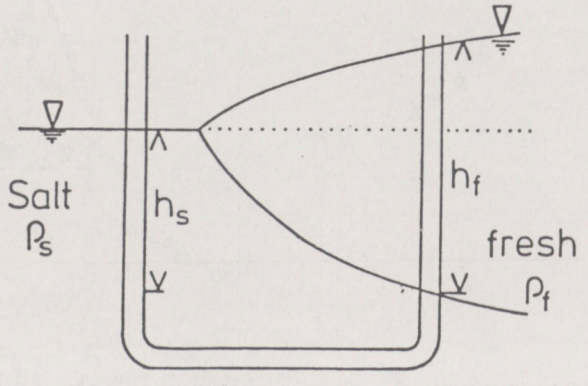
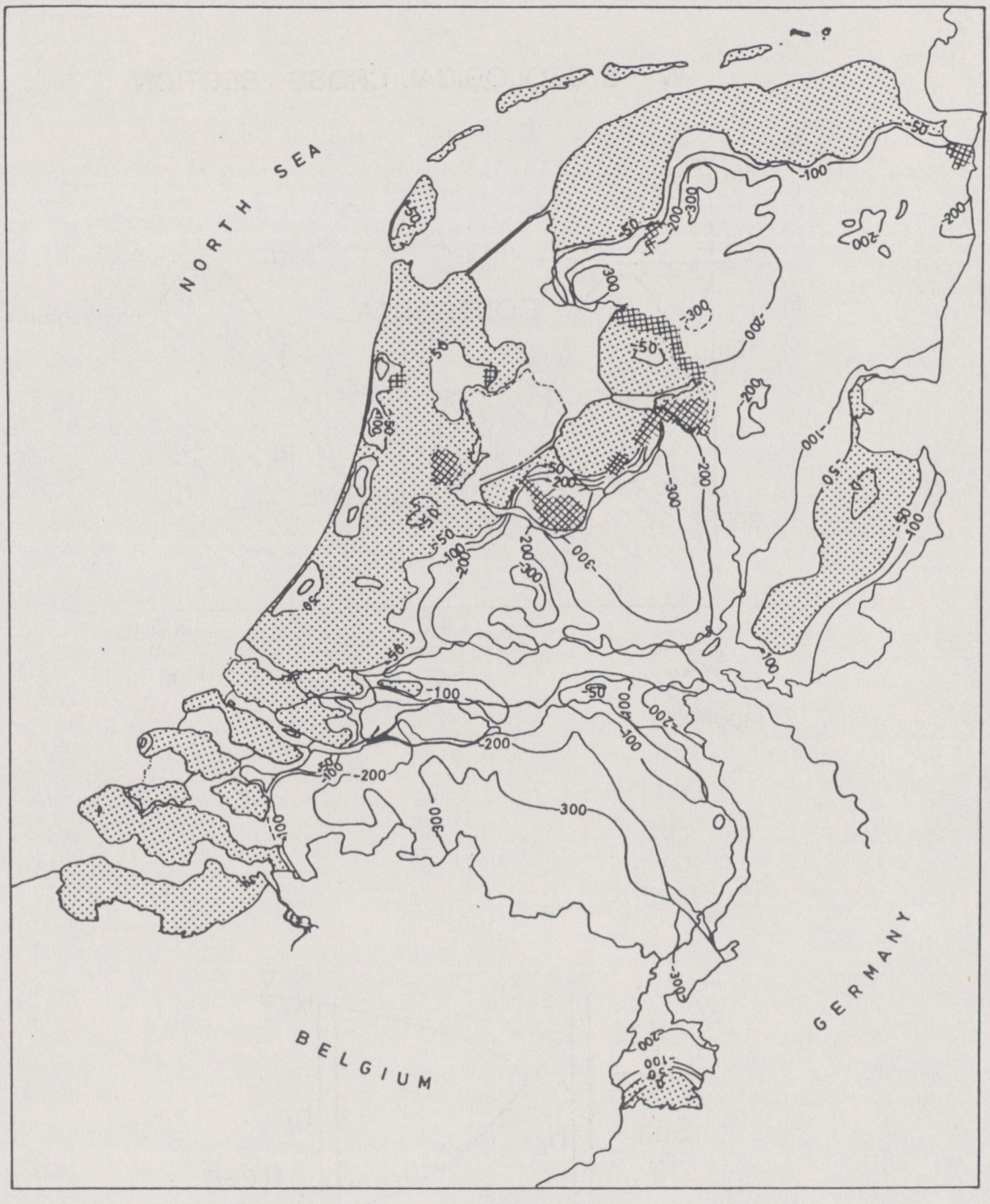


FIGURE 1



$$\rho_s g h_s = \rho_f g h_f$$

FIGURE 2



Depth to transition zone from fresh to brackish/salt groundwater (Cl' > 150 mg/l) related to Mean Sea Level

Inversion

Depth <50m below Mean Sea Level

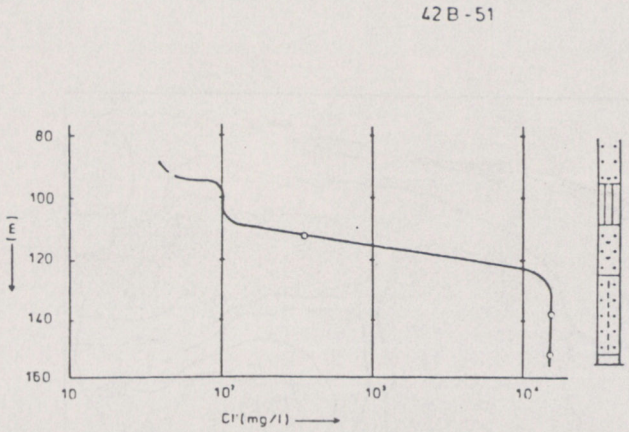
FIGURE 3

—10— Piezometric level related to Mean Sea Level (m)

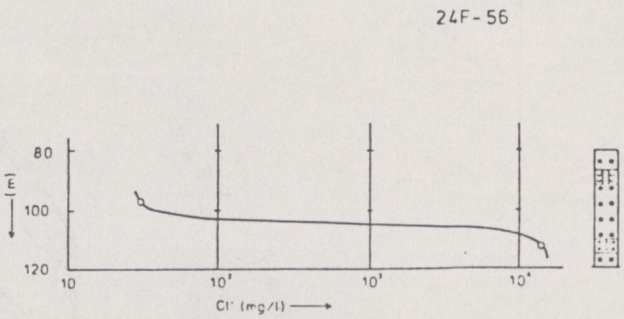
Piezometric level is below MSL

FIGURE 4

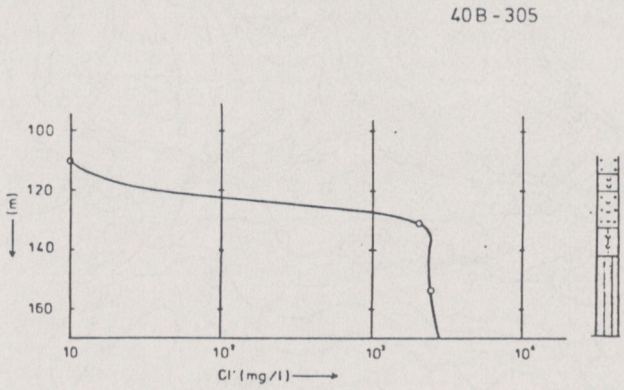
A



B



C



coarse sand

medium sand

fine sand

slightly clayey

clayey

very clayey

clay, sandy clay

alternating sand and clay strata

peat

shells

o groundwater sample analysis

FIGURE 5

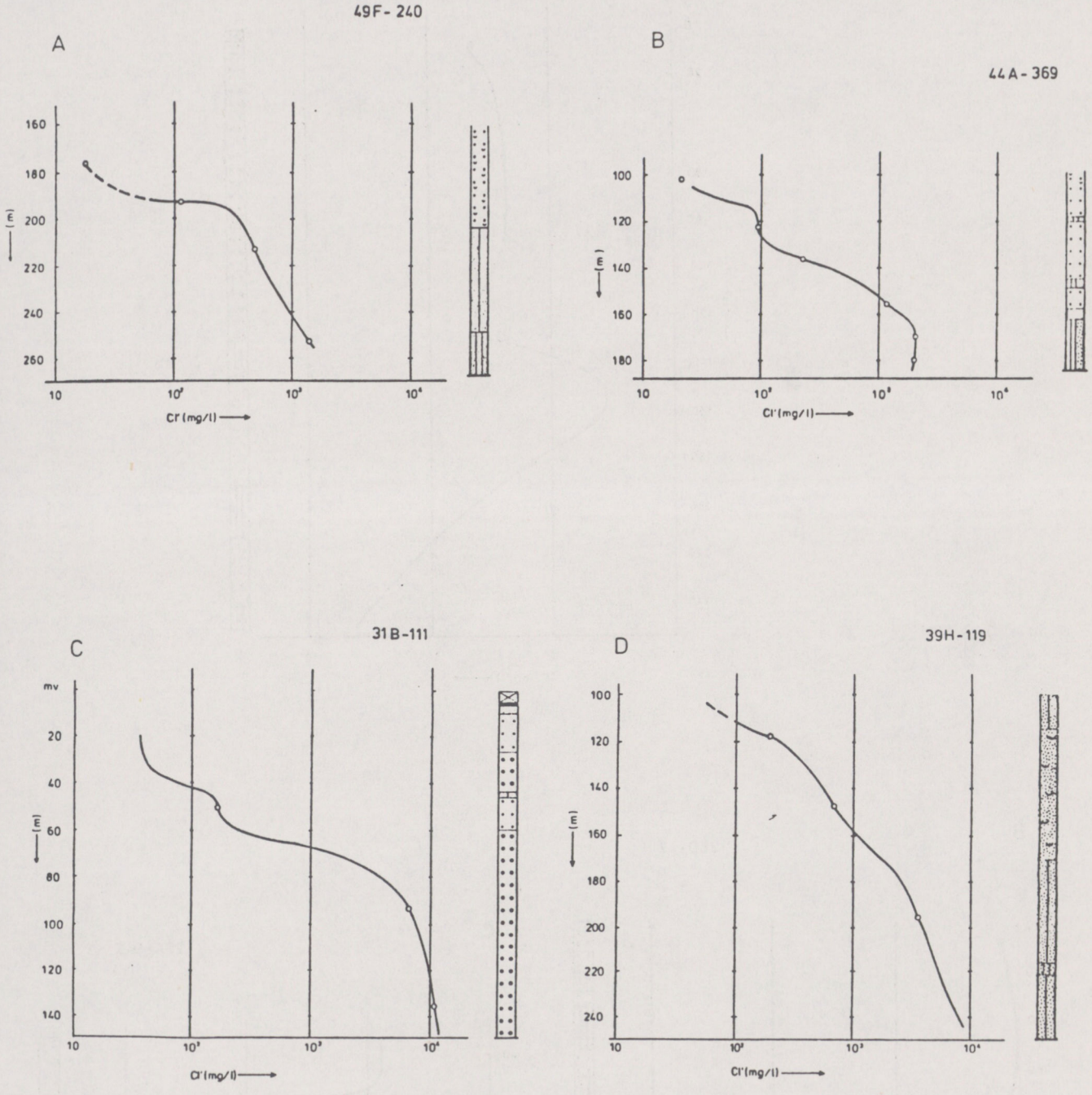
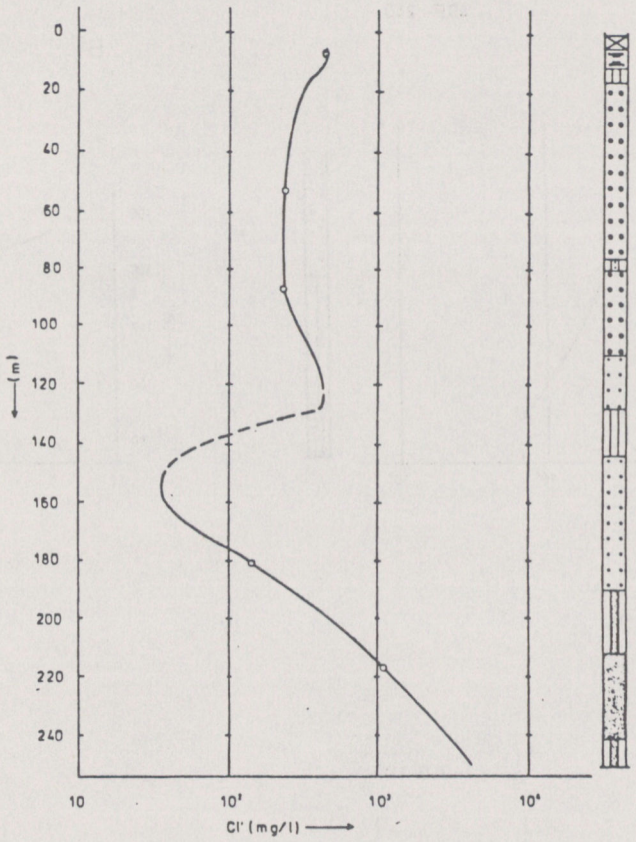

49F-240
A
160
180
200
220
240
260
(m)
10
10²
10³
10⁴
Cl⁻(mg/l)
B
44A-369
100
120
140
160
180
(m)
10
10²
10³
10⁴
Cl⁻(mg/l)
C
31B-111
mv
20
40
60
80
100
120
140
(m)
10
10²
10³
10⁴
Cl⁻(mg/l)
D
39H-119
100
120
140
160
180
200
220
240
(m)
10
10²
10³
10⁴
Cl⁻(mg/l)

FIGURE 6

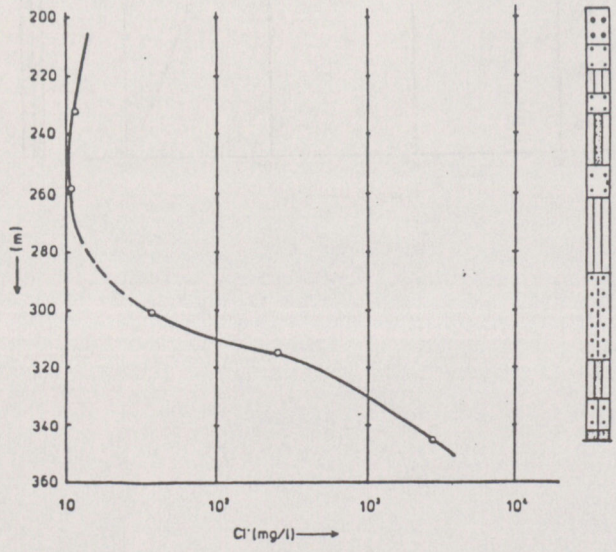
A

21C-113



B

26D-42



C

25A-943

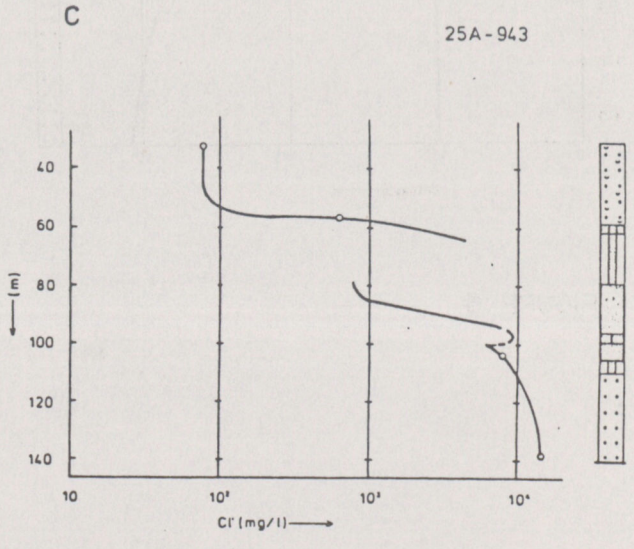


FIGURE 7

SALINE CONTAMINATION PHENOMENA IN THE KARSTIC AND FISSURED CARBONATIC AQUIFER OF THE SALENTINE PENINSULA (SOUTHERN ITALY) AND THEIR EVOLUTION

V. COTECCHIA* - T. TADOLINI* - L. TULIPANO*

ABSTRACT

Carbonatic mesozoic rocks represent a huge aquifer where fresh waters float on intruding land mass sea waters. As for other areas in the world, also in the Salentine Peninsula a continuous degradation of waters quality, due to the saline contamination processes, is on foot. For a long period, observations have been carried out on some observation wells drilled for scientific purposes and many other wells drilled for agricultural water supply. In this way it has been possible to underline the main features through which saline contamination takes place. In the aquifer such a contamination occurs also in relation with the high degree of anisotropy, being mostly vertical. Among the main modes, mention should be made of the one causing an upward expansion of the dispersion zone between fresh water and salt water. Such a phenomenon also occurs without significant piezometric levels changes. Considerations have been carried out about the influence exerted by sea water intrusion phenomenon and by feeding process on fresh groundwater temperature.

INTRODUCTION

Salentine peninsula is a region prevailingly karstic in type, almost lacking in surface water resources. Then underground waters account for great importance, since they are the only local water resources avaiable. Their use very often is limited because of sea intrusion phenomena occurring in the territory, being the region of coastal nature. The geological basament, in which underground waters move, is made up by limestones, dolomitic limestones and dolomities belonging to the Cretaceous, generally very permeable by fissuring and by karst phenomena. The high degree of permeability of the rocks outcropping along most coastal zone causes the penomena to regulate the discharge of underground waters into the sea by means of a moltitude of subaereal or subaqueous springs being concentrated or spread, generally they discharge brackish waters, with a saline content seldom less than 3 g/l. Moreover the same high degree of permeability favours the inland sea intrusion, occurring everywhere at the bottom of fresh groundwater. Then it is the classical hydrogeological example of a coastal fissured and karstic aquifer /1/. Because of the development of agricultural and industrial activities in the last decades there has been an ever-increasing demand of water supply and, while waiting the realization of artificial basins and water conduction works from surrounding regions, such a demand has been satisfied by an ever increasing groundwater drawing. Maybe,

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as it happens everywere, such massive drowings have been carried out without knowing the real availability of the aquifer, considering the eventual saline contamination due to the variation of fresh water-salt water equilibrium. Nowadays a remarkable degradation of ground waters quality has been recorded in some large zones of the region so that waters have been prevented from being drawn for any use. An over ten years study, about the subject, has aimed at the realisation of the best investigation methodologies about sea water intrusion and fresh groundwater saline contamination and the check of the real trend of the phenomenon also in relation with groundwater drawing /2, 6/. In virtue of such scientific investigation programs, till now in the region some observation wells have been drilled allowing detailed data about the trend of transition zone between fresh waters and salt waters.

Hydrogeological notes

As we have mentioned before, the aquifer is made up by a cretaceous carbonate formation, generally very permeable by fissuring and by karst phenomena. For, such an aquifer is strongly anisotropic as for the special distribution of permeability features. Such anisotropy, expecially vertical, relates to the paleogeographic evolution peculiar to coastal regions. Underground water moving base level, as at present, also in the past has been represented by the sea horizon. It is well know that the average sea level, in the run of geological eras has undergone even remarkable changes as to the continent. Karst dissolution phenomenon took place where, from time to time, underground water moving happened, also considering the presence, in the Mesozoic formation, of levels more or less able to become karst in type according to their lithological nature. For such reasons such an aquifer is made up by a series of preferential water levels higly permeable by karst phenomenon and by rocky levels having a lower permeability only by fissuring /4/. The main hydrogeological features concerning the mentioned aquifer, generally show good permeability features. Almost everywhere the aquifer shows phreatic conditions, normally hydraulic heads are very low, seldom exceeding 3 m above m.s.l.; hydraulic slopes are also very low, about 0.25‰. Water table trend of ground water (fig. 1) shows how also in horizontal position the irregular distribution of the features of permeability determines the presence of preferential flow lines, and since they are situated in high permeable zones as far as the coast, often they coincide with salt water intrusion lines from the sea to inland. Since groundwater has low hydraulic heads, its thickness, above the transition zone which separates it from intruding sea water is not very high; in the innermost zones it reaches about 150 m. Underground waters, of the mentioned region, have been originated by the infiltration of a high rate of precipitation during the autumn-winter semester;

moreover an abundant underground waters contribution from the near Murgia's region has been remarked. This region represents a hydrogeological bordering unity through having uncertain limits, since they are originated by quite different hydrogeological and morphological conditions, and not by geolithological reasons.

Mechanism of groundwater saline contamination

By means of periodic measurements of saline and thermal stratification along the water column of observation wells, a lack of balance, more or less marked, between intruding sea water and fresh groundwater, has been recorded /5/. Such a phenomenon is evident in some areas subject to massive groundwater drawings. Particularly it has been observed that the relationship between hydraulic head and the level of the top and the bottom parts of transition zone are not in agreement with the vertical distribution of saline content and then with the real trend of water density with the depth /5/. According to the real trend of waters density for a given real hydraulic head measured in the observation well, intruding sea water should be at a level higher than the one at which generally it is found /6/. Near to the part characterized by a lack of balance, detectable both a lack of head and by the difference between the theorectic level and the real one of intruding sea water, one observed how transition zone presents the greatest thicknesses; this implies that for an aquifer, as the one in question, massive water drawings do not cause reductions of the piezometric head, but a progressive salting process from the transition zone to upwards, viz fresh groundwater thickness narrows causing an expansion of the brackish zone. Evidently this salting process takes place by a mobilisation of intruding sea water which surely turn out to be favoured in those areas characterized by a higher degree of permeability or, better, by a higher communicability with the sea. That is proved by the distribution of saline content in groundwaters of the whole region in question (fig. 2). The deep inland deflections shown by isohalines in some areas, surely coincide with the areas having a high degree of permeability as it can be deduced by the trend of groundwater counturs in the same zones (fig. 1). In other words the replacement of that part of sea water, salting the groundwater by an increasing of diffusion zone thickness, takes place more easily along the same lines of preferential discharge peculiar to the zones with a higher permeability.

Underground waters temperature as environmental tracer with reference to the sea water intrusion phenomenon

What we have been saying up to now shows the importance, in the study of fresh water-salt water equilibriums, of knowing with enough

precision the features of underground water moving, from the feeding process to the discharge of groundwaters into the sea. Besides all investigations depending on traditional and advanced methodologies, the use of underground waters temperature have been of great help. It is know that the distribution of underground waters temperature in a coastal region is the resultant of several factors among which, the average temperature of feeding waters, the residence time of underground waters and the thermal influence exerted by intruding sea waters on groundwaters, are the most important /3/. The average temperature of feeding waters of the region in question is about 12°C since the feeding phase occurs only in the autumn-winter semester /3/. Underground waters residence time depends on many factors, the permeability first of all. As for thermal influence exerted by intruding sea waters, it is worthy to point out that since they have a low degree of mobility, or even being motionless in the innermost parts of the peninsula, they reach the equilibrium with the geothermal gradient; then starting from temperatures of about 17,5°C peculiar to the top of intruding sea water surely there will be an increasing of temperature of the same waters with the depth. The figure 3 shows the distribution of grounwater temperature on a horizontal section drawn 10 m under mean sea level. Considering what has been above-mentioned, and particularly the relationship between the degree of underground waters mobility and their temperature, from that figure one can observe the effects of local feeding marked by temperatures of about 15-16°C. From those areas to the zones in which the discharge of groundwater takes place, the temperature will increase with horizontal gradients as lower as higher is the permeability of the aquifer. The area surrounded by 18°C isotherm near to Lecce's town, turns out to be particularly singular. The temperature relatively high of underground waters in such a zone, can be explained by watching the trend of vertical thermal gradients which are negative in the same zone, since the temperature decreases with the depth. This case is an example of the superposition of different water circulations. For, the upper part of the aquifer has copious underground water contribution coming from the near Murgia's region beneath such waters, those originated by the infiltration of meteoric waters, through the near areas, are found. This phenomenon will be better classified by means of isotopic analysis in progress, since in this particular zone, it is quite sure that there is an inversion of ages along the vertical of the aquifer. As for the information available from the distribution of temperature about intruding sea water influence on groundwaters, one can note how same zones remarkably concerned with saline contamination, are characterized by an increase of temperature of underground waters; this relationship is particularly evident along the Ionic coastal zone of the central part of the peninsula. In some cases one could assert that thermal influence caused by sea water intrusion happens

in advance as to the salting process of groundwaters.

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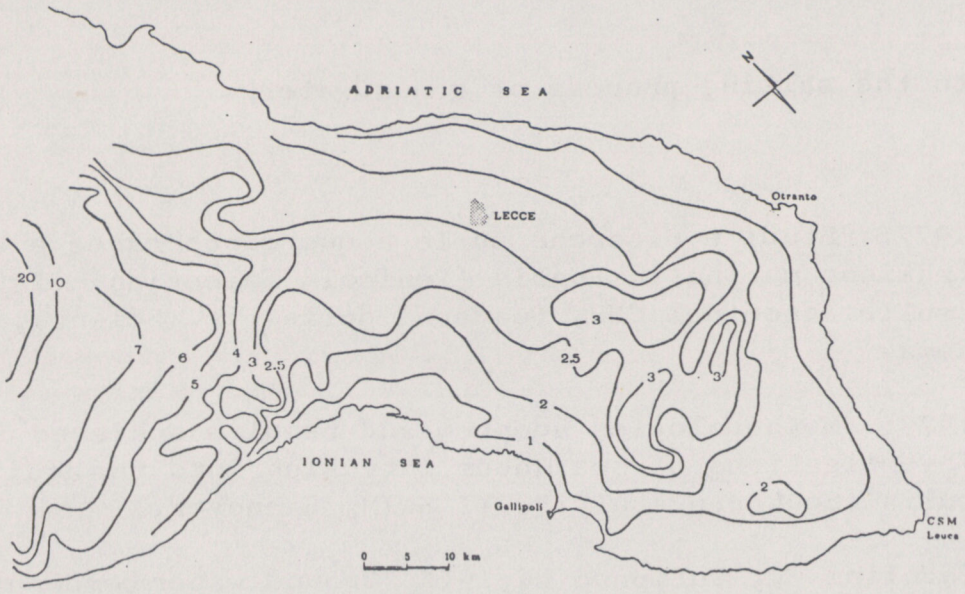

Fig. 1
Water table contours

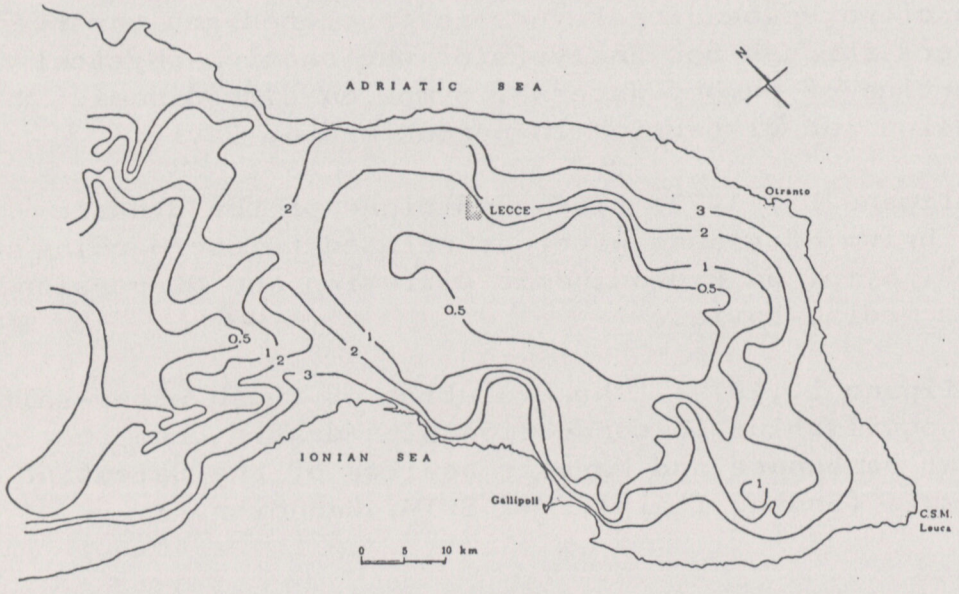

Fig. 2
Salt content distribution

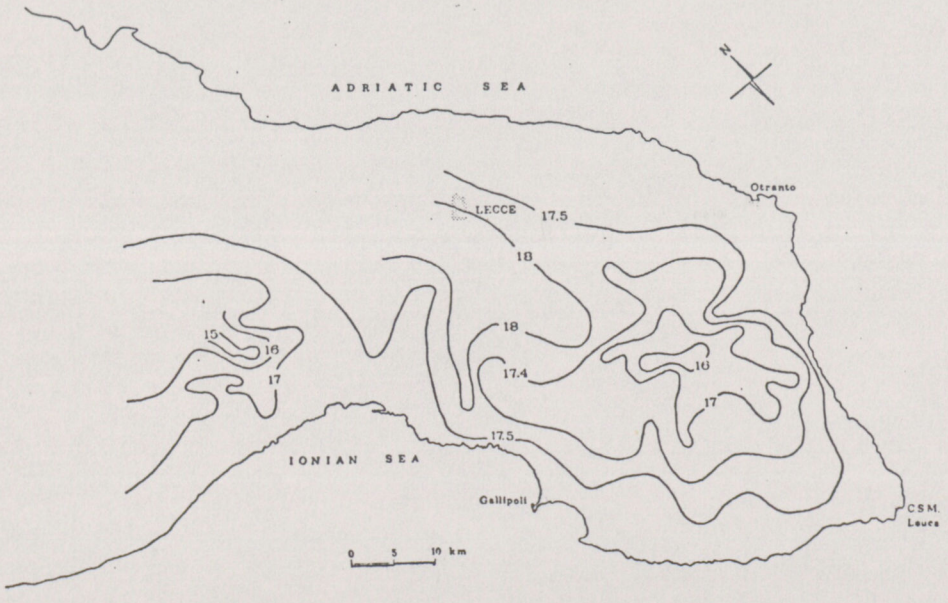

Fig. 3
Groundwater temperature measured 10 m below m.s.

SALT WATER INTRUSIONS INTO COASTAL AQUIFERS OF GDAŃSK REGION

By B. Kozerski[1]

ABSTRACT

In the coastal zone of Gdańsk region the ground water of great usability occurs in Cretaceous, Tertiary and Quaternary sediments. The water has been intensely exploited. Present yield amounts to 550000 m^3/d. As a result intrusion of sea water into the coastal aquifers has been observed. The problem has been investigated and a observation system is to be developed.

INTRODUCTION

There are a few aquifers rich in water along the coast of Gdańsk Bay. That water has been intensely exploited for tens of years; the present yield amounts to nearly 550 000 m^3/d. The ground water is the main source of supply for a number of towns of Gdańsk region of 1 million inhabitants. A deterioration in the quality of ground water has been observed for several years, particulary with regard to its salinity due to sea water penetration into the coastal aquifers. All greater ground water intakes are situated at a distance of 1 - 10 km from the coast, and numerous single wells are located in the vicinity of the sea.
The above mentioned fact and the constant increase of water abstraction as well as other factors have seriously complicated the water supply of Gdańsk and Gdynia.

DESCRIPTION OF THE AREA

In the coastal zone of Gdańsk Bay the ground water of great usability is present in the Cretaceous, Tertiary and Quaternary sediments. Under natural conditions, not disturbed by exploitation, the water flows down to the sea which is the regional drainage base for it.

The changes and trends being observed in this respect will be referred to on the basis of general hydrogeological conditions of the coast of Gdańsk Bay.

Cretaceous and Tertiary aquifers

Fine-grained glauconitic sands included in Coniacian and Santonian constitute the main aquifer of the Cretaceous strata. These strata are widely spread in the Gdańsk region forming a vast artesian basin. The aquifer appears regularly at the depth of 150 m below the sea level, the thickness being 80 - 100 m. The aquifers are covered by limestones and Campanian marls of a thickness of 50 m which are only locally pervious. Above them there are Tertiary and Quaternary sediments /Fig. 1/.

The original piezometric surface of the Cretaceous aquifer at the coast of Gdańsk Bay has been at 15 - 18 m above the sea level. The recharge area is the moraine hills of the Kashubian Lake District, the drainage area the coastal plains and the Bay of Gdańsk.
The mean yield of the wells equals to abt. 1000 m^3/d and transmissivity coefficient varies between 100 - 500 m^2/d.

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The water is of good quality. The content of chloride ions is below 15 ppm Cl^-, hardness i - 4 m-val/l, iron 0.3 - 0.6 ppm Fe.

As a result of eighty years exploitation which is still increasing, the piezometric surface has fallen in the area of about 400 km sq. In the centre it is 5 - 10 m below sea level. The zone of depression also covers the coastal part of Gdańsk Bay. The present piezometric surface at the coast is at the height of the sea level.

The natural flow of water has been reversed, hence the conditions have been created for leakage of sea water into the Cretaceous aquifer. Because of a low salinity of the Baltic Sea and an even lower one of Gdańsk Bay, and also because of a considerable thickness of the covering layers, no increase of water salinity in Cretaceous aquifers has noticed so far. In the course of time this phenomemon, however, may be of signigicance from a practical point of view.

There is one factor more that should be taken into consideration. In earlier Mezosoic formations of Gdańsk region there is fossil water of high salinity. The lowering of pressure in the Cretaceous aquifer faciliates an upward migration of older water thus forming an additional threat to fresh ground water.

The Tertiary in the coast of Gdańsk Bay is represented by Miocene and Oligocene strata. These strata are only locally composed of sands and do not form aquifers of regional extent. The total ground water discharge does not exceed 20 000 m^3/d. A dispersed and insignificant exploitation dimmishes the possibility of intrusion of sea water into the Tertiary sediments. One cannot exclude, however, an increased salinity indirectly from higher situated Quaternary sediments.

Quaternary aquifers

The occurence of ground water in the Quaternary sediments and its contact with sea water is differentiated and depends mainly on the geomorphological conditions. The following formation may be mentioned here as water bearing structures: /Fig 2/

- the Vistula Delta,
- marine terrace,
- ice marginal valley,
- bay barrs,
- moraine hills

Below is the hydrogeological characteristics of the Quaternary sediments of Gdańsk Bay presented in the sequence as above.

The Vistula Delta: encloses the Bay from the south over the length of more than 30 km. Along the coast line, on alluvial deposits the sands of marine and eolic accumulation form narrow and specific forms of bay-bars.

Among the Quaternary sediments of the Delta region there are aquifers of Pleistocene and Holocene. The sandy strata of Pleistocene are 20 to 50 m in thickness. Beneath them there are clays and loams and they are covered with muds, peats and Holoceneous sands. Alluvial sands are

often join to water bearing Pleistocene sediments of the delta area forming one quaternary ground water horizon having at certain places the thickness of 40 - 60 m /Fig. 3/.

In the area of the Vistula Delta the salinity of ground water has been observed first. It is typical for the area situated below the sea level. These areas have been, for centuries, intensively drained. In some parts of the Vistula Delta the salinity of Quaternary water reaches 2000 ppm Cl^-/l. It may be assumed, however, that we are dealing here with young fossil water formed during the Delta evolution and which has been preserved due to a slow exchange of water. An important part in the salinity of aquifers of the Delta play also the water of natural and artificial dykes and particulary the Dead Vistula River. This water is salty as a rule and penetrates quite easily into the wells situated nearby where the chloride content exceeds 1000 ppm. The increase of chloride ion content in Pleistocene sediments of the Delta area is at present observed at a distance of about 5 km from the Dead Vistula River.

The Marine Terrace covers a flat area between Gdańsk and Sopot. From the surface down to the depth of 30 - 40 meters there are highly pervious sands and gravels of Pleistocene age extending down under the bottom of Gdańsk Bay. Below, there are boulder clays or low permeable Tertiary sediments /Fig. 4/. The marine terrace constitutes an abundant reservoir of ground water being recharged partly from the moraine hills and partly directly by precipitations. The coast of the Bay of Gdańsk constitutes the zone of salt and fresh water contact. The ground water of the marine terrace is intensively exploited and the wells of one of the intakes are spaced at a distance of about 1 km from the shore. The rate of discharge of a single well amounts to approximately 1000 - 1500 m^3/d. Lately the conditions of exploitation of this intake has become considerably worse, the progressive urbanisation has significantly decreased the direct recharge. The necessity of a constant discharge of about 24 000 m^3/d has caused the increase of depression in the wells down to 6 - 10 m. Consequently the penetration of sea water has started and the front of salt water proceeds towards the wells. The intensity of this process is hampered by the bottom sediments of the Bay of lower permeability and by fine - grained sands of sea origin covering the water bearing layer /Fig. 4b/.

At present the main problem to be explained is the rate of proceeding salinity of water in connection with the decreased lateral recharge from the land side. For this purpose a piezometric net has been installed and observations have started. Random observations show the content of Cl^- being 80 mg/l at a distance of 150 m from the sea shore.

The aquifer in Reda ice marginal valley is partially a confined and partially an unconfined one.

This stratum is exploited with the rate of discharge of 180 000 m^3/d. In some parts of the valley a constant lowering of the ground water level can be observed. In the estuary part of the valley there are locally two aquifers. The first one is composed of river and sea sands and includes polluted subsurface water thus not being exploited.

The second aquifer which is lower in the estuary part of the valley is divided from the first by a layer of clays of the thickness of 20 m. Below it there is a confined aquifer consisting of sands and gravels of a thickness of up to 30 m.
Both the covering layer and the aquifer spread under the coastal basins of Gdańsk Bay. The recharge of the aquifer takes place by direct infiltration and lateral inflow from the moraine hills of the Kashubian Lake District. The ground water is discharged at the outcrops at the sea bottom. So there is a natural zone of contact with sea water in this place. It occures within a distance of a few or more kilometers from the shore. A long lasting exploitation of this water from the intakes situated at the shore has led to the lowering of the piezometric surface below the sea level. The ground water under the bottom has been directed towards the intakes /Fig 5/.

There is no possibility, however, to analyse this phenomenon on the basis of observation because the distance between the outcrop of the aquifer and the shore is not known. The gradual increase of salinity is shown directly in wells in the vicinity of the Bay. Any assumptions in this respect are only approximations and must only be based on uncertain data obtained by indirect methods.

Sea bars constitute a part of the coastal area of Gdańsk Bay. The Hel peninsula of about 30 km length encloses the Bay from the north and the Vistula Bay Bar from the south along 40 km separating the Bay water from the areas of Vistula Delta.

As concerns geological structure of the Bay bar the predominating sediments are sand of a thickness between 50 - 60 m. The depth of the water table varies between 1 - 10 m depending on the height of dunes. The level of the ground water table in relation to sea level is higher; 1 to 2 m. Under natural conditions fresh water occurs at the floor of sea sediments. It is because of the low salinity i.e. 4 - 8 $^{o}/_{oo}$ Gdańsk Bay and Baltic respectively.

The conditions of balance between sea and fresh water have been disturbed in places of increased exploitation of ground water. Jurata on the Hel peninsula may serve as an example. The consumption of ground water is considerably increasing in summer and its salinity as well.

In the present paper the problem of sea and ground water contact within moraine hills has been omitted, because it is of no practical importance.

CONCLUSIONS

An intense exploitation of ground water in Gdańsk region has led to disturbance of the natural balance between the sea and ground water. There have been observed negative effects due to intrusion of sea water to aquifers of the Vistula Delta, marine terrace and sea barres. The ground water of Reda ice marginal valley is also threatened.

The proceeding process of intrusion has become the subject of investigations. A special system of observation is being developed in order to obtain data on the ratio and extent of intrusion.

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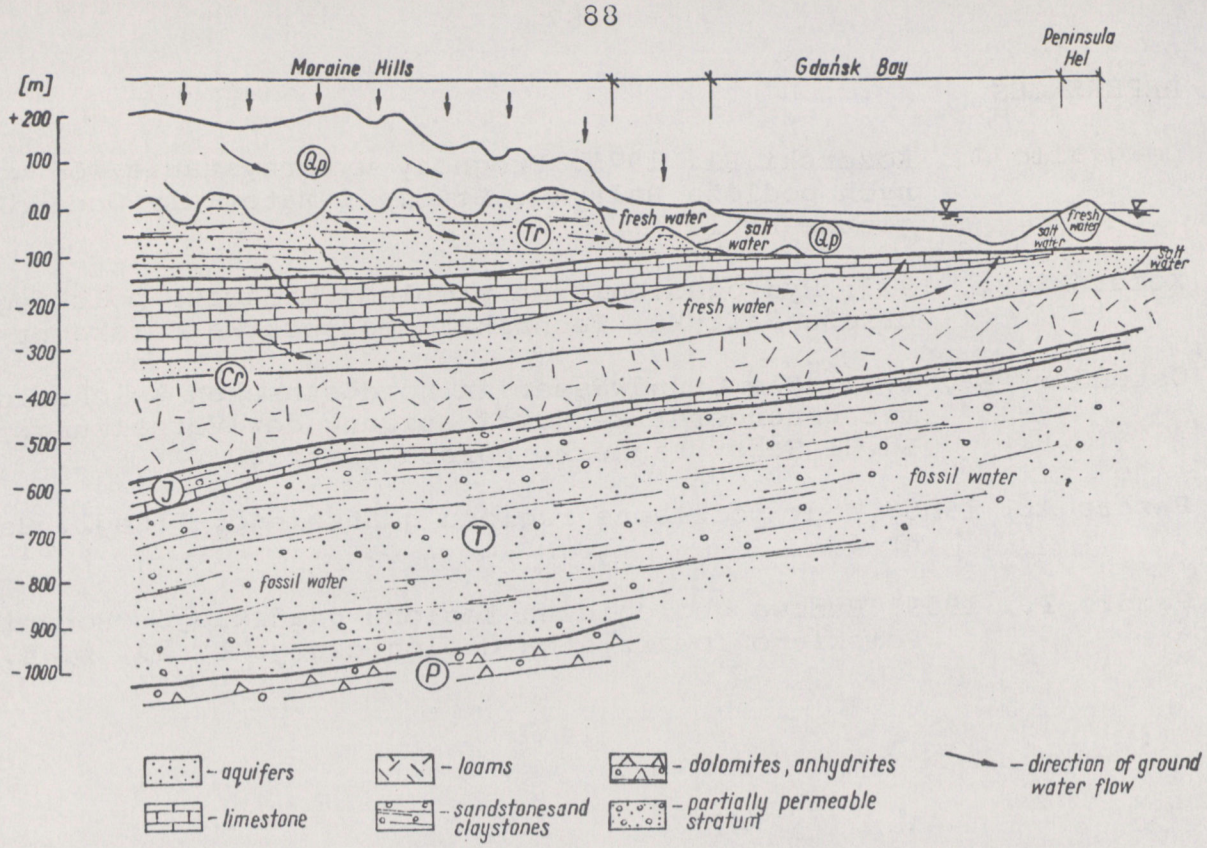

Fig 1. General geological cross-section of Gdańsk Region

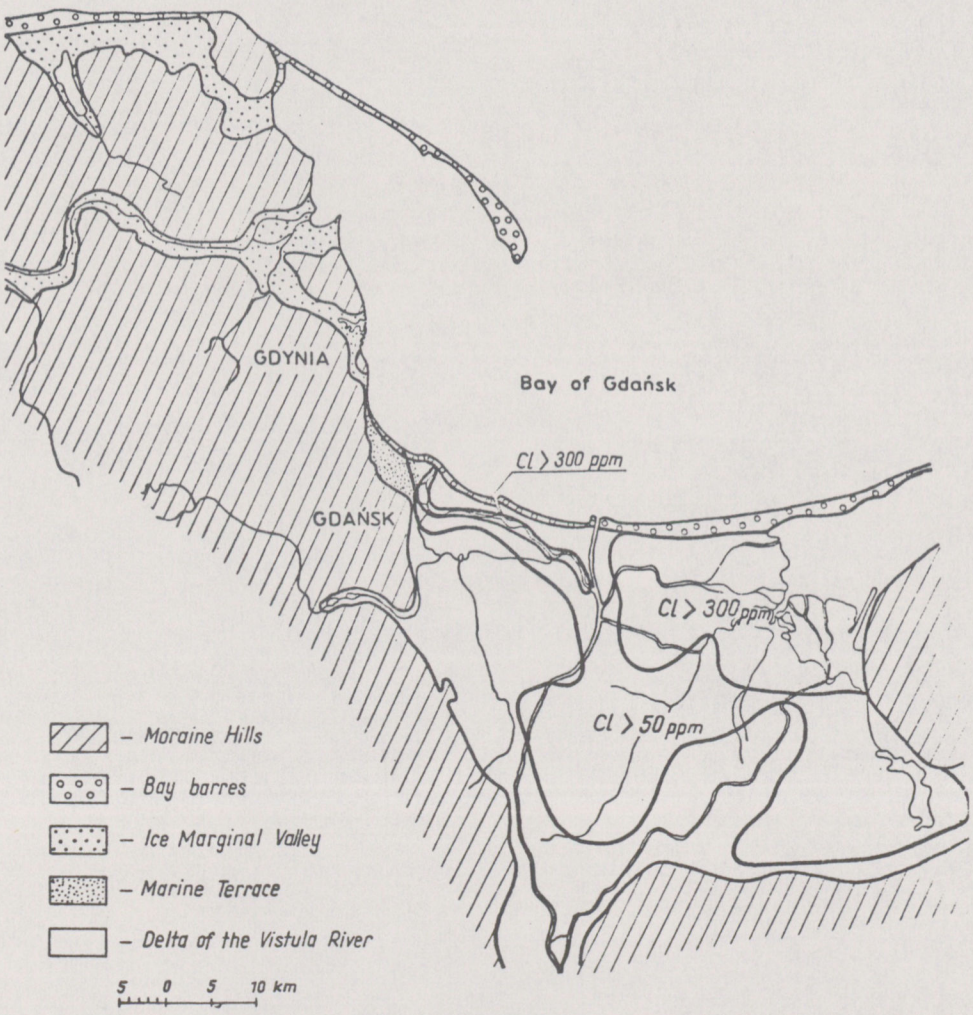

Fig. 2. Geomorphological structures

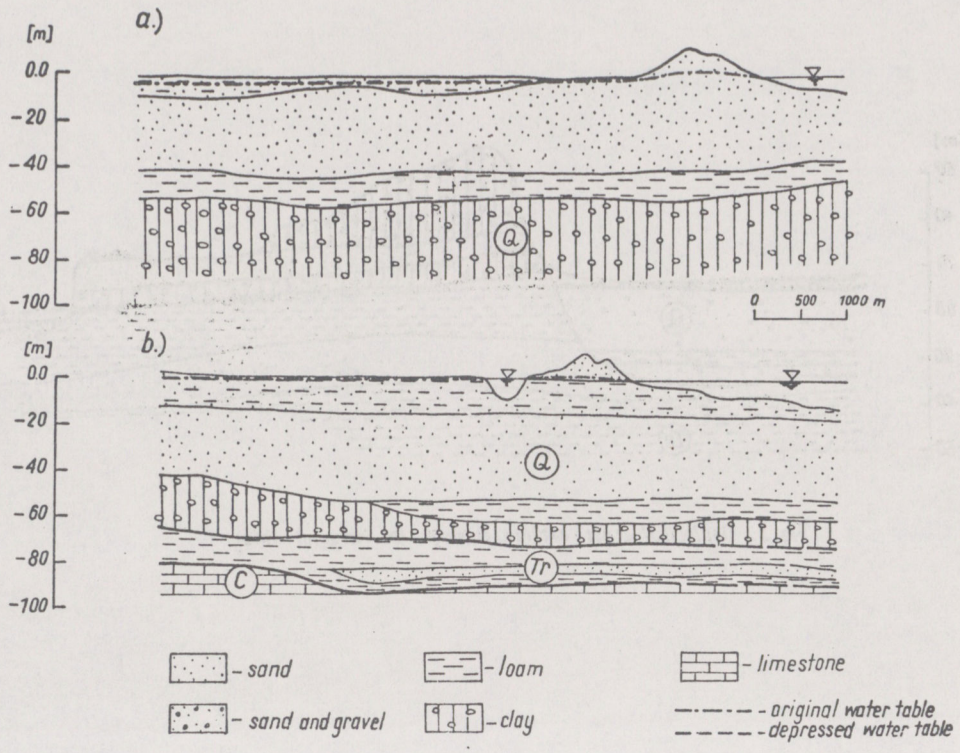

Fig. 3. Geological cross-section of the Vistula Delta

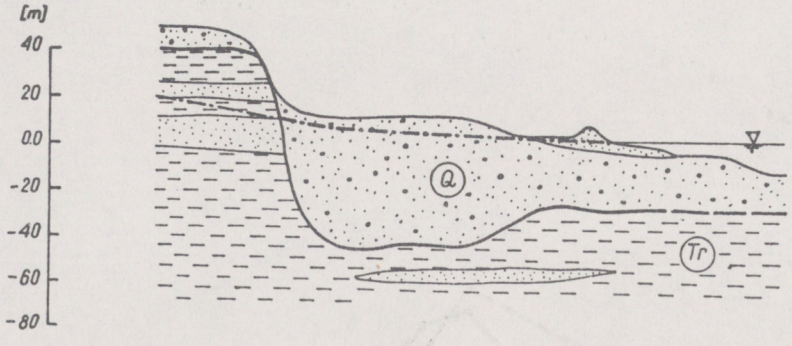

Fig. 4 a. Geological cross-section of the Marine Terrace

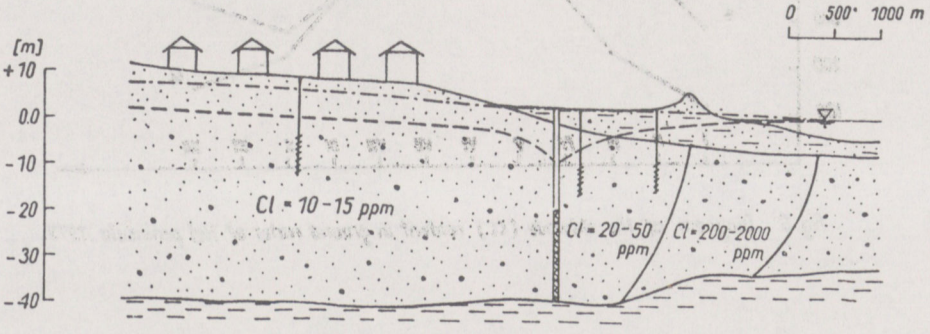

Fig. 4 b. Cross-section of ground water intake area

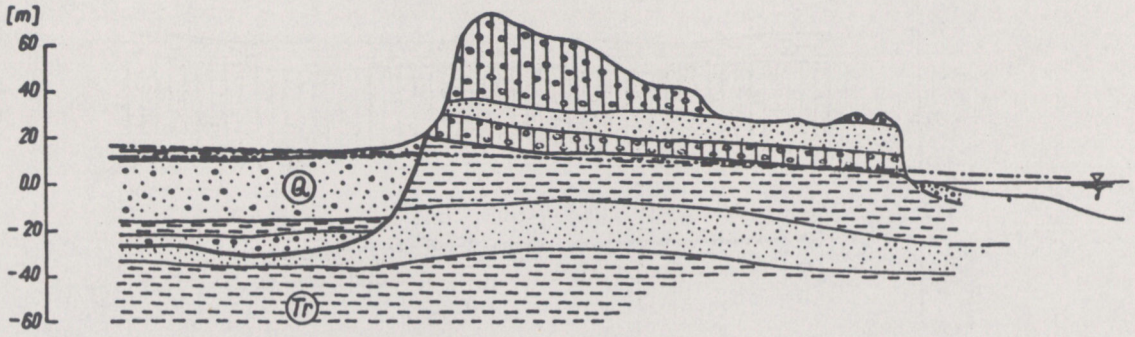

Fig. 5. Geological cross-section through Ice Marginal Valley

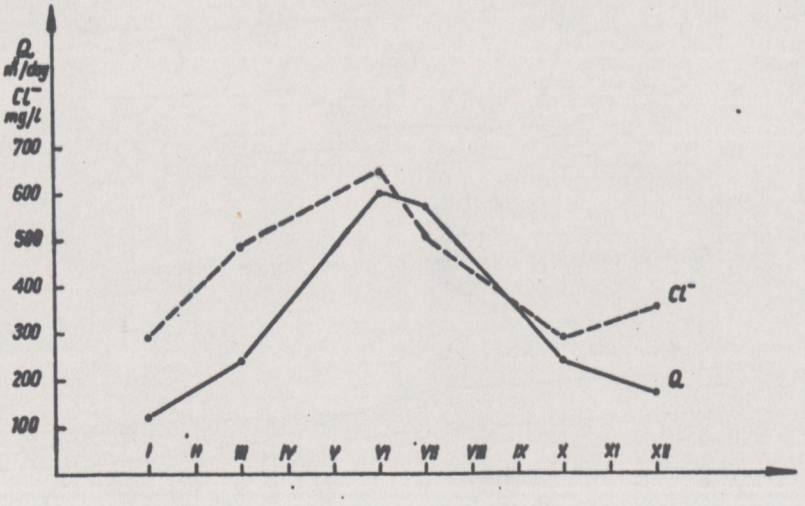

Fig. 6 Diagram of the chloride (Cl^-) content in ground water of Hel peninsula .1975.

TESTING VLF-RESISTIVITY MEASUREMENTS IN ORDER TO LOCATE SALINE GROUNDWATER

By C.-F. Müllern[1] and L. Eriksson[1]

ABSTRACT

VLF-resistivity measurements are being tested as a rather cheap and rapid tool for making rough estimations of the occurrence of saline groundwater. In this study the method has been tested in order to: 1. locate fossil saline groundwater in a Pleistocene esker (sand and gravel) aquifer, 2. locate fossil saline groundwater in the Precambrian crystalline basement, 3. locate the saline-water/fresh-water boundary on the coast of the Bothnian Sea, 4. outline the boundary of pollution from a storage of salt used for de-iceing and anti-dusting on the roads. The results of these tests are presented and discussed.

Fig. 1. Geographical positions of the sites where VLF-resistivity measurements of this study were carried out.

INTRODUCTION

During the last few years VLF-measurements have been used to locate groundwater bearing fracture zones. This method has in many cases been successful and many geological consultants, and others, in Sweden have acquired this kind of equipment. Since it also is possible to make at least approximate resistivity measurements with this kind of equipment, preferably in single- and two-layer cases, it was thought worth-while to test it in connection with problems concerning saline

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groundwater. Under favorable geological conditions, the measurements are not very complicated, quickly made and inexpensive. It must be pointed out, however, that the results of this study are the first obtained in Sweden in order to locate saline groundwater, and some of the results are not yet properly verified. The different sites where the measurements of this study were carried out are shown in Fig. 1.

THE VLF-RESISTIVITY TECHNIQUE

The VLF-instrument including the attached resistivity device is mainly a radio receiver measuring the ratio and phase angle between the horizontal electric and magnetic fields of very low frequency radiowaves (15 - 20 kHz) transmitted from distant stations. The instrument is calibrated to read the resistivity directly. If the earth is of constant electrical conductivity down to the depth of penetration the phase angle is 45°. If in a two-layer case the top layer has lower resistivity than the bottom layer, the phase angle is less than 45°. If the situation is the opposite, the phase angle is more than 45°. The penetration depth depends on the measured resistivity and can be determined from a graph, supplied with the instrument. From other graphs the thickness of the upper layer, the resistivity of that layer and of the lower layer can be determined. One of these three parameters must however be known (or almost correctly inferred).

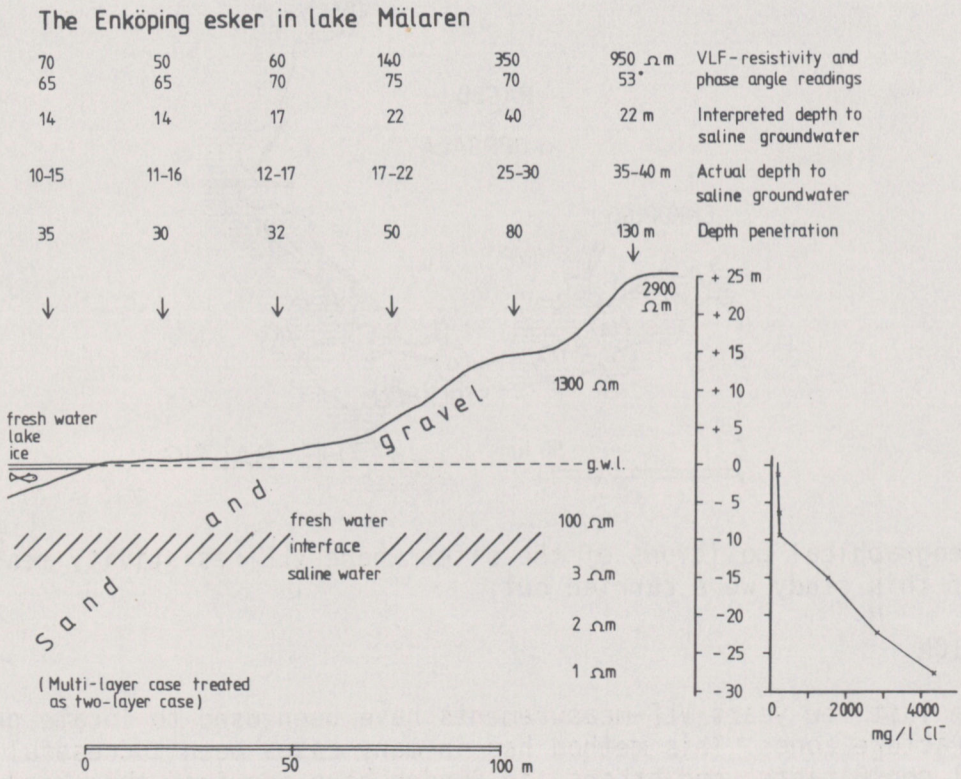

Fig. 2. VLF-resistivity measurements were carried out in order to determine the depth to the interface between fresh and saline groundwater on an island constituting a part of the Enköping esker in lake Mälaren.

FOSSIL SALINE GROUNDWATER IN A PLEISTOCENE ESKER

During the uplift of land after the latest glaciation, saline sea water has been trapped in different aquifers under certain geological and morphological conditions.
About 70 km west of Stockholm a row of small islands in lake Mälaren constitutes a part of the Enköping esker. On one of the islands (Tallholmen) drilling for groundwater was undertaken by a Swedish consulting company (AIB). It was found out that underneath a fresh groundwater layer, about 10 m thick, there is saline groundwater. This is shown in Fig. 2.
The resistivity values of the sand and gravel above the groundwater table were obtained from Beam Slingram readings[1]. The resistivity values below the groundwater table have been calculated from the chemical analyses of the groundwater. In the left part of Fig. 2 the geological setting constitutes a rather clear-cut two-layer case, with fresh groundwater on top of saline down to penetration depth. In this part the high phase angles indicate low resistivities below high resistivities, and the interpreted depths to the saline groundwater are in rather good agreement with the actual depths. In the right part of Fig. 2 the geological setting constitutes a multi-layer case. Down to penetration depth the geological sequence is, from top: 1. dry sand and gravel, 2. sand and gravel with fresh groundwater, 3. ditto with saline groundwater, 4. Precambrian crystalline bedrock, probably with saline groundwater (not shown in Fig. 2). It can be seen that the more pronounced the multi-layer case, the more difficult it is to make correct interpretations from the measured data. (At the right-most part of Fig. 2, the interpreted depth to the saline groundwater more likely is the depth to the fresh groundwater table.)

FOSSIL SALINE GROUNDWATER IN THE PRECAMBRIAN CRYSTALLINE BASEMENT

At Östervåla, about 100 km north-north-west of Stockholm and about 40 km inland from the coast of the Bothnian Sea, two wells were drilled into the Precambrian basement for the supply of water to the municipality. One well 79 m deep yields fresh water and the other 69 m deep yields saline water. The wells are 1 km apart and are drilled in two different, almost parallel fracture zones. The geological setting at the two well sites is almost equal, see Fig. 3.

Calculations made from the resistivity readings at A show that the resistivity ρ_2 of the second layer, the granite with fresh groundwater, is 30 000 ohm m. At A' the resistivity of the granite with saline groundwater is only 1 000 ohm m. Due to this large contrast in resistivity it probably is possible roughly to outline the borders of the saline groundwater body.

In this context it can be mentioned that the total output from the two wells over a ten year period is about 1 000 000 m^3 of saline groundwater and 1 500 000 m^3 of fresh groundwater, respectively. The chloride concentration has been almost constant in the two wells over this period of time. This indicates that the body of the trapped saline groundwater is rather large.

THE SALINE - FRESH WATER BOUNDARY ON THE COAST OF THE BOTHNIAN SEA

In order to test the VLF-resistivity method in connection with salt water intrusion, a profile was measured from the shore of the Bothnian Sea to 20 km inland. This was done in the vicinity of Forsmark about 120 km north of Stockholm. The results of these measurements are shown in Fig. 4.

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1 A moving electromagnetic source receiver instrument (9.8 kHz) calibrated for direct resistivity reading.

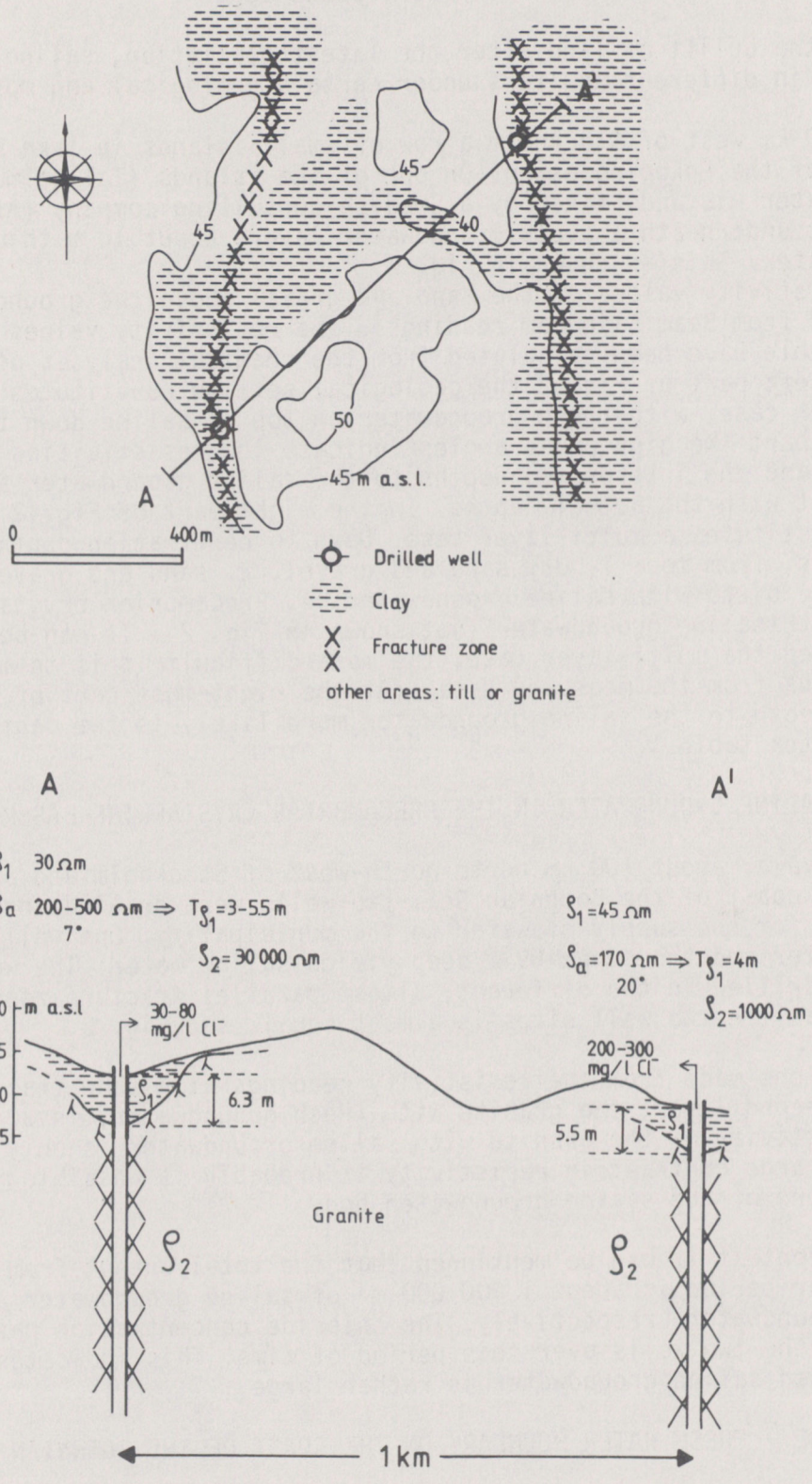

Fig. 3. Two wells supplying drinking water to the municipality of Östervåla. One well with low salinity, another with high. The resistivity of the granite aquifer was in the first case found to be normally high, in the second case definitely lower than normal.

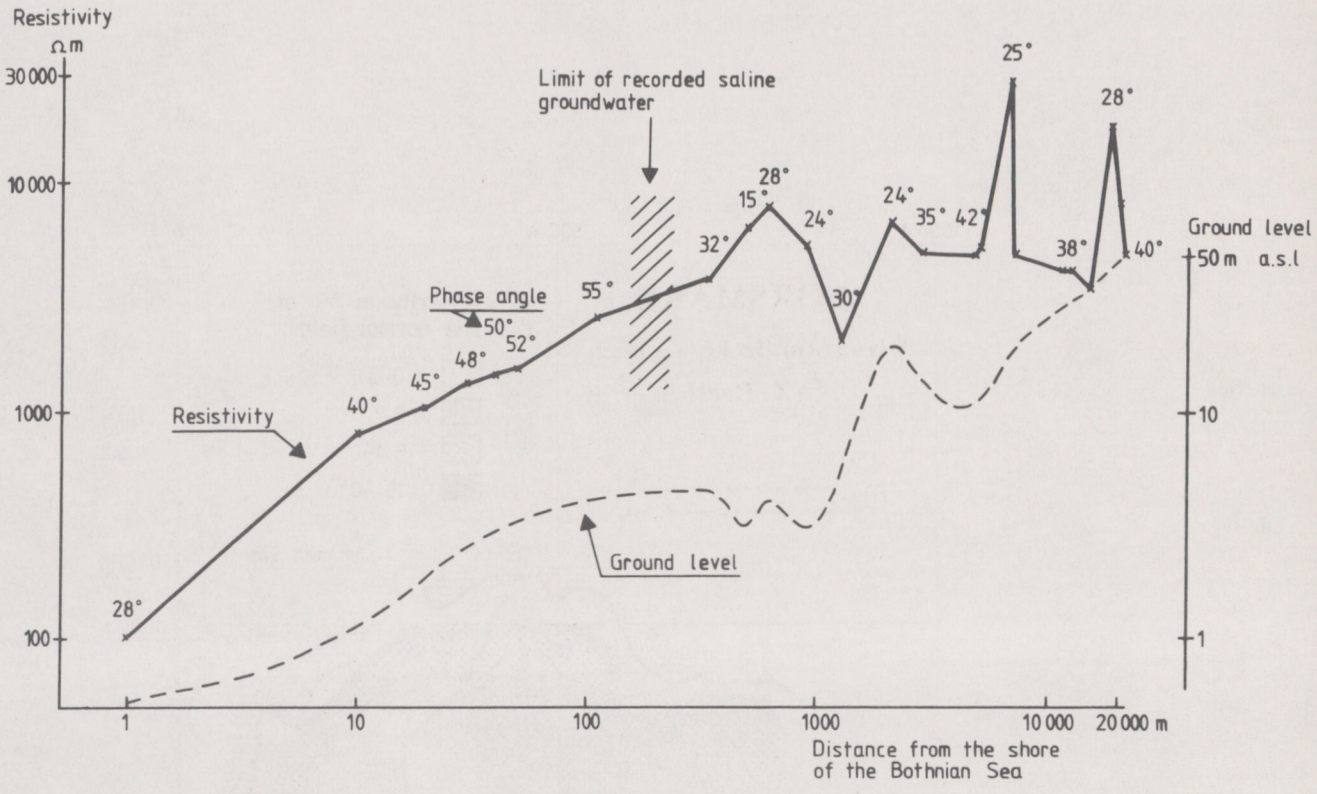

Fig. 4. A VLF-resistivity profile was measured from the shore of the Bothnian Sea and 20 km inland. Both resistivity and phase angle readings indicate saline groundwater a few hundred metres inland.

The geological setting is the same in all the measured points: 1 - 3 m of till on top of the Precambrian basement consisting of granites and metamorphic volcanics. In Fig. 4 it can be seen that the resistivity rises from 100 ohm m close to the shore to roughly 5 000 ohm m a few hundred metres inland (except two peaks and one low). The high phase angles (more than 45^{o}) close to the shore indicate the occurrence of saline groundwater within the depth of penetration. Somewhere around 200 m inland there is a sudden change in phase angle, where the values turn from high to low. This means that further inland there is no saline groundwater within the depth of penetration along this profile, and that contact with saline groundwater is lost at about 175 m depth. Obviously this does not exclude saline groundwater at greater depth.

Fig. 5 shows a map of Slingram (horizontal-loop electromagnetic survey, 18 kHz) measurements, from the area corresponding to the first 1 000 m in Fig. 4. On this map the influence of the sea water can be seen to reach generally between 100 and 150 m inland except in certain north-south and northwest-southeast trending zones, probably related to minor fracture zones. The penetration depth of the Slingram instrument is in this case about 80 m.

Measurements with the Beam Slingram instrument recorded influence of the sea water between 25 and 50 m from the shore. The penetration depth of this instrument is usually about 6 m (perhaps in some cases about 10 m).

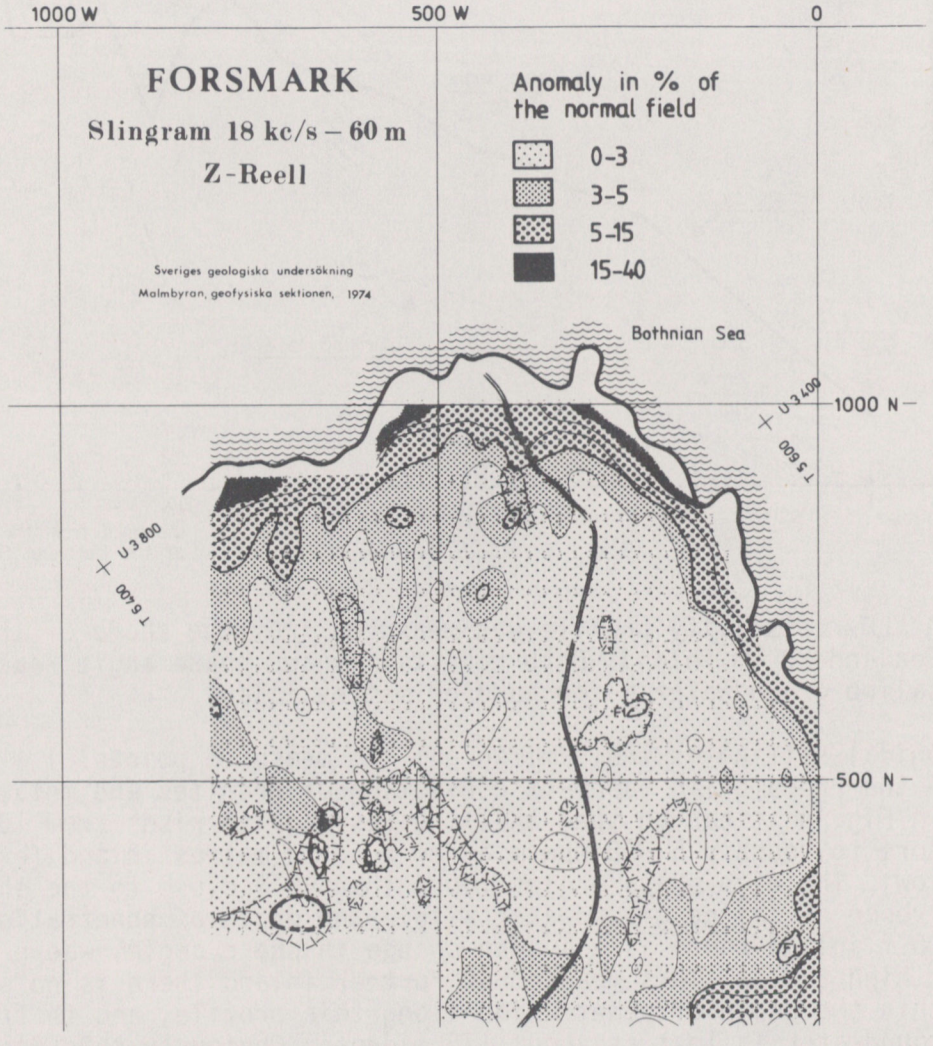

Fig. 5. Slingram measurements have earlier been carried out in this area, corresponding to the first 1 000 m in Fig. 4. These measurements recorded saline groundwater generally 100 - 150 m inland except in certain north--south and northwest-southeast trending zones, probably related to fractures.

Using these data the slope of the interface between saline and fresh groundwater can roughly be calculated to be something like what is shown in Fig. 6. However, the interface can be expected to be rather irregular due to the nature of the different fractures in this kind of aquifer.

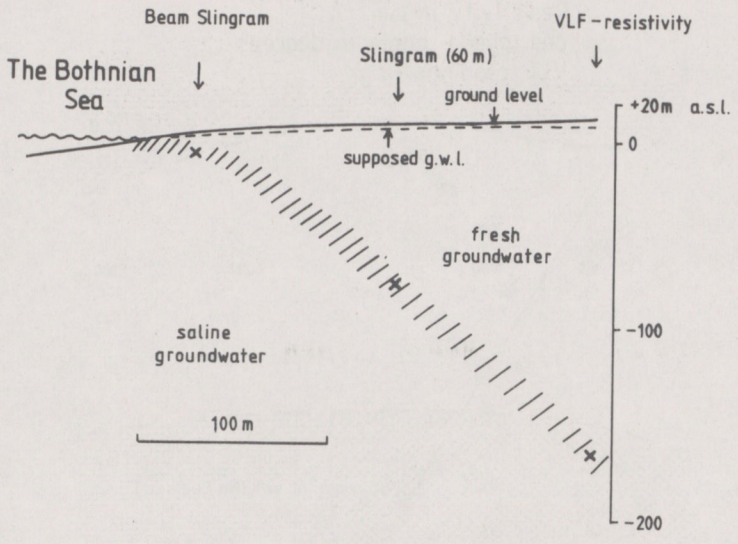

Fig. 6. A rough estimate of the slope of the fresh water/saline water interface, interpreted from VLF-resistivity, Slingram and Beam Slingram readings.

POLLUTION OF GROUNDWATER FROM A STORAGE OF SALT

At Rasbo, about 70 km north of Stockholm, salt (NaCl and $CaCl_2$) used for de--iceing and anti-dusting on the roads has been stored for a few decades. The geological setting is shown in Fig. 7a and b.

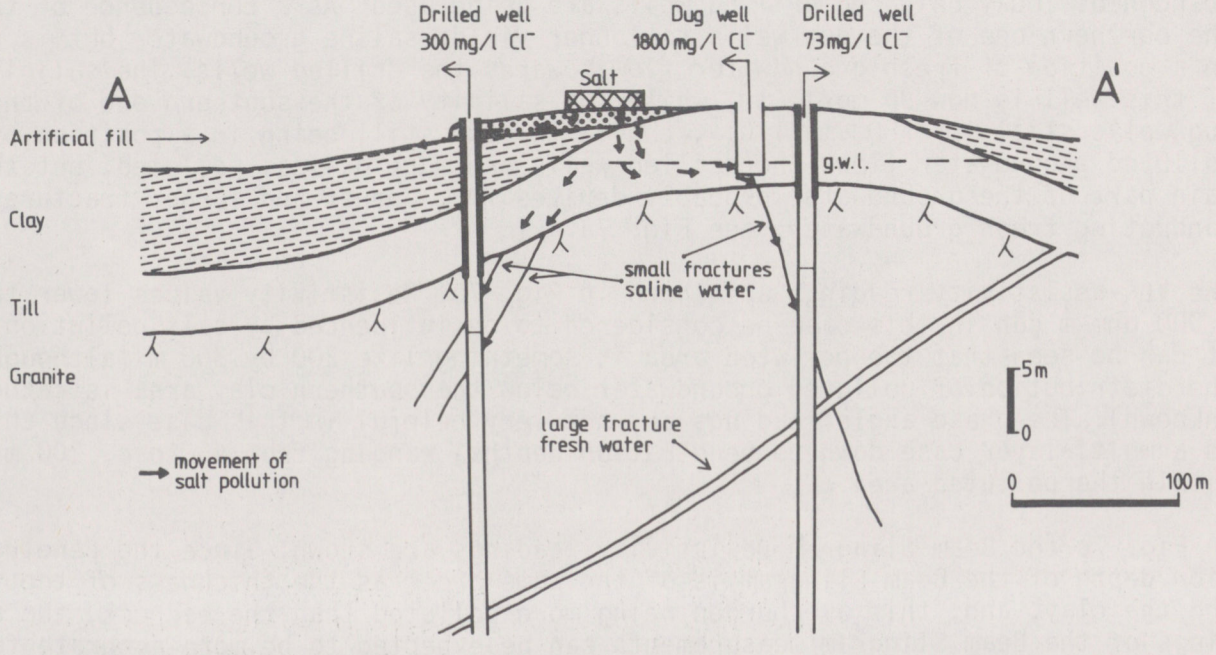

Fig. 7a. Section showing the geology and position of wells at a storage of salt at Rasbo. The salt has polluted the groundwater.

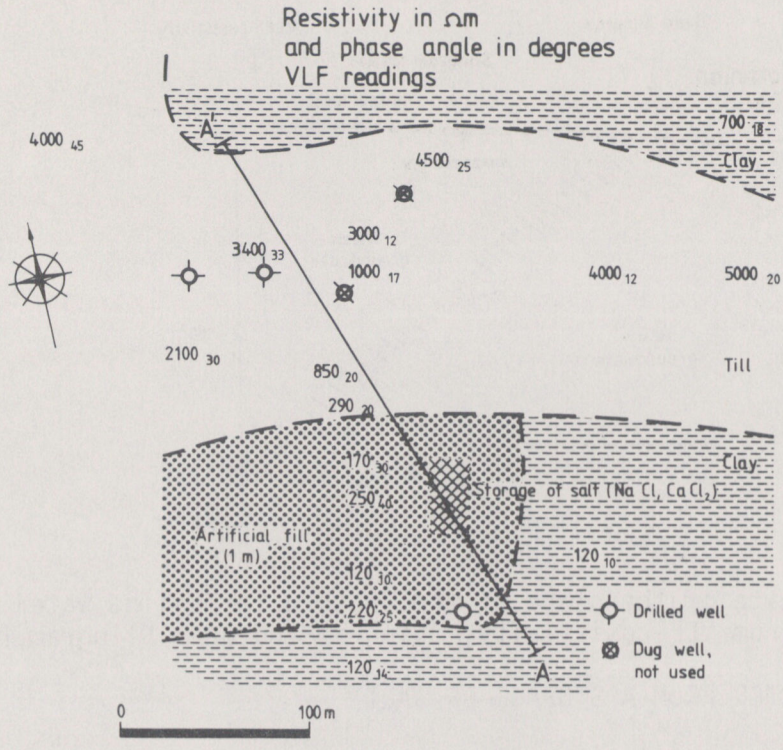

Fig. 7b. VLF-resistivity and phase angle readings at the storage of salt at Rasbo.

Concerning the pollution of the wells, the two dug wells had a very high content of cloride (between 1 800 and 1 900 mg/l Cl^-) during some time before they were abandoned. Today only the drilled wells are being used. As a consequence of this, the northern one of the dug wells no longer yields saline groundwater but is now in a position of fresh groundwater flow towards the drilled wells. The salinity of this well is now 96 mg/l Cl^-, while the salinity of the southern one of the dug wells still is 1 800 mg/l Cl^-, the latter one still being in a position of polluted groundwater flow. The drilled wells are more or less polluted, but the main part of the groundwater probably derives from deeper and larger fractures conducting fresh groundwater, see Fig. 7a.

The VLF-resistivity readings are shown in Fig. 7b. Resistivity values lower than 4 000 ohm m can in this case be considered to be influenced by salt pollution. It can be seen that the polluted area is something like 200 by 300 m (although the distribution of polluted groundwater below the southern clay area is rather unknown). The phase angle readings are not very helpful in this case since this is a multi-layer case down to penetration depths, ranging from 45 to c. 200 m within the polluted area.

In Fig. 7c the Beam Slingram resistivity readings are shown. Since the penetration depth of the Beam Slingram is of the same order as the thickness of the till and the clay, and, this overburden being more polluted than the bedrock, the readings of the Beam Slingram measurements can be expected to be more accurate for outlining the borders of the salt pollution. Resistivities lower than 800 ohm m measured with this instrument in the till area can be considered to be influenced by salt pollution. Comparing Fig. 7b and c one can see that the polluted area shown by the Beam Slingram measurements is somewhat larger than what is shown by VLF-resistivity measurements.

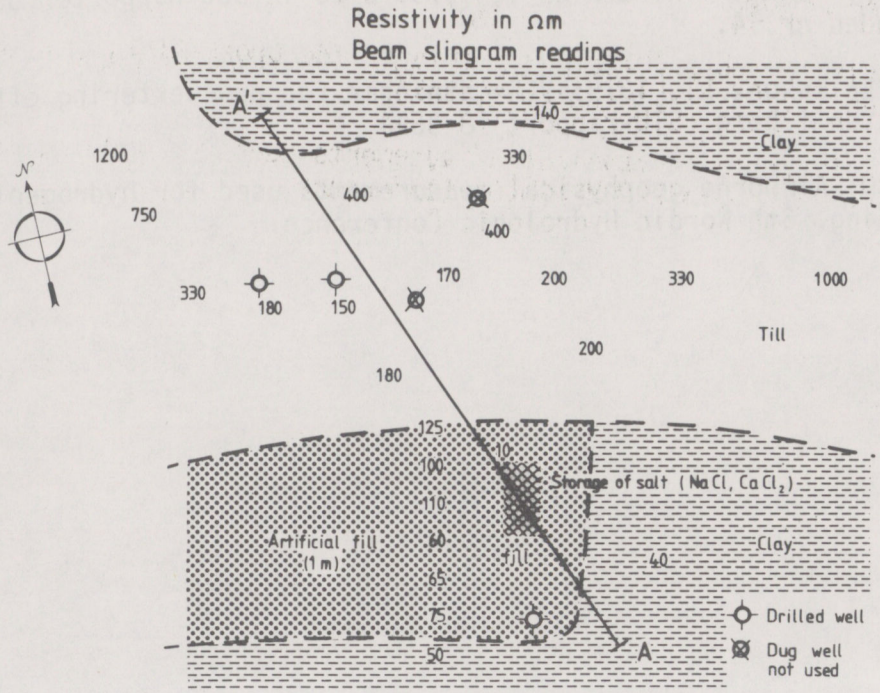

Fig. 7c. Beam Slingram readings at the storage of salt at Rasbo.

In order to protect the wells, salt is not going to be stored at this place any more.

CONCLUSIONS

This study has shown that when the VLF-resistivity technique is used it is necessary to have some knowledge of the geology including groundwater conditions in the measured area down to penetration depth, in order to make correct interpretations of the readings. Supporting information can rather easily be gained by other electromagnetic measurements in the same frequency interval (10 - 20 kHz). The obtained depth and resistivity values are often not completely exact due to the geological conditions not being as uniform and simple as one often tends to think they are. But mostly they are good enough for approximate estimations.

The VLF-resistivity measurements are more useful in pure one- or two-layer cases, but important information on the apparent resistivity in multi-layer cases can often be obtained.

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HYDROCHEMICAL OBSERVATIONS IN THE SUBSOIL OF THE WESTERN PART OF THE NETHERLANDS

by A.B. Pomper[1]

ABSTRACT

The presence of saline waters in the subsoil, consisting of a thick layer of marine and fluviatile sediments, of the western Netherlands indirectly has a great influence on the economic activities of the area. Therefore it has been given much attention of hydrologists and geologists.

It seems that there is no direct relation between the location of marine and fluviatile sediments on the one hand and the presence of saline respectively fresh waters in the subsoil. This paper gives a description of the location of the waters with different salinities and of the establishment of a parameter for their residence time, i.e. by comparing their sulfate contents.

INTRODUCTION

Most papers earlier presented by the author were based on observations in the region between the towns of Amsterdam and Rotterdam. It concerns a low lying area with a flow of saline groundwater from the seaward direction as well as a flow of fresh groundwater from the inland direction. This resulted in a complicated hydrochemical groundwater situation, as described in these earlier papers (Pomper, 1972, 1977). In the region North of Amsterdam, on three sides bordered by open water and recently investigated, a fundamentally different situation exists.

Hydrogeological and hydrochemical maps and profiles of the two regions have been combined and they are the subject of the present paper.

Although the explanation of a strongly schematized version of the groundwater situation along the lines of mathematical modeling appeared to be possible, here it will be tried to keep in contact with the nonschematized situation and to give a more detailed description of the situation.

GEOHYDROLOGICAL SITUATION AND GROUNDWATER FLOW SYSTEMS

It is well-known that the region discussed is part of the large West-European sedimentary basin. During the Tertiairy and the first part of Pleistocene, marine sediments in salt formation water were deposited. Then a period of predominantly fluviatile sedimentation started, locally with intercalations of marine interglacial sedimentation. Although many interglacials occurred (Zagwijn, 1975, 1978), marine sediments of only the last three interglacials, i.e. Holsteinian, Eemian and Holocene, have been found. It is possible that older marine sediments never did exist or that they have been removed by erosion. In that case we have luck to find remnants but up to that moment we must state the upper marine layers are the only ones present.

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The thickness of the mostly fluviatile Pleistocene layers overlying the marine lower Pleistocene increases from 80 meters near Rotterdam up to more than 250 meters in the North near the town of Hoorn. Unless the sediments originate from different river systems (northern Germany, Rhine, Meuse), the horizontal variations in grain size composition are not large.

Roughly the subsoil can be divided into:

a, a covering layer consisting of Holocene fine grained sediments;
b, the upper aquifer consisting of fluviatile, aeolean and marine sediments of different grain size composition and younger Pleistocene origin;
c, the semiconfining layer consisting of sediments from the marine Eemian (clay), Saalian glaciation (boulder clay and warven clay) and middle and lower Pleistocene fluviatile clay;
d, the lower aquifer consisting of coarse grained fluviatile sediments of different origin. In the northern half of the region this layer is the main aquifer;
e, the hydraulic base consisting of fine grained marine sediments of lower Pleistocene and Tertiairy origin.

The vertical resistance of the covering layer shows many variations. The upper aquifer has transmissibilities varying from 500 to 1000 $m^2.d^{-1}$. The lower aquifer mostly shows little variations in hydraulic conductivity but large variations in thickness. The thickest parts are to be found in the northern half of the region, where it functions as the main aquifer. The transmissibility of the lower aquifer varies from 3000 $m^2.d^{-1}$ in the South up to more than 10,000 $m^2.d^{-1}$ near Hoorn and somewhat lower values more to the North.

The direct consequence of the differences in transmissibility of the lower aquifer between the northern and the southern part of the region is that in the northern part the groundwater flow is mainly governed by the hydrological situation in the lower aquifer, whereas in the southern part the upper aquifer is the main influence on the situation near the ground surface. So in the northern half of the region the lower aquifer is the main aquifer and in the southern half there is no clear main aquifer.

Groundwater flow in the upper aquifer is mainly governed by differences in ground level. Except from deep polders, a large part of the region outside the dune area consists of polderland with ground levels near mean sea level. In the deep polders mainly upward seepage occurs whereas in the remaining higher part of the region at many sites infiltration is present. This results in a fundamental difference between the northern and the southern half of the region. In the southern half infiltration sites alternate with seepage sites, whereas in the northern half the intermittance of seepage and infiltration concerns large areas.

FRESH AND SALT WATER

Earlier (Pomper, 1972) a map was shown, giving the topography of the 500 ppm chloride plane in the southern half of the region. This map now has been extended to include the North (fig. 1). Unfortunately it was not possible to carry out this extension up to the northernmost town of Den Helder because of lack of data.

As was earlier noted in the different systems of groundwater flow, in this map also a fundamental difference between the northern and the southern half of the region can be observed. In the southern part of the region large differences in the topography of the 500 ppm Cl^--level are found. Fresh water bodies are

alternating with upwellings of saline water. The ground surface situation is closely related to the chemical groundwater situation.

In the northern half of the region the topography of the saline groundwater shows at first sight a less complicated situation. Along the coast large fresh water bodies are found. In this part of the region they are larger than in the southern part. Behind the dunes one large area with upwelling saline water is present from the Haarlemmermeer polder in the South up to Den Helder in the North, with only a small interruption in the neighbourhood of the town of Alkmaar. More to the East a large fresh water body is found, which will be discussed later in this paper.

The fresh water bodies along the coast need some explanation. These fresh water bodies originate from infiltration into the dunes of precipitation water. Aside from deep infiltration, the precipitation in the dunes is giving a subsurface runoff.

In the southern part of the dune area the fresh water bodies have a limited extension. Remarkable are the locally occurring extensions of the fresh water bodies in inland direction. In general a rather gradual boundary of the fresh water bodies exists in horizontal direction.

The fresh water body in the dune area in the northern part shows no natural interruptions. The only interruptions to be observed are due to human activities. There exist no inland extensions of the fresh water body and there is an abrupt boundary in the East. This fresh water body is strongly affected by the pumping for drinking water use, although its influence has been limited by infiltration into the dunes of river water obtained elsewhere during the last decennia.

The salt water outcrops West of the dunes are divided into cells by saddle shaped fresh water bodies along the rivers and larger canals. As already mentioned there are hardly any interruptions of this body in the northern half of the region.

Special attention is to be given to the fresh water body in the eastern part of North Holland. This fresh water body is the largest in the region examined. Near Hoorn the base of it is found at a depth of more than 90 meters below sea level. In contrast with the inland fresh water body in the southern part of the region, this body receives no fresh water supply. And yet it exists because of the fact that the area has been inundated by fresh water both during the Eemian and the Holocene transgressions. Although data were not available of the situation beneath the bottom of the nearby IJssellake, it seems that this fresh water body continues there. This is in agreement with Van Dam's (1962) geophysical observations.

A remarkable phenomenon is again the occurrence of inversions in the vertical saline - fresh water sequence. In earlier papers attention was also paid to them. Their occurrence in the southern half of the region is neglectable in comparison with that in North Holland. Especially along the fringes of the central fresh water body a very complicated situation exists. Figure 2 gives an illustration of this situation as compiled by Witt (1980). Due to the complexity of the situation, it is hardly possible to give a presentation in a plane. So Witt compiled maps with the chloride content at different levels and the map shown has been compiled by projecting the 200 ppm Cl^--lines of the different maps on one surface. Because of its limited readability, however, it needs some study.

Along the inland fresh water body in the southern part of the region, inversions were not observed. This body receives a fresh water supply by infiltration of precipitation water in the Utrecht hills, where outcrops of coarse grained Pleistocene sands occur.
During the past years much of my attention was spent on the question how this situation has come into existence. Applying radiocarbon datings would have been a useful procedure if the money to do this had been available. So other possibilities had to be looked for.

It appeared to be useful to make a study of the sulfate content of the groundwater. Sulfate is an important ion in sea salts and it forms a rather constant fraction of them. Under surface water conditions it is chemically inactive. Normal seawater contains about 2650 ppm sulfate, so after chloride it is its most important anion.

When sulfate comes under anaerobic conditions it will be affected by microbiological processes. So the solvable sulfate will be reduced to unsolvable sulphur hydrogen and pyrite. It is a rather slow process depending not only on the presence of microbes but also on the availability of organic matter and the redox potential. The data needed were available although not numerous; less than half of the water analyses did include sulfate.

Whereas the contents of chloride and sulfate in river and precipitation water can be neglected, the ratio of both contents in brackish water will be constant if no sulfate reduction has occurred. From figure 3 this can be observed. It concerns a water analysis of a deep boring near Den Helder (9D/186). The chemical composition is given in a modified Schoeller-diagram. Whereas Schoeller (1962) presents the different components in absolute values, this modified version gives them in relative percentages with regard to the composition of seawater. In this way differences of seawater mixtures are easily visible.

In the modified Schoeller-diagram of boring 9D/186 the connecting lines between the of chloride and sulfate are horizontal for water near the ground surface and at a depth of about 350 meters. But at a depth of 250 meters the relative percentage of chloride appears to be much higher than that of sulfate. The sulfate content has clearly been reduced by sulfate reduction, which indicates a longer residence time of brackish water at 250 meters.

In order to create a parameter for comparison of waters with different salt contents the contents of chloride and sulfate are expressed in their ratio. If the found chloride - sulfate ratio (from now on briefly called ratio) is equal to that of seawater recently infiltrated water is concerned and when it is considerably larger a long residence time since infiltration must have occurred. Unfortunately it is not (yet?) possible to make a absolute relation between ratio and residence time, as it proved to be impossible to get information from literature about rates of sulfate reduction.

The ratio of seawater amounts to 7.4 (data obtained from Westerveld, 1951). If the values are very much lower than 7.4 there is no or only a small amount of seawater in the mixture. Such extremely low values only are observed in fresh water bodies with very low chloride contents. Application of the ratio method requires a chloride content of at least 500 ppm, however, so for fresh waters radiocarbon dating must be applied.

The ratios found have been sketched into the profiles of the chloride contents found in the subsoil (figs. 4 and 5). The first profile (fig. 4) starts near the town of Maassluis in the South and goes up to Den Helder in the North and it

crosses the upcomings of saline water behind the dunes. The other profiles (fig. 5) are perpendicular to it. Aside from isochlorides, the values of the observed ratios are presented. The mentioned differences in the chloride situation in the northern and the southern part of the region are very clear. In the area South of Delft brackish water occurs both in the lower and upper aquifer. In both aquifers high values of the ratio are to be observed with the exception of the fresh water bodies. The high ratios indicate a long residence time of brackish waters.

Remarkable are the lower values observed in the three wells in the semipermeable Maassluis Formation. The number of data available there is too small to make conclusions, however.

A very interesting situation is found under the Haarlemmermeer polder. Fresh water originating from infiltration in the nearby dune area, has formed a fresh water body under the polder. It shows high ratio values for the lower part of this fresh water body whereas they are lower in the upward direction. In the upper thirty meters values usual for infiltrated precipitation water are found. The underlying salt water shows values equal of that of seawater. Calculations of Wit (1974) showed that at this site a strong inland-directed subsurface flow takes place. It is the result of the low phreatic levels in the Haarlemmermeer polder in the neighbourhood.

North of this point a totally different situation is found than in the southern half of the region. Due to the low position of the hydraulic base in this area the main aquifer, consisting of very coarse grained sediments, shows very high values for the transmissibility. In the part of the region crossed by this profile the main aquifer contains water with chloride values of more than 15,000 ppm. Detailed studies of this area, having been carried out during the last four years, showed that there is heavy salinization fed by an inland flow of saline water from a seaward direction. The ratios in this aquifer come near to that of seawater. Only near Den Helder somewhat higher values are found.

The upper aquifer, containing saline water as well, shows extremely high values of the ratio. It seems that waters with long residence times are being washed out by seawater recently infiltrated in the lower aquifer. This is in agreement with calculations by Wit (1980).

Similar remarks can be made about the other profiles shown. Under the dunes the fresh water body mostly shows very low ratios, with values increasing with depth. In South Holland (fig. 5a) also the inland fresh water body has very low ratios. Profile 2 (fig. 5b) shows a cauliflower shaped intrusion of fresh water from the lower aquifer into the upper aquifer. The ratios in this fresh water intrusion are similar to those in the fresh water body near Hoorn (fig. 5c) but show an increase near the interface with the brackish water in the upper aquifer.

CONCLUSIONS

- The shape of the level of the 500 ppm chloride plane in the subsoil water of the southern part of the western Netherlands is fundamentally different from that in the northern part.
- This difference can be explained by differences in the geohydrological situation in the subsoil.
- The various fresh water bodies can be divided into three groups with different origin:

a. the coastal fresh water body consisting of recently infiltrated precipitation water;
b. the inland fresh water body in South Holland consisting of a mixture of precipitation water and infiltrated river water;
c. the large inland fresh water body in the North of the region consisting of fossil water from unknown origin.

- The ratio between chloride and sulfate contents of the saline and brackish waters appears to be a useful tool to differentiate between seawater having infiltrated in recent and in earlier periods.
- Pure seawater never has been found in the subsoil of the region.

ACKNOWLEDGEMENTS

Thanks to Mr. H. Witt, who compiled fig. 2 and Mrs. Lenie van Goor who assisted with analyzing of data in the northern part of the area (Van Goor, 1981).

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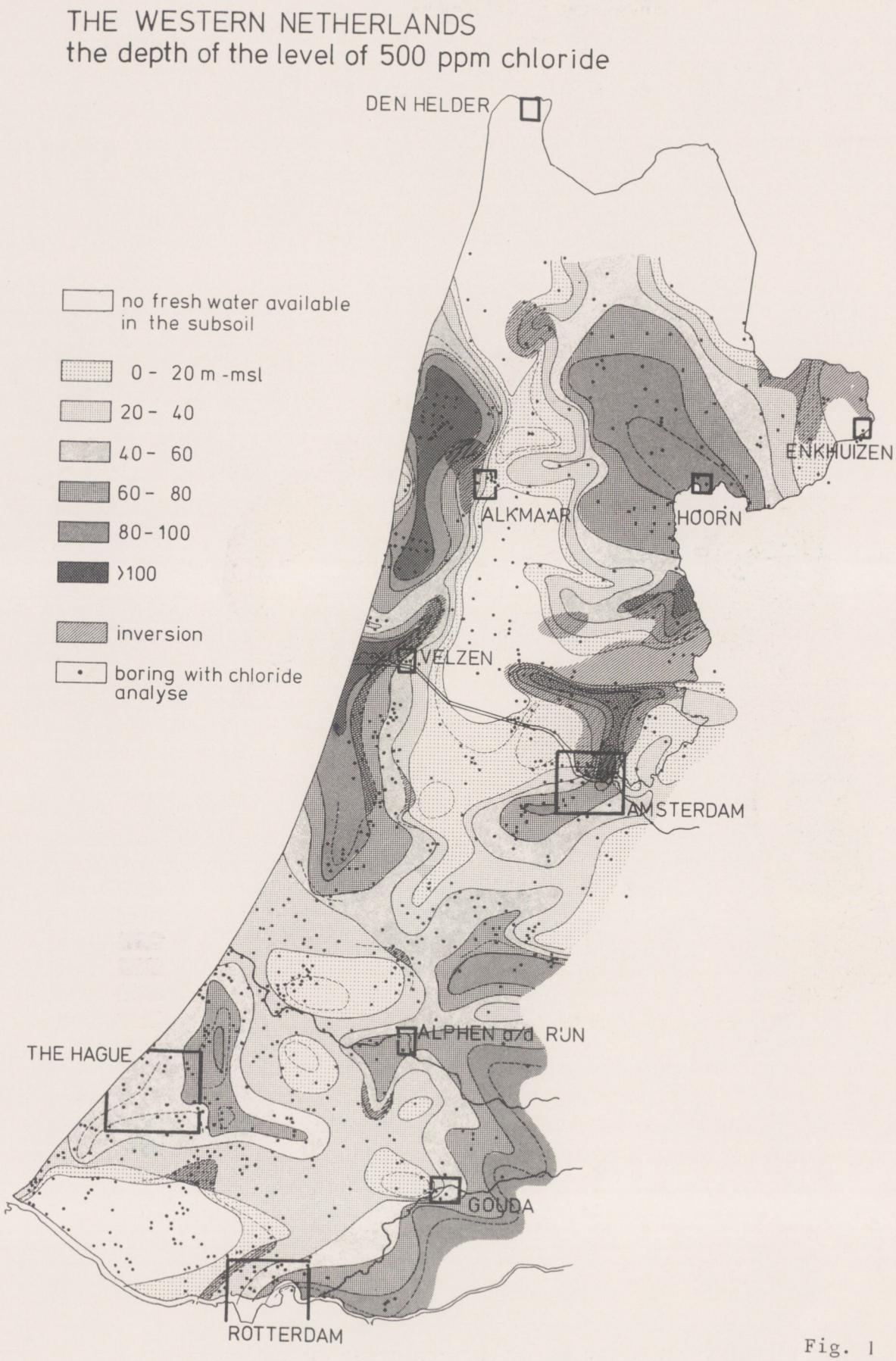

THE WESTERN NETHERLANDS
the depth of the level of 500 ppm chloride
DEN HELDER
no fresh water available in the subsoil
0 - 20 m -msl
20 - 40
40 - 60
60 - 80
80 - 100
>100
inversion
boring with chloride analyse
ENKHUIZEN
ALKMAAR
HOORN
VELZEN
AMSTERDAM
ALPHEN a/d RIJN
THE HAGUE
GOUDA
ROTTERDAM

Fig. 1

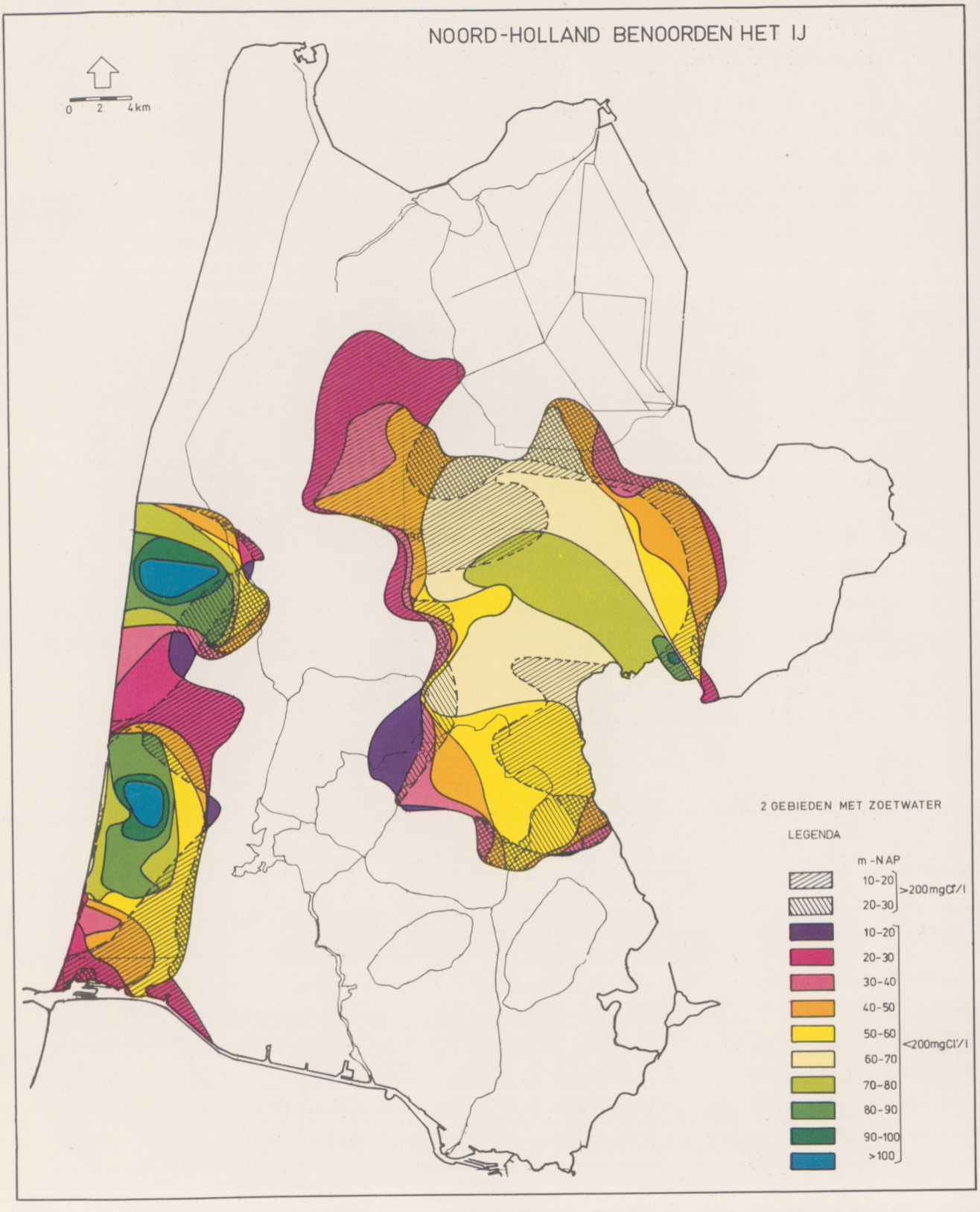


Fig. 2. The depth of the plane of 200 ppm chloride in the province of North Holland and the occurrence of inversions along the edges of the fresh water bodies in the area (Witt, 1980)

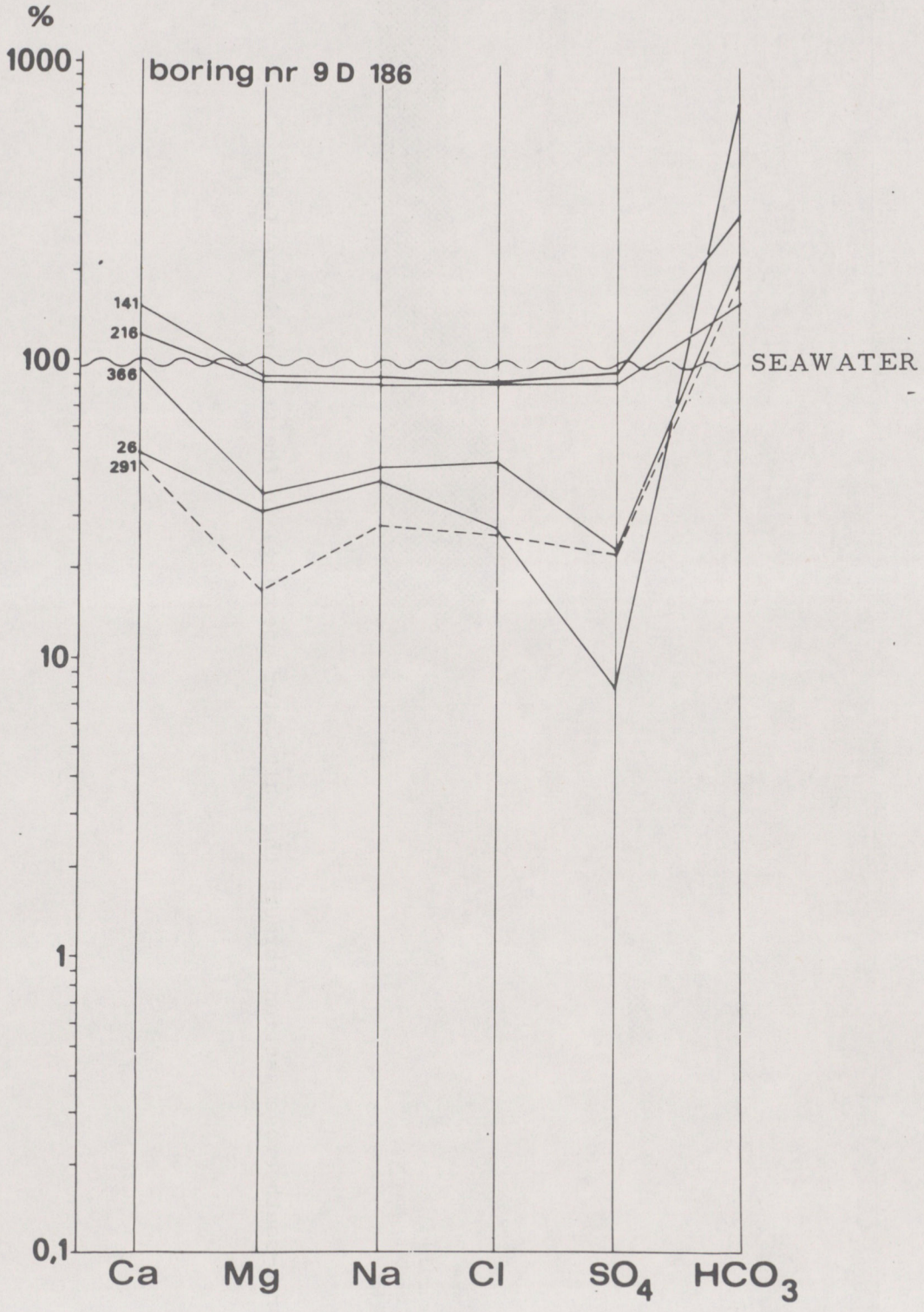


Fig. 3. A modified Schoeller-diagram of the boring 9D/186 near Den Helder. The contents of several components of groundwater are expressed in percentages of that of seawater

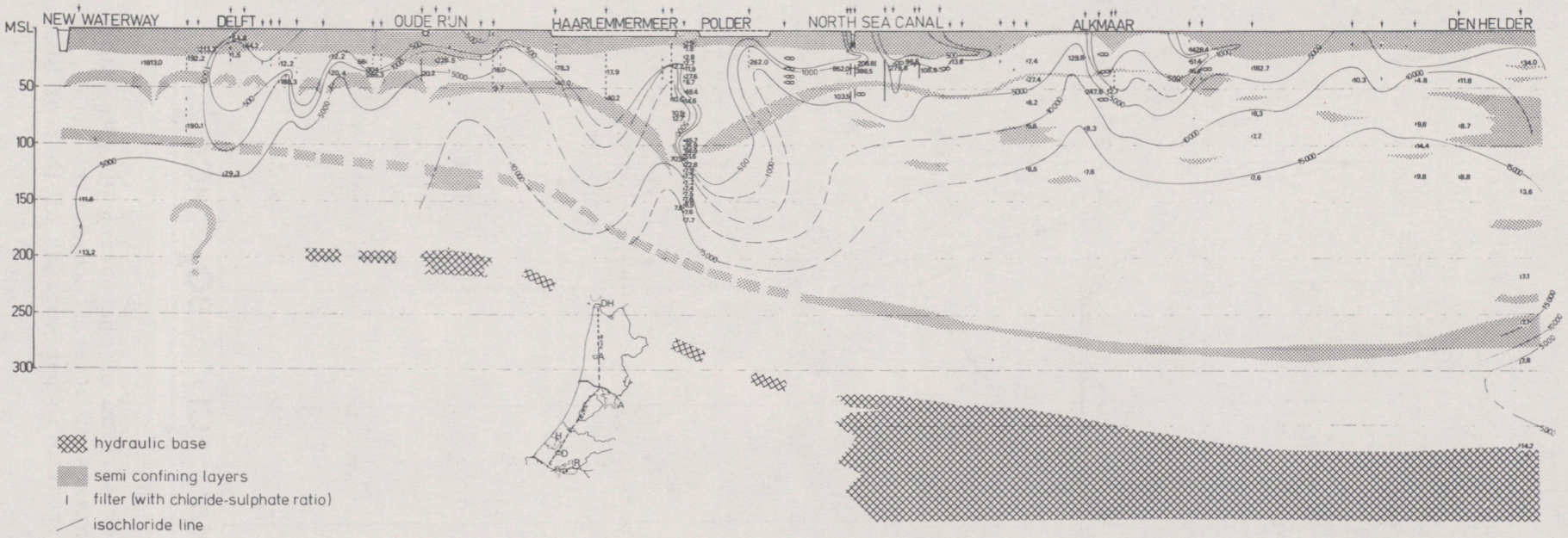


Fig. 4. A north-south cross section through the salt water upcomings in the western Netherlands

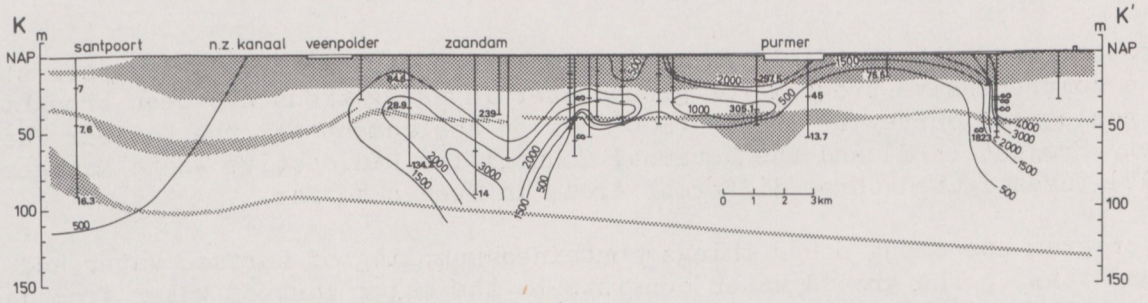
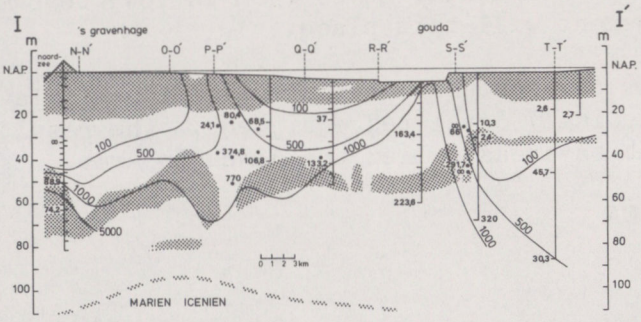


Fig. 5a



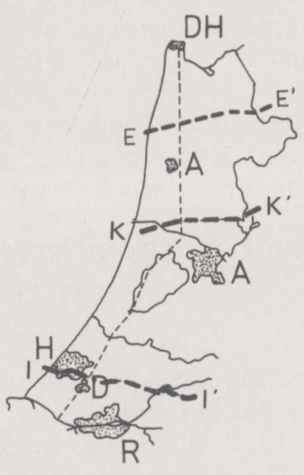


Fig. 5b

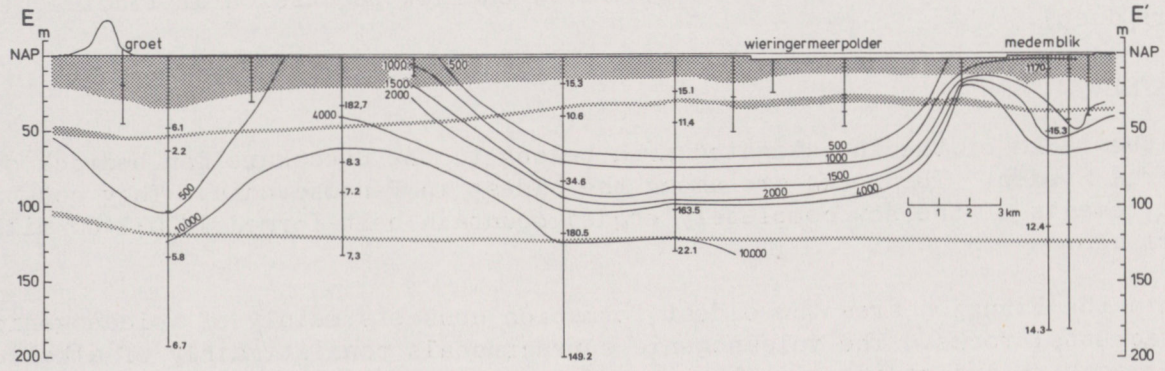


Fig. 5c

Fig. 5. Three east-west cross sections through the region

SALT WATER IN CONNECTION WITH GROUNDWATER SAMPLING IN DEEP BOREHOLES

By G. Gidlund[1]

ABSTRACT

In connection with the field investigations for future radioactive waste storage a comprehensive program for groundwater investigations has been brought out. The geological survey of Sweden (SGU) has on assignment for the Nuclear Fuel Safety Project (KBS) and the National Council for Radioactive Waste Management (PRAV) investigated three different areas in Sweden so far.

The program has among other things contained sampling of surface water and ground water. The ground water consists to the major part of water from deep coredrillholes (500-800 metre).

INTRODUCTION

Before it is possible to store radioactive waste in the bedrock one has to know what chemical and physical environments occur at the actual deposite depths. Other important questions are how the groundwater moves from inflow- to outflow-areas and at what velocity this movement will take place.

This paper describes the program for this groundwater investigation with a presentation of the results and the specially developed equipment. The results originate from one of our investigation-areas called the Finnsjön-area.

DESCRITPION OF THE INVESTIGATION-AREA

General

The Finnsjön area is situated approximately 100 km north of Stockholm and 20 km from the Baltic coastline in the northern part of Uppland in central Sweden (Figure 1). The area is one chosen for long term hydrogeological investigation where different methods and techniques are developed and tested.

In the investigation area, which is a 25 km^2 drainage basin, there are 8 deep core drillholes (500-700 metre deep) and 17 shallow percussion drillholes (100 metre deep).

Geology

All the rocks of the investigated area belong to the Svecokarelian bedrock of central Sweden. The rocks are among the oldest in Fennoscandia. They constitute remants of the now completely eroded mountain belt formed 1800-2100 million years ago.

Within the Finnsjön area the oldest formation consists mainly of volcanogenic supracrustal rocks. The volcanogenic supracrustals consist mainly of alkali-intermediate- and sodium-rich leptites. Interlayered in the leptites are basic vulcanites with accompanying iron ore, crystalline limestones and minor parts of clastic sediments. The supracrustals are usually stratified, foliated and often exhibit small-scale folding. The whole rock sequence is plastically deformed.

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1. SGU Geological Survey of Sweden, Uppsala, Sweden

In the Finnsjön area, as well as regionally, a differentiated intrusive suite ranging from ultrabasic to acid in composition has intruded the supracrustal formation. Most frequently, these intrusives have a granodioritic-granitic composition.

The supracrustal and intrusive rocks were later cut by numerous basaltic dikes, which are now altered to amphibolite.

The younger granite is the last big intrusion in this area and together with the granodiorite dominate the Finnsjön area. Earlier mapping of drill-cores shows that the granodiorite is limited in depth by younger granite. This granite has probably intruded in several phases but is not differentiated. It is massive and it brecciates all the other rocks which form the bedrock of the Finnsjön area. The younger granites can locally be regarded as macrobreccia. The central parts of the granite are not very well exposed and as a consequence the extension of the granites is somewhat uncertain.

The younger granite is followed by pegmatite and aplite which frequently crosscut the older rocks. Pegmatites also occur relatively often as small bodies in the eastern parts of the granite area. The most common fracture fillings in the area are calcite, prehnite and cloride.

Quaternary geology

The Late Quaternary development of the Finnsjö-area consists of the glaciation, deglaciation and land uplift. Peat deposits are widespread, particularly in the central parts. The most common sediment is a sandy silty till. Its $CaCO_3$ content varies between 10-20 %. The moraine is - because the area was situated under the coastline - more or less washed out. The outwashed material is distributed everywhere in depressions in the area. In the eastern parts, up to 3 m thick beds of outwashed sands were measured. The depressions are also partly filled with clays, especially with varved $CaCO_3$-rich glacial clay. In the central part of the drainage basin some minor glacio-fluvial deposits exist which are supposed to belong to the northern part of a more regional esker.

Hydrology

The northern part of Uppland has rather few lakes. Most of the region is drained towards the north to the Baltic Sea. The streams generally go in a north-southerly direction following the main feature on the regional tectonic lines.

The annual average precipitation is about 670 mm. Most of the precipitation falls as rain. From the middle of November to the middle of April the precipitation falls as snow.

The Finnsjön investigation-area is drained by two main water courses. The average run-off from the northern Uppland is calculated to around 240 mm/year.

The distance between groundwater level and ground level is in the higher part of the area between 3-4 metre. In the lower parts, the outflow-areas, the groundwater head conicide with the ground level or is above.

INSTRUMENTATION AND METHODS

The down-hole piston pump

One of the problems in connection with sampling of water in drillholes is how to know from which part of the hole the water originate. In our different investigations we have noticed rather big movements of water along the drillholes from different conductive zones that have been penetrated by the drillhole. In order to avoid to recieve an integrated sample of water from the entire hole the geological survey of Sweden has developed a groundwater pump which allow us to take samples between packers.

This pump give us opportunity to pump up water down to depths of 1000 metre in drillholes of a diameter of 56 mm. The pump is equipped with packers on each side of the intake section. The packers function as seals against the wall of the hole so that we know where the collected water come from. It is possible to use any distance from 0.5 m to 50 metre between the packers (Fig. 2).

Methods

From every core-drillhole ground water from about four different levels are sampled. The levels are chosen from earlier investigations brought out in the drillhole as for example, core-logging, geophysical logging and water loss measurements. The water is, as mentioned above, sampled between packers in order to determine its origin.

RESULTS OF THE INVESTIGATION

In connection with our standard investigation of the ground water conditions we have found salt water in crystalline bedrock. Based on analyses from our eight core-drillholes and seventeen percussion drillholes we have found that;

- Fresh water originate from the less permeable formations in the area.
- Saline water originate from the highly permeable tectonic lines.
- Fresh water in the formation from the same depths as the saline water in the surrounding tectonic lines are younger according to the age determination.

From a total of 39 analyses of the age of the water in the area we have found that the saline water have an average age of approximately 7000 years and the fresh water have an average age of about 4400 years. (See Table 1 and 2.)

Table 1. The ages of the saline ground water in the Finnsjön investigation area.

	Core drillhole	Level	Age BP (corrected $\delta^{13}C$)
SALINE WATER	Fi 5	141 m	$\bar{A}$ = 10000 YEAR
		205 m	$\bar{A}$ = 10500 YEAR
		297 m	$\bar{A}$ = 4800 YEAR
		384 m	$\bar{A}$ = 10500 YEAR
	Fi 7	301 m	$\bar{A}$ = 4400 YEAR
		322 m	$\bar{A}$ = 4500 YEAR
		511 m	$\bar{A}$ = 4400 YEAR

Table 2. The ages of the fresh water in the Finnsjön investigation area.

	Core drillhole	Level	Age BP (corrected $\delta^{13}C$)
FRESH WATER	Fi 1	206 m	$\bar{A}$ = 2100 YEAR
		293 m	$\bar{A}$ = 2500 YEAR
	Fi 2	385 m	$\bar{A}$ = 3750 YEAR
		504 m	$\bar{A}$ = 3750 YEAR
	Fi 4	152 m	$\bar{A}$ = 3750 YEAR
		247 m	$\bar{A}$ = 6100 YEAR
		368 m	$\bar{A}$ = 5400 YEAR
		534 m	$\bar{A}$ = 5400 YEAR
	Fi 7	123 m	$\bar{A}$ = 3900 YEAR

All watersamples from the core-drillholes are plotted in Piper diagrams shown in Fig. 3 and 4.

All drillholes except one which contains saline water are placed into or nearby a major tectonic zone.

The way the saline water appears in this area it will act as very effective barriers and prevent any regional groundwater flow.

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Gidlund, G., 1978: KBS technical report nr 62.

Hultberg, B., 1981: KBS, Grundvatten i kristallin berggrund./ Groundwater in crystalline bedrock/

Jacks, G., 1978: KBS technical report nr 88.

Fig 1. Location of the Finnsjön test area

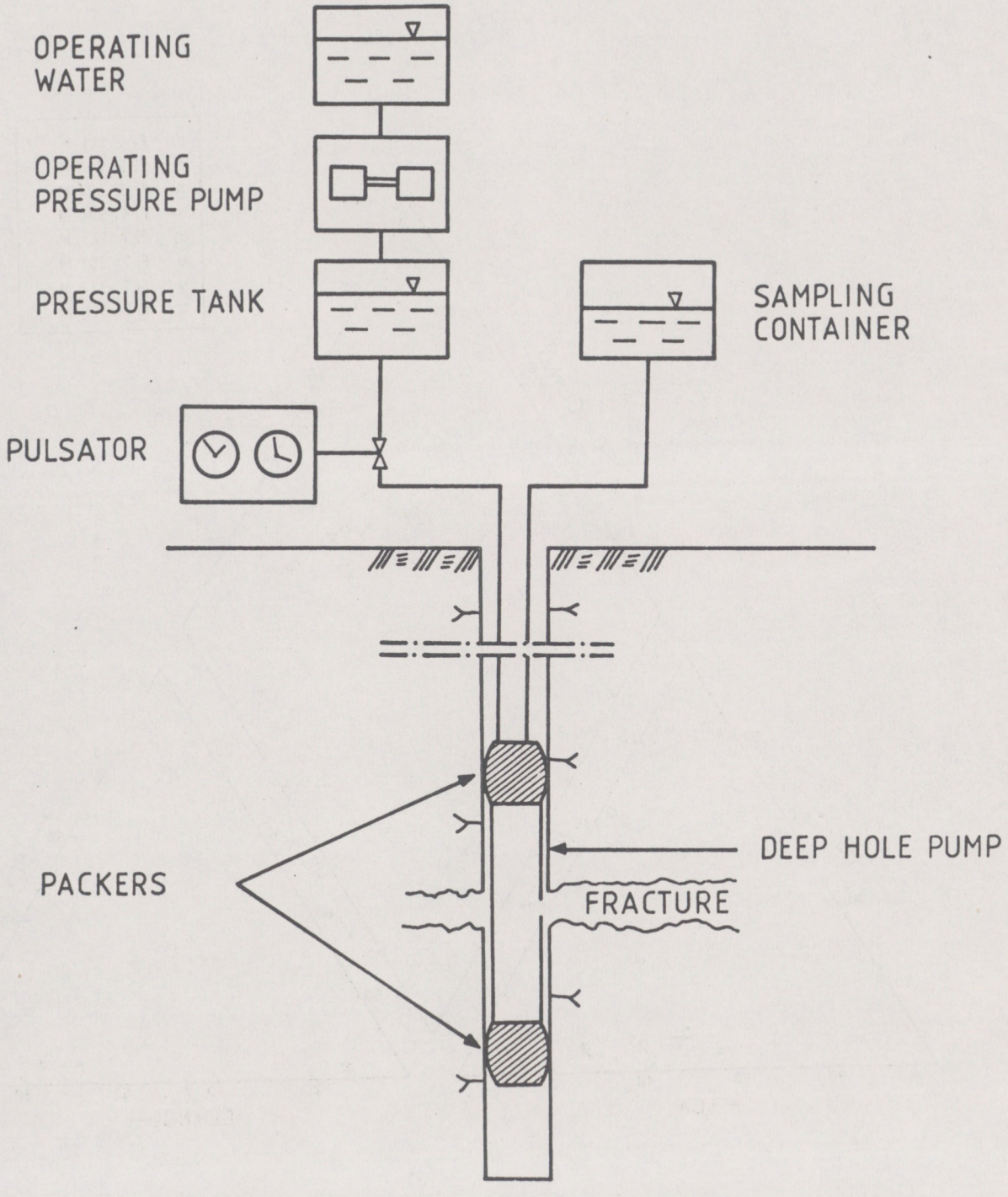

Fig 2. Schematic illustration of deep hole pumping

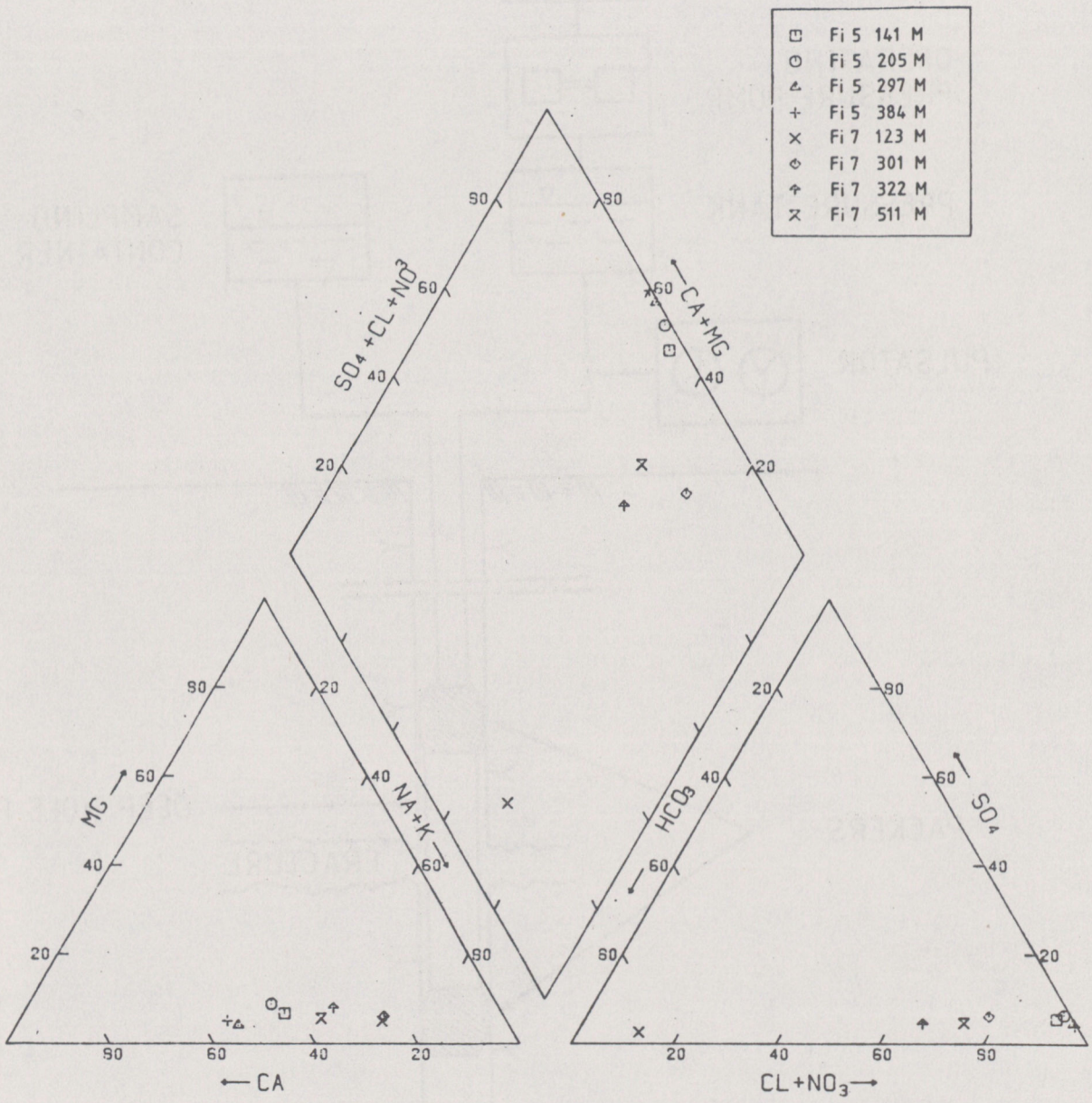

Fig 3. Piper diagram for the saline groundwater in Finnsjön test area

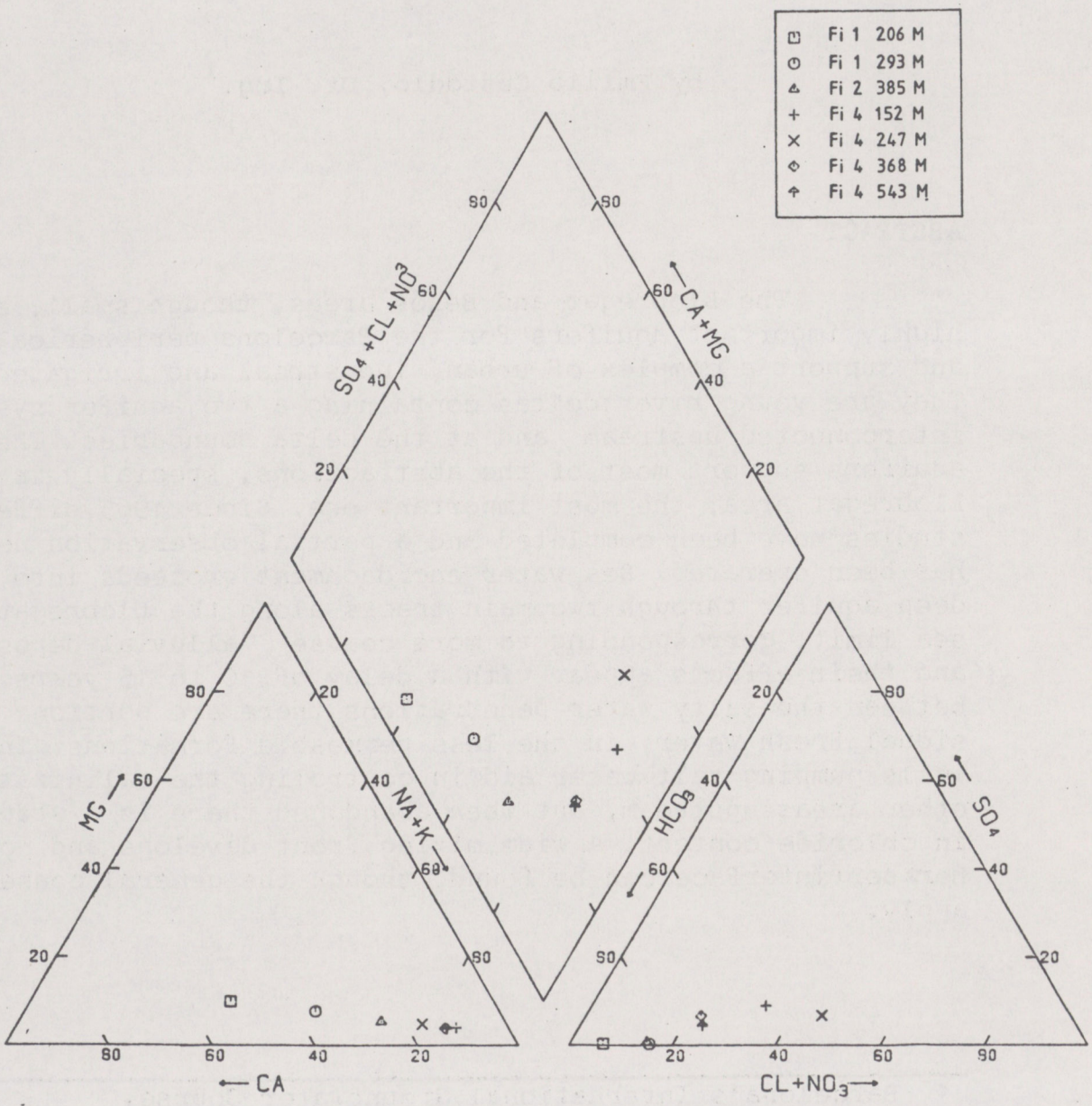

Fig 4. Piper diagram for the fresh groundwater in Finnsjön test area

SEA WATER ENCROACHMENT IN THE LLOBREGAT AND BESÓS AREAS, NEAR BARCELONA (CATALONIA, SPAIN)

by Emilio Custodio, Dr. Ing.[1]

ABSTRACT

The Llobregat and Besós areas, though small, are highly important aquifers for the Barcelona peripherical zones, and support a complex of urban, industrial and irrigated areas. They are young river deltas containing a two aquifer system, interconected upstream and at the delta boundaries. The deep aquifers support most of the abstractions, specially in the Llobregat area, the most important one. Since 1965 different studies have been completed and a partial observation network has been operated. Sea water encroachment proceeds into the deep aquifer through two main tracts along the Llobregat delta sea limit, corresponding to more coarse alluvial deposits, and their effects appear with a delay of 10 to 15 years. In between the salty water penetrations there are portions of residual fresh water, in the less permeable formations. Industrial wells pumping salt water aid in controling the salinization of other areas upstream, but when abandoned there is a step increase in chloride content. A wide mixing front develops and no Ghyben-Herzberg interface can be found, though the general consequences apply.

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1 Barcelona's International Groundwater Course. ETSIIB.- Polytecnical University of Barcelona and Eastern Pyrenees Water Agency.

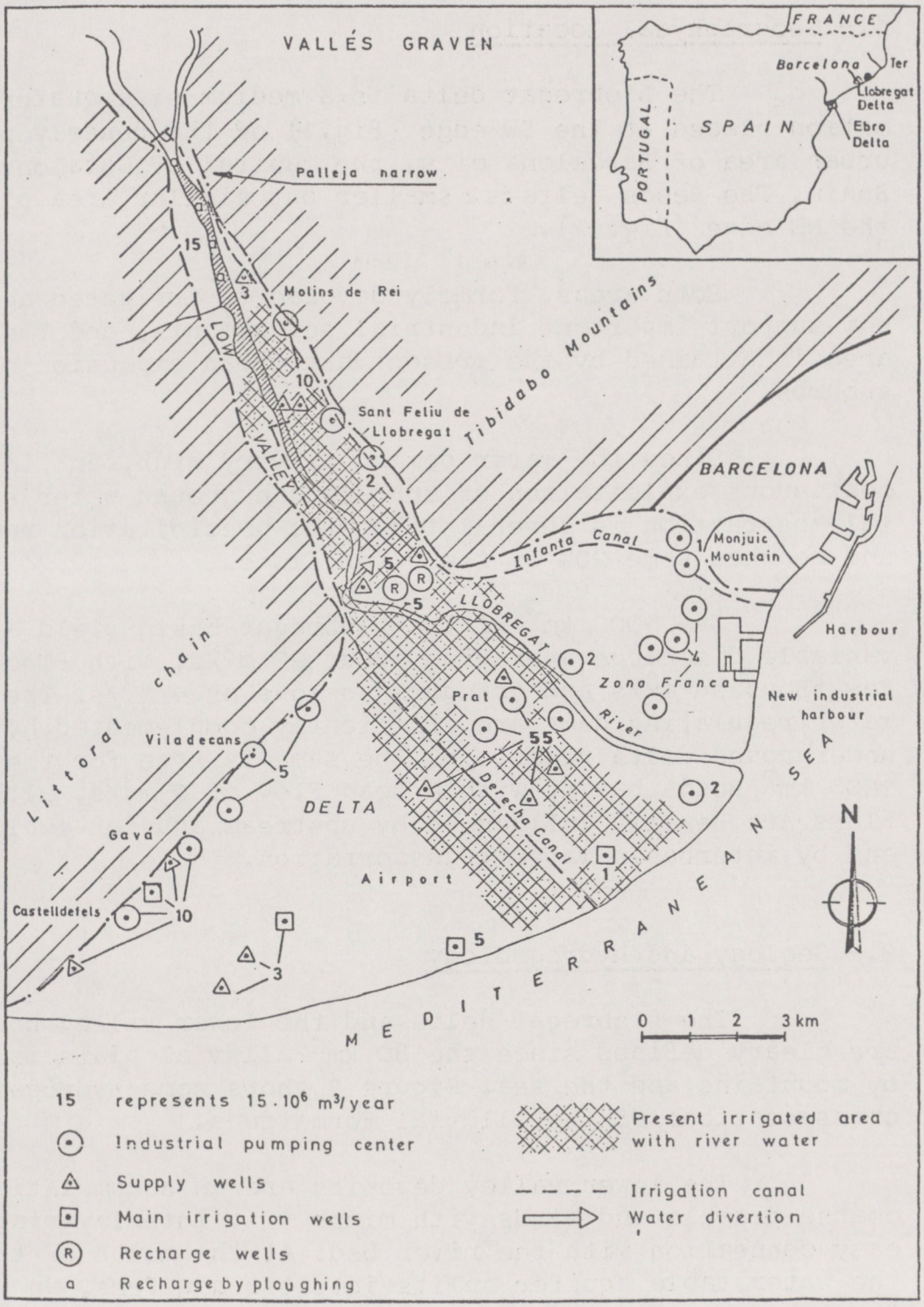

Fig. 1

Situation of the Llobregat delta and ground water abstraction in million m^3/year, in 1980.

1.- Geographycal location

The Llobregat delta is a medium-size quaternary formation placed at the SW edge (fig.1) of the densely populated urban area of Barcelona city, the capital of Catalonia, in NE Spain. The Besós delta is a smaller but similar area placed in the NE edge (fig. 2).

Both areas, formely devoted to irrigated agriculture, now support important industrial settlements, and the Besós area is attained by the present fast urban expansion of the city suburbs.

Since the water demand is very high, an intense and continuous exploitation of surface and ground water exists. Total abstraction is several times the precipitation recharge, and reaches some 20% of the river flows.

The 5000 km^2 of the Llobregat basin yield a highly variable discharge averaying about 20 m^3/s, with floods until a few thousand m^3/s and draughts so lows as 4 m^3/s. The surface water regulating dams are efficiently complemented by the underground delta reservoir. The same is true for the smaller, 1000 km^2 Besós basin, with a mean flow of 2 m^3/s, although the flows are heavily influenced by upstream aquifer exploitation and by interbasin water transportation.

2.- Geology and hydrogeology

The Llobregat delta and the lower valley boundaries are cleary defined since the 80 km^2 alluvial plain is limited by mountains and the sea. Figure 3 shows some hydrogeological cross-sections of the alluvial formations.

The lower valley deposits are an accumulation of coarse gravels and sands with minor silt interlayerings, in easy connection with the river bed. At the mouth of the valley, the water table aquifer splits into two aquifers, an upper water table aquifer and a deep confined aquifer, separated by a low permeability clay, silt and fine sand lens, that thickens towards the sea the lens reaches a thickness of about 50 meters near the shore.

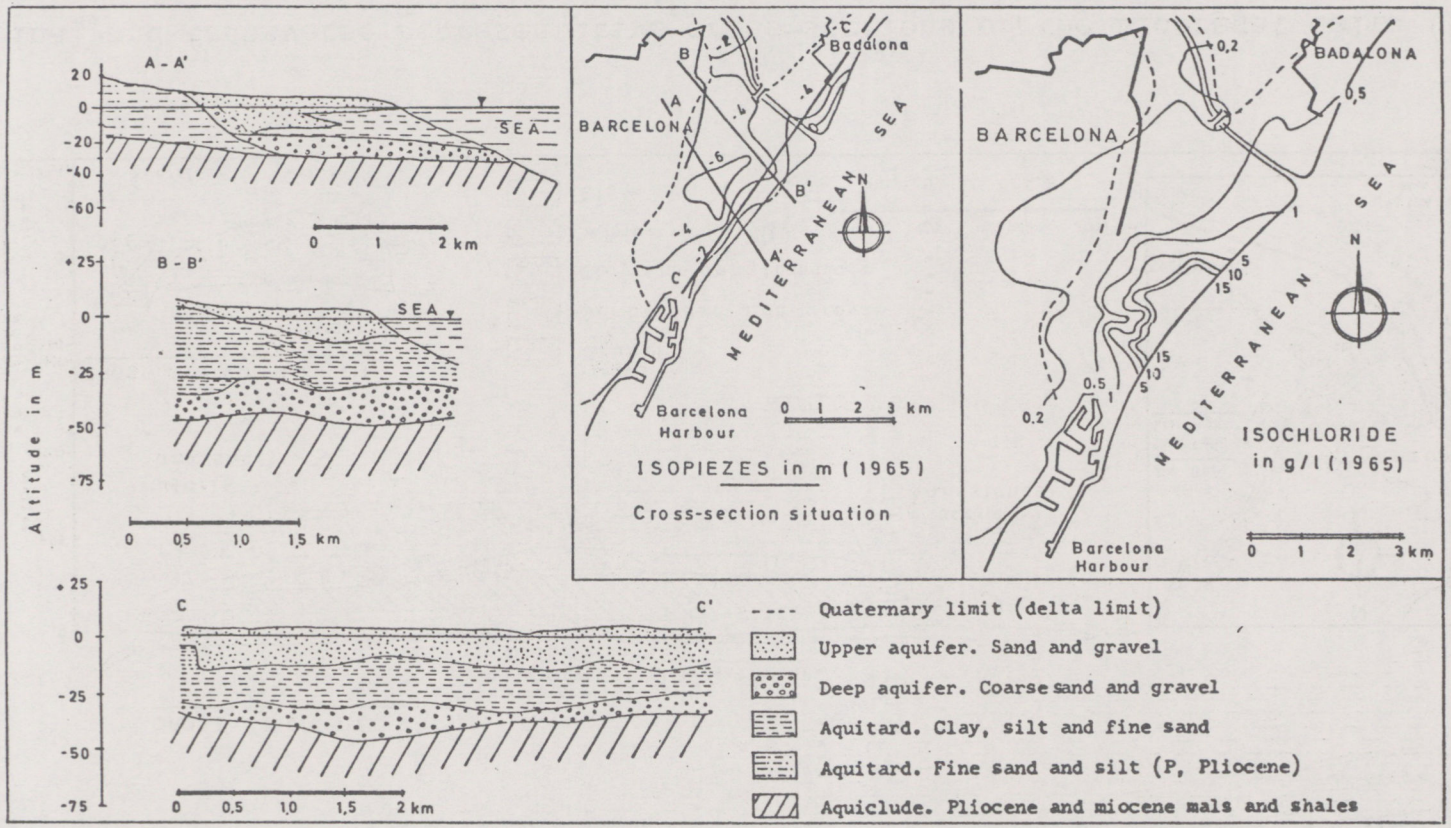

Fig. 2.

Hydrological cross-sections, piezometric map and isochloride map of the Besós delta deep aquifer. The upper aquifer was almost fully drained (11) (17) (23).

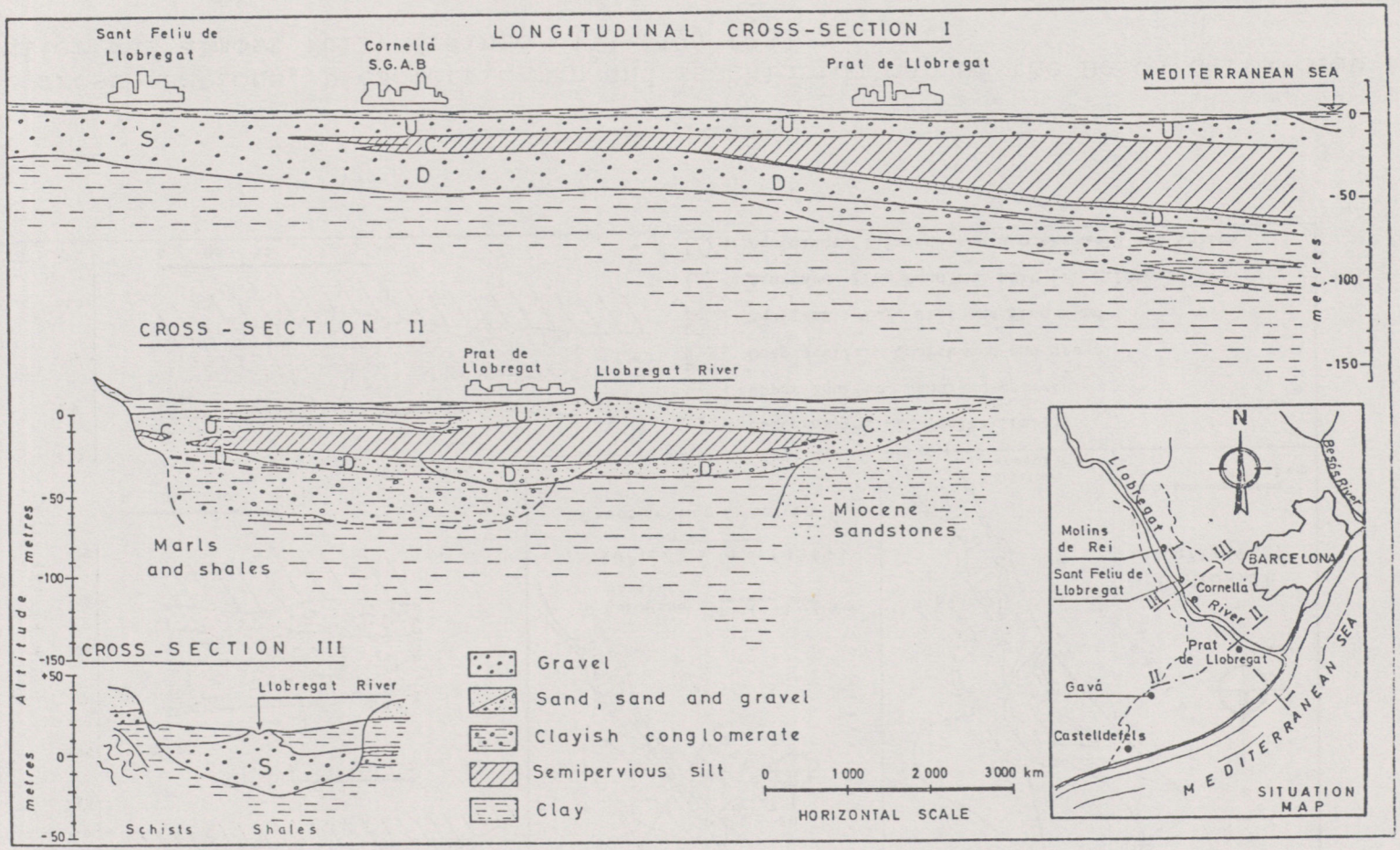

Fig. 3.

Longitudinal and transverse representative cross-sections of the Llobregat delta (5) (17).

This lens may be regarded as an aquitard whose vertical permeability varies from a negligible value near the central coastal zone (almost impervious) to a high value at the valley mouth and near the delta boundaries, where the two aquifers are clearly connected or no separation exist at all.

Near the delta boundaries, the situation may be more complicated in detail, as shown in figure 4. In these areas, lagoon, marsh and beach conditions ocurred alternatively during the sediment built-up, jointly with alluvial influences from local creeks and sometimes from the main river.

The upper aquifer discharges directly into the sea through a sandy bottom, except at the eastern part, where the sea floor is dominantly muddy. The coarse highly pervious gravel and sand sediments of the deep aquifer concentrate main in a strip area through the center of the delta, with a short lobe extending towards the eastern border. Sea-deep aquifer connections are poorly known, but it is believed that the main formation may outcrop on the sea floor some 4 km off-shore, bellow some 110 m of water, in front of the center of the delta. Probably the effective thickness and the permeability decreases sharply towards the outlet.

The deep aquifer lye on blue-clay Pliocene deposits, which can attain more than 700 m thickness (17). Near the coast some clayish conglomerate lens appear, but they may be regarded as relatively impervious and of a small extension.

The geographic delta is a relatively young formation because in Roman times the shore was almost 2 km inland, as shown by some archeological founds. In the Middle Ages, the Barcelona harbour was placed at the shelter of the Western side of Montjuic Mountain, some 1.5 km landward from present shore line. Increased solid transport from the rivers was caused probably by intensive deforestation and inadequate agricultural practices, which lead to an intense land erosion.

The Besós delta is similar, as shown in figure 2, but at a smaller scale (17).

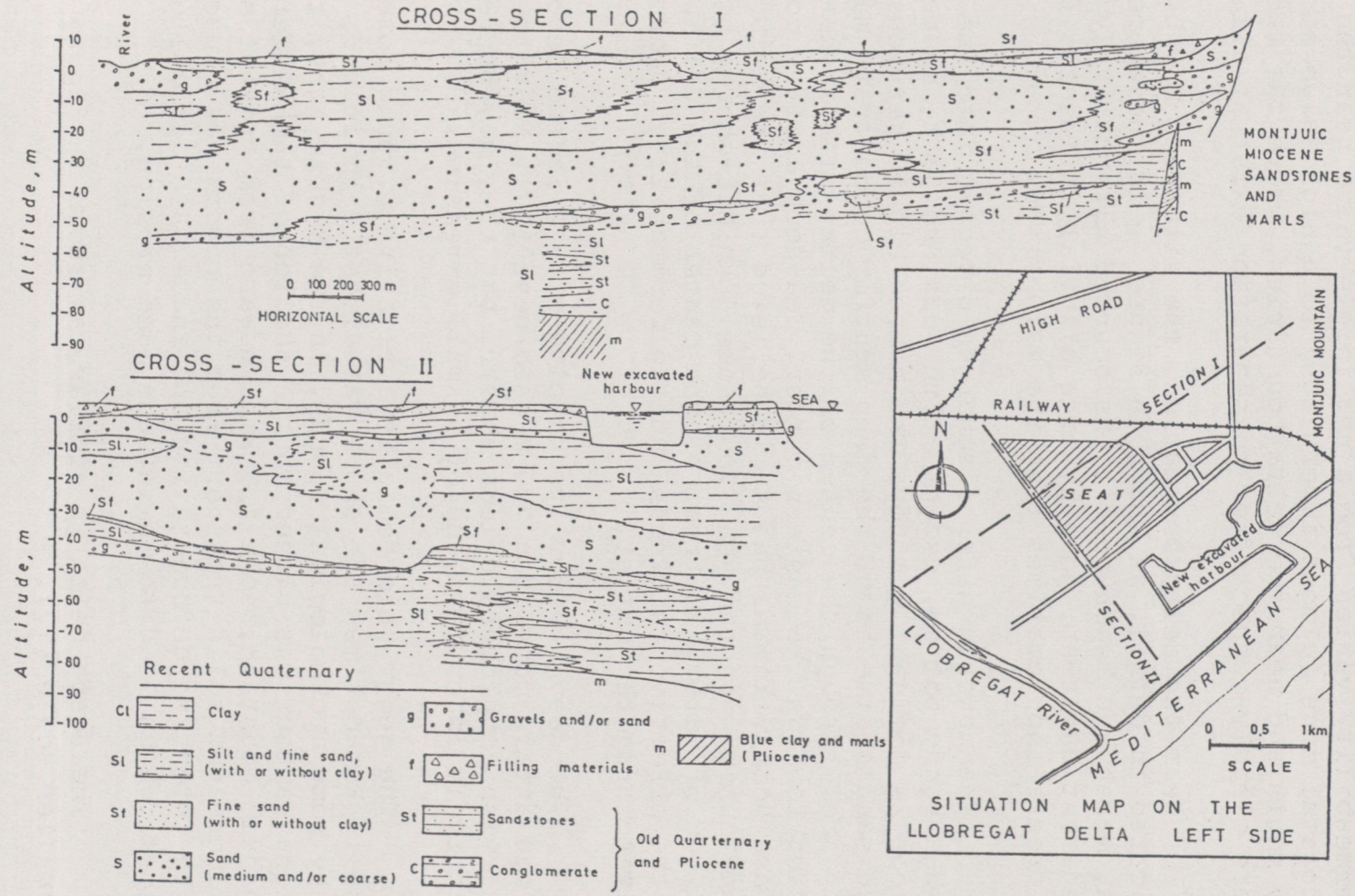

Fig. 4.

Geological cross-sections in the NE border area of the Llobregat delta (1) (2).

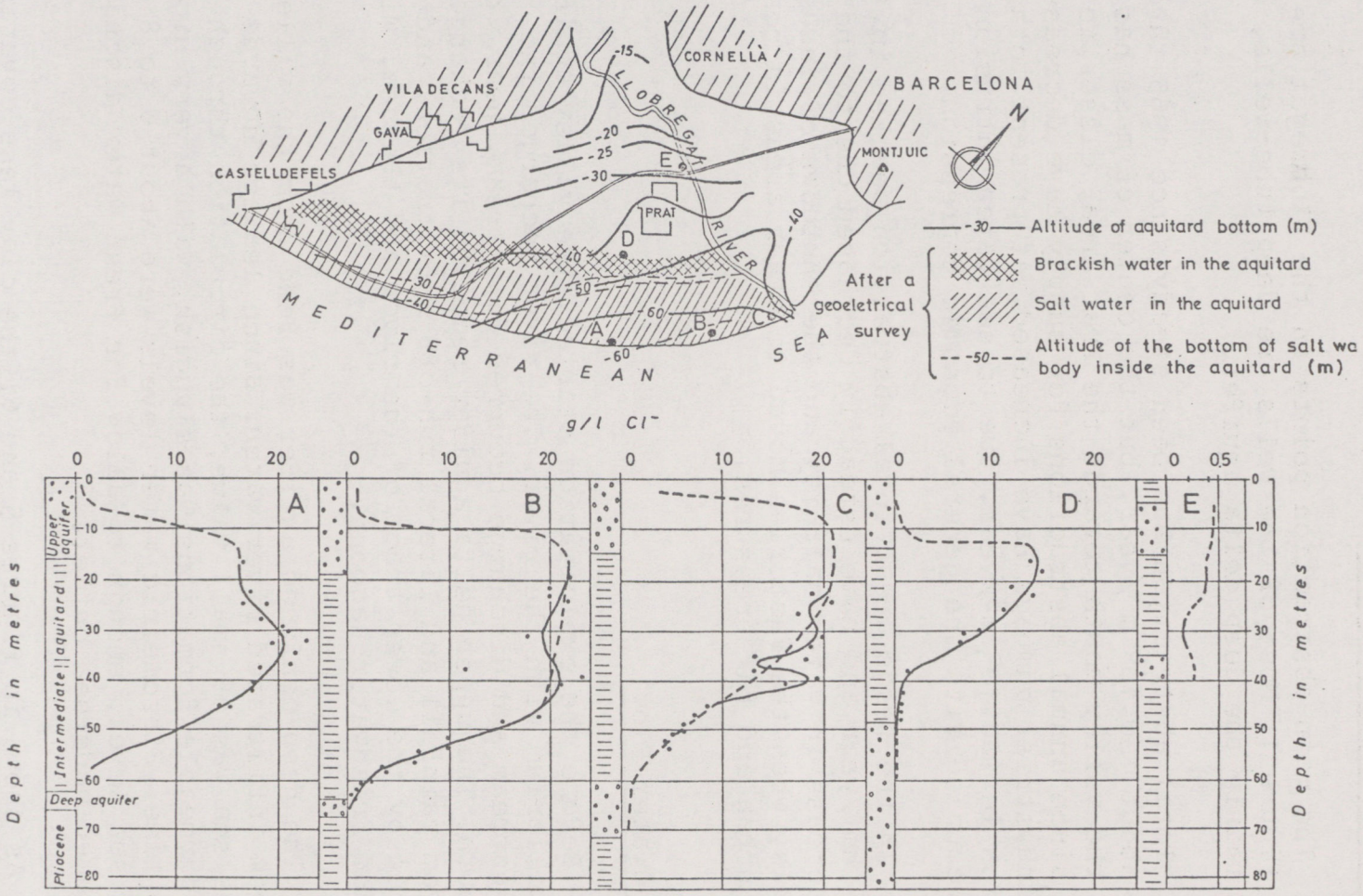

Fig. 5. Chloride-ion logs of several cored bore-holes in the Llobregat delta (8) (10).

3.- Groundwater abstraction

The main abstraction points in the Llobregat are shown in figure 1. Most of the wells are deep tube-wells, exploiting mainly the deep delta aquifer.

Total abstraction has been steady since 1965 through 1978, at about 130.10^6 m^3/year, but the center of mass has moved from the valley mouth towards the Prat area, closer to the sea, since annual abstractions for supply have decreased and the industrial pumpage have increased. At present total abstraction is 100.10^6 m^3/year, due to sea water intrusion problems coupled with the general economic depression.

In the Besós area, total abstraction peacked in 1965 at 60.10^6 hm^3/year and have esteadily decreased due to the excesive increase in water salinity and the shutdown of industrial wells, when the factories moved to other areas in response to urban and tax pressures.

4.- Groundwater flow

Before the exploitation started, the valley aquifer was in equilibrium with the river. The river recharged the aquifer upstream and discharge occurred downstream, near the valley mouth. The upper delta aquifer received its recharge mainly from rainfall and lateral run-off infiltration, and it was drained by the lower tract of the river and the sea, with marshes and freshwater lagoons.

The deep confined aquifer was feed from the valley, mainly from infiltrated river water. Since leakage upwards and to the sea was of small value, the circulation through the more permeable formations was sluggish, with a very high residence time. Piezometric water levels were about 6 to 8 m over mean sea level, enough to allow for fresh water discharge at the assumed undersea outlet.

As shown in figures 5 and 6, the clay lens covering the deep aquifer still contains connate sea water, trapped between 5000 to 10000 years ago. Therefore, vertical upward flow can be taken as negligible. Only in marginal areas, where ver-

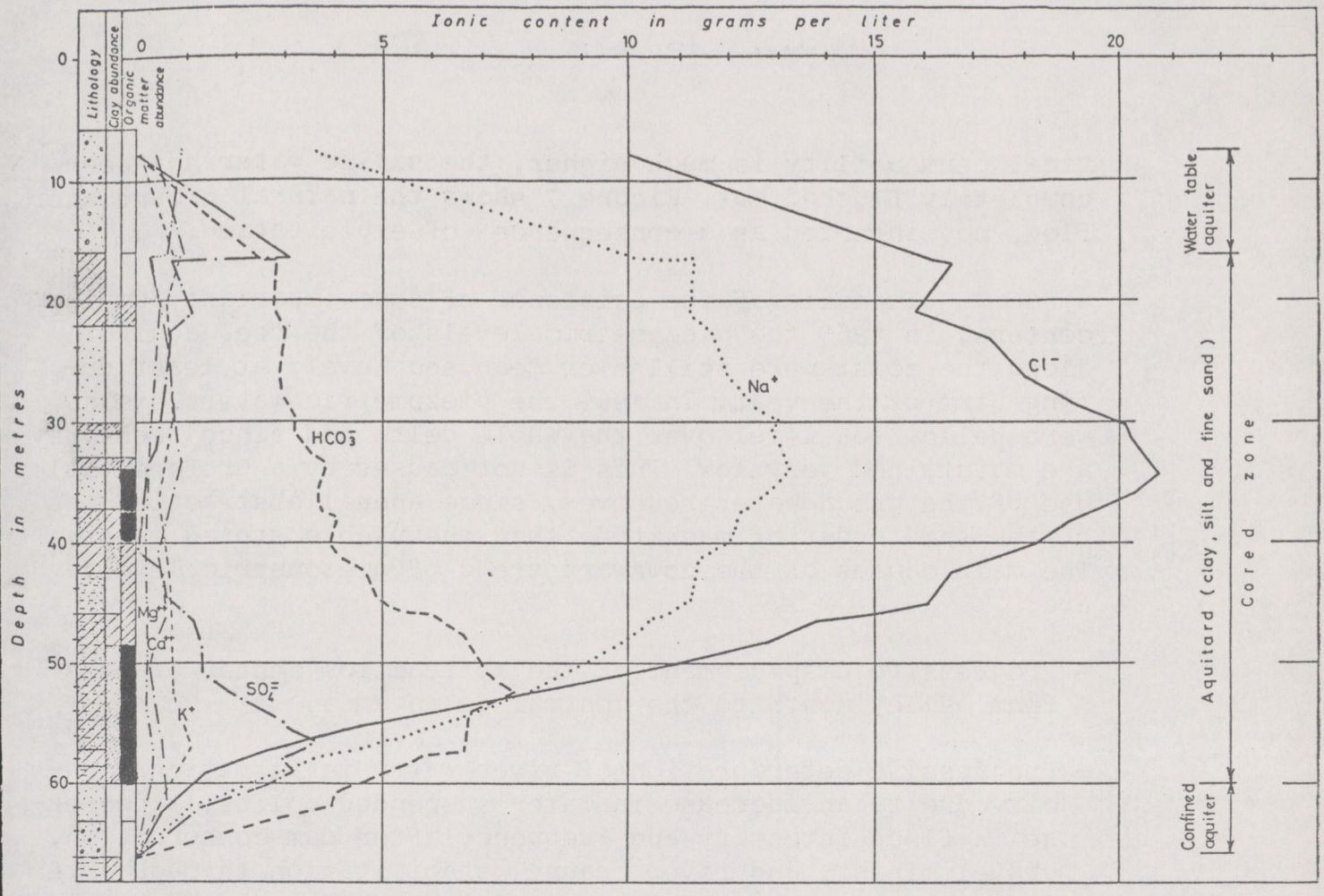

Fig. 6. Ion concentration in the pore-water in the delta sediments cored near the Llobregat delta shore (8) (10).

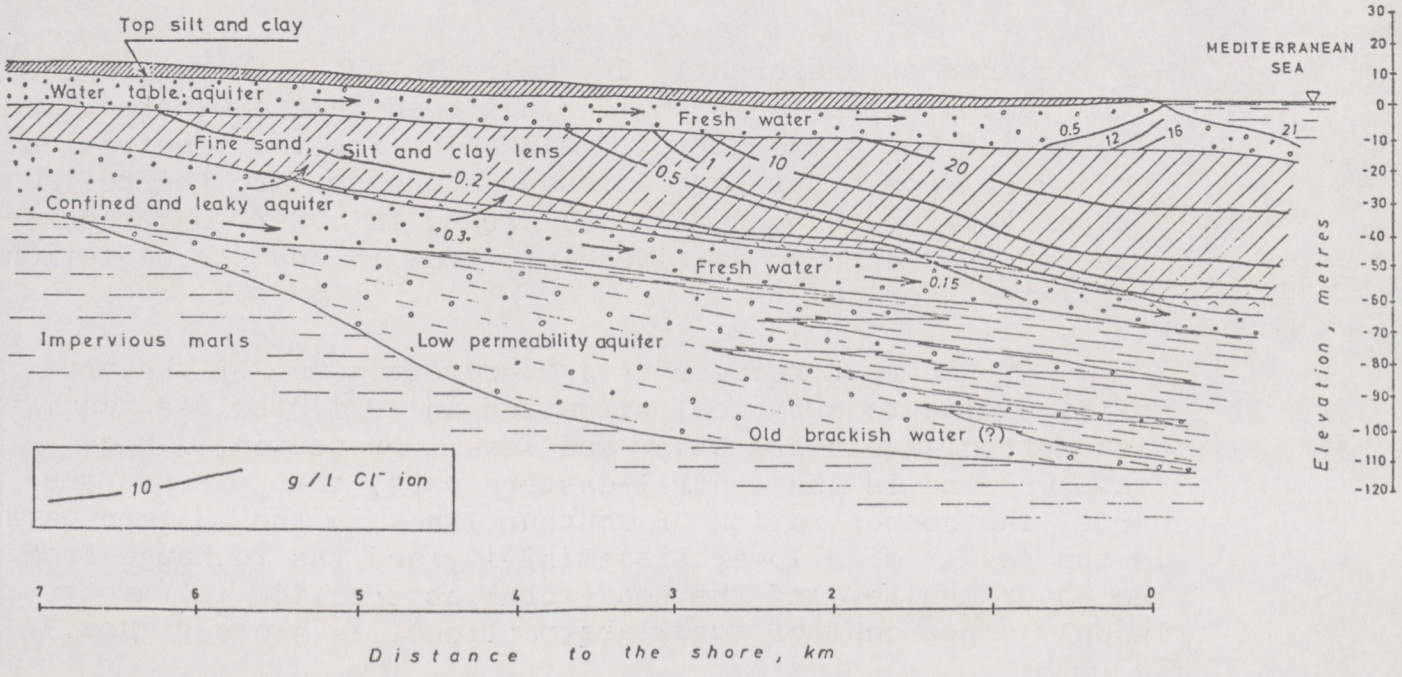

Fig. 7. Groundwater flow and salinity distribution in a longitudinal cross-section of the Llobregat delta (8).

tical permeability is much higher, the saline water has been completely flushed out. Figure 7 shows the natural groundwater flow, now inverted as a consequence of exploitation.

In spite of the existence of some important pumping centers, in 1960 the piezometric levels of the deep aquifer along the coast were still over mean sea level, at least during part of the year. In 1965 the piezometric water levels were bellow sea level over the whole delta and since then they are maintained very low. This is not caused by a preferential use of the groundwater reserves, since annual abstraction is of the same order of magnitude than the usable stored water. The main causes of the downward trend of piezometric levels are:

- Progressive displacement of the abstraction center of mass from valley mouth to the central delta area.

- Progressive deterioration of river water infiltration capacity due to an increase in water suspended solids, a decrease in flood intensity and frequency after dam construction, gravel mining, and river channel stabilitation through artificial banks. Now the river remains perched over the water table and then, the river infiltration is not regulated by the water table position. Only can be increased by some artificial recharge activities, applied from time to time (7) (12) (13)

- A sequence of preferently dry years in which the storage of the valley water table aquifer has been half exhausted. The resulting low piezometric levels allow for the penetration of water from other sources such as the sea and the delta upper aquifer through marginal areas, in order to compensate for the decrease in river infiltration and the depletion of water storage.

Figure 8 represents a piezometric map of the area corresponding to a typical situation in 1971. The deep aquifer piezometric levels are below sea level, presenting a huge drawdown cone in the central-eastern part, that extends seawards. The concentration of contour lines in the eastern part is the result of a lower transmisivity and the recharge from the upper aquifer and the sea (total abstraction in this area is only 7 per cent of total abstraction), as shows figure 9.

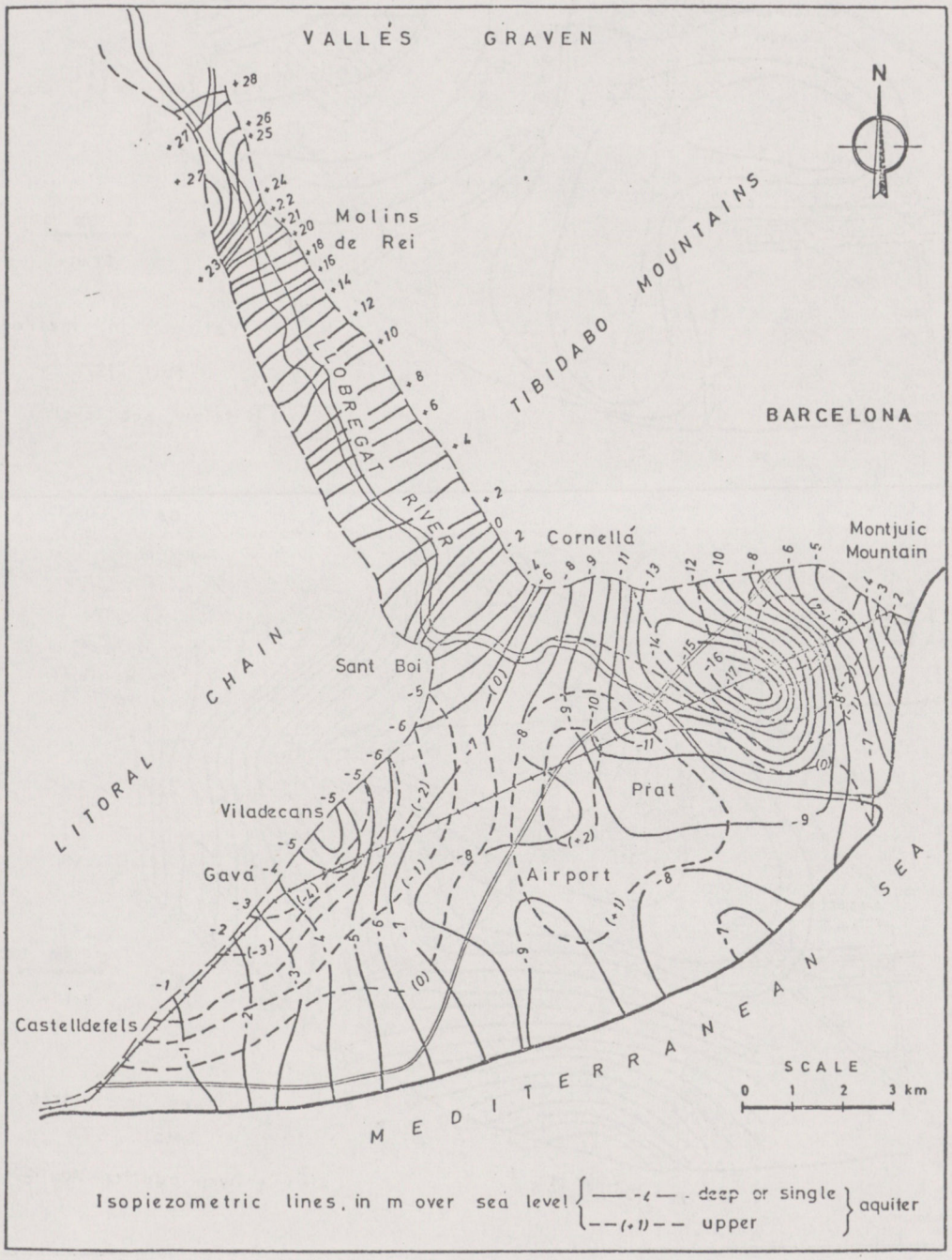

Fig. 8.

Isopiezometric map of the upper and deep Llobregat delta aquifer in 1971.

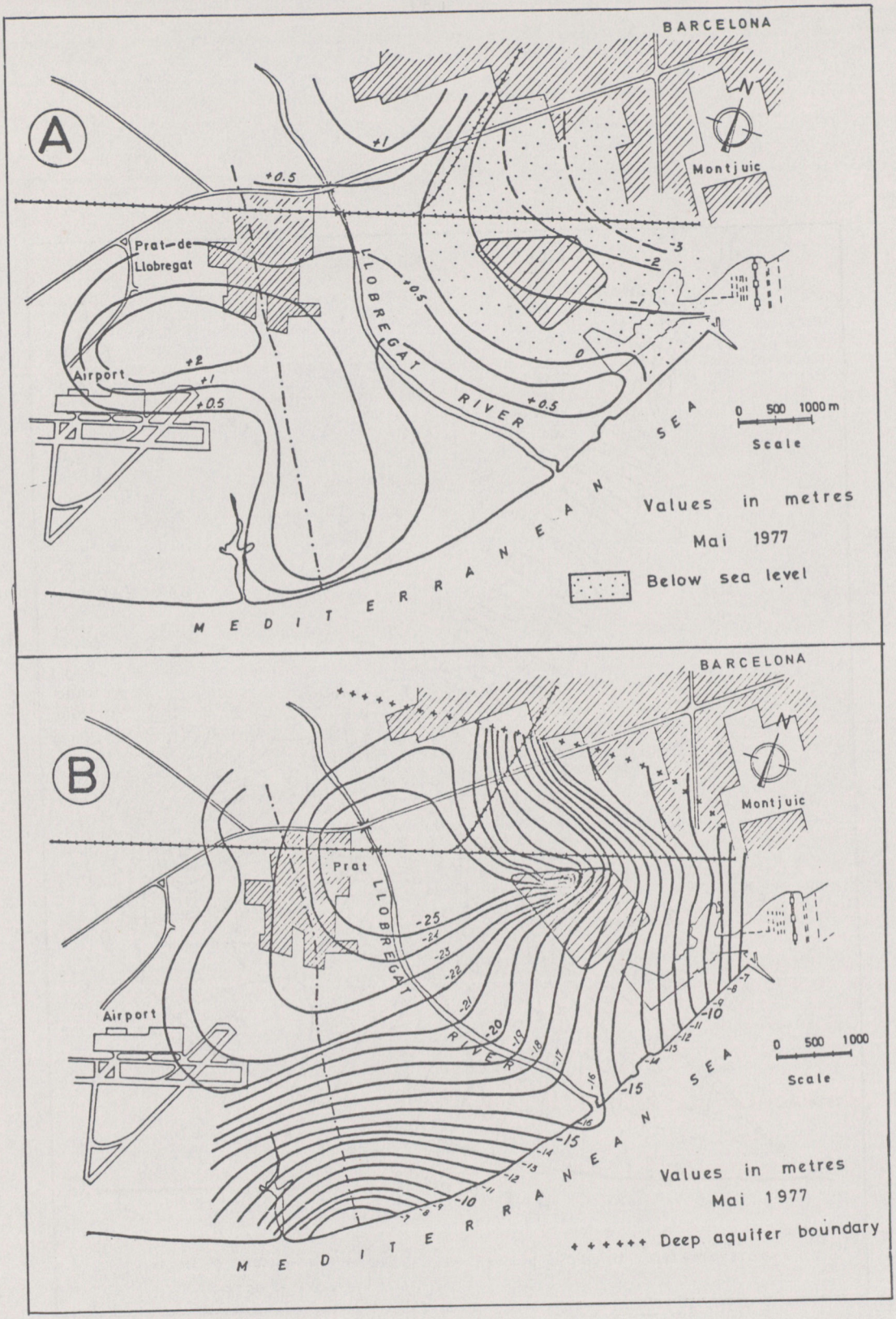

Fig. 9. Detailed isopiezometric map of the eastern part of the Llobregat delta. A: upper aquifer. B: deep aquifer. Data from 1977.

In the central coastal part, a small elevation in water level corresponds with the area of higher transmisivity and shows the effect of an easier connection with the sea. The east-west movement of groundwater in the deep aquifer is negligible since transmisivity in the western area is small.

In the central part of the delta upper aquifer, water levels are over sea level because recharge from rainfall, canal losses and return irrigation water from canals is high. This water mound descends towards the river and the sea and also towards the western delta boundary, where water is transfered from the upper to the deep aquifers.

A similar situation existed in the upper aquifer in the left delta side until 1970. Afterwards, irrigation with river water has been supressed in this area, and intensive industrial settlements, with paved areas and intense land occupation took place. Then, recharge was greatly decreased, Drainage towards the deep aquifer through the boundary areas creates a wide zone with water levels below sea level, prone to sea water intrusion. This situation has been agravated with the inland extension of the industrial harbour. Figure 9 shows the detailed situations in the central-eastern zone in May 1977.

Different studies have been carried out in order to understand the groundwater movement in the area from a qualitative and quantitative point of view (4) (5) (6) (8) (10) (15) (16) (20) (22).

Some observation bore-holes existed in the valley aquifer since 1944 and, in 1965, a dense but not complete network of observation bore-holes were stablished and since then are routinely observed.

Figure 10 shows the natural radioisotope content of the Llobregat's deep delta aquifer water. Data are ajusted to the year 1973. Recent waters from the river fill only the inner part of the delta until the main pumping centers. Infiltration in the river or through canals and irrigated land, account for about 60 to 75 % the total recharge.

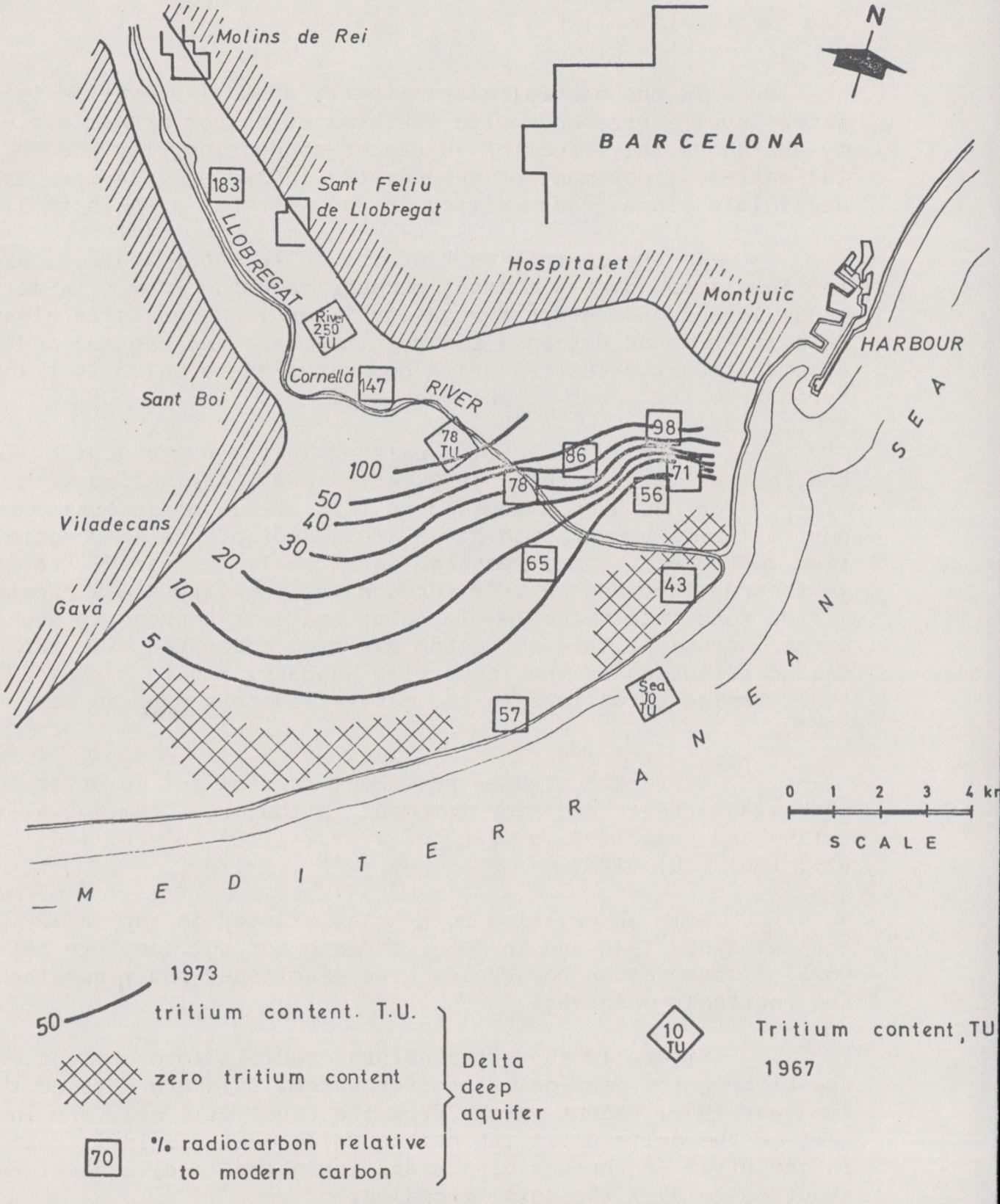

Fig. 10. Tritium and radio-carbon content in the Llobregat's deep aquifer waters.

Along the coast the tritium content is zero because there is old fresh water. The situation is more clear in the less permeable materials of the deep aquifers below the river mouth area.

The radiocarbon content shows the same facts, with nuclear bomb radiocarbon until the 30-50 TU line. Along the 10 TU line, the radiocarbon content shows just recent water without nuclear bomb contamination, and along the coast the radiocarbon content is clearly below that of recent non-contaminated groundwater. Maximum water ages are from 3000 to 4000 years in the less permeable sand formations. No data exist of the two main areas through which sea water intrusion proceeds.

Water in the 50 and 100 TU lines in 1973 corresponds to rainfall and river water in 1962, and water in the valley mouth is less than one year old. Then transit time between these two places 6 km far is about 11 years.

A similar figure is obtained when the changes in the chloride content in the river water are compared with the changes in the Prat wellwater, as shown in figure 11. The great increase in chloride content in river water is a consequence of mining activities upstream, to extract ClK from salt deposits. Thus, recent river water is easily identifiable (5), and their underground movement can be studied.

Pore water in the intermediate lens is devoid of tritium, but in the top layer.

5.- Groundwater salinity

In the Llobregat area the deep aquifer groundwater, without man's influence, reflects the salinity and chemical composition of the river water before the upstream salt mining and the main industrial processes started. Since the river crosses areas where evaporite salts (mainly sodium chloride and gypsum) outcrop, natural background in the lower Llobregat is about 80 to 120 ppm Cl. This is the chlorinity of the fresh water, now found in the central part of the delta. But, as shown in figure 12, in some areas near the coast, specially in the western area, much higher values are found, resulting from a mixing of this water with old entrapped sea water, in

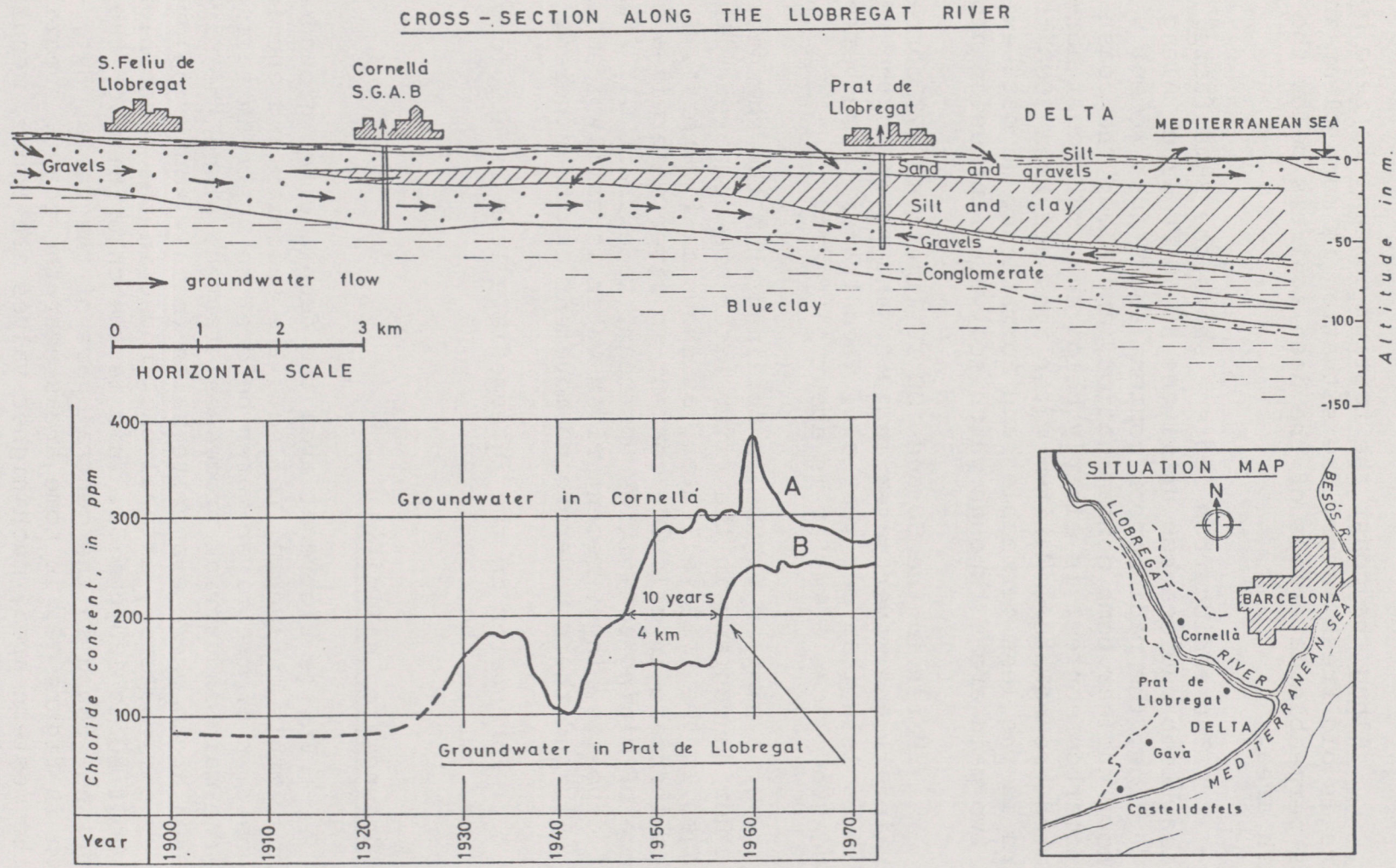

Fig. 11. Movement of industrial chloride contaminated water in the Llobregat area (5).

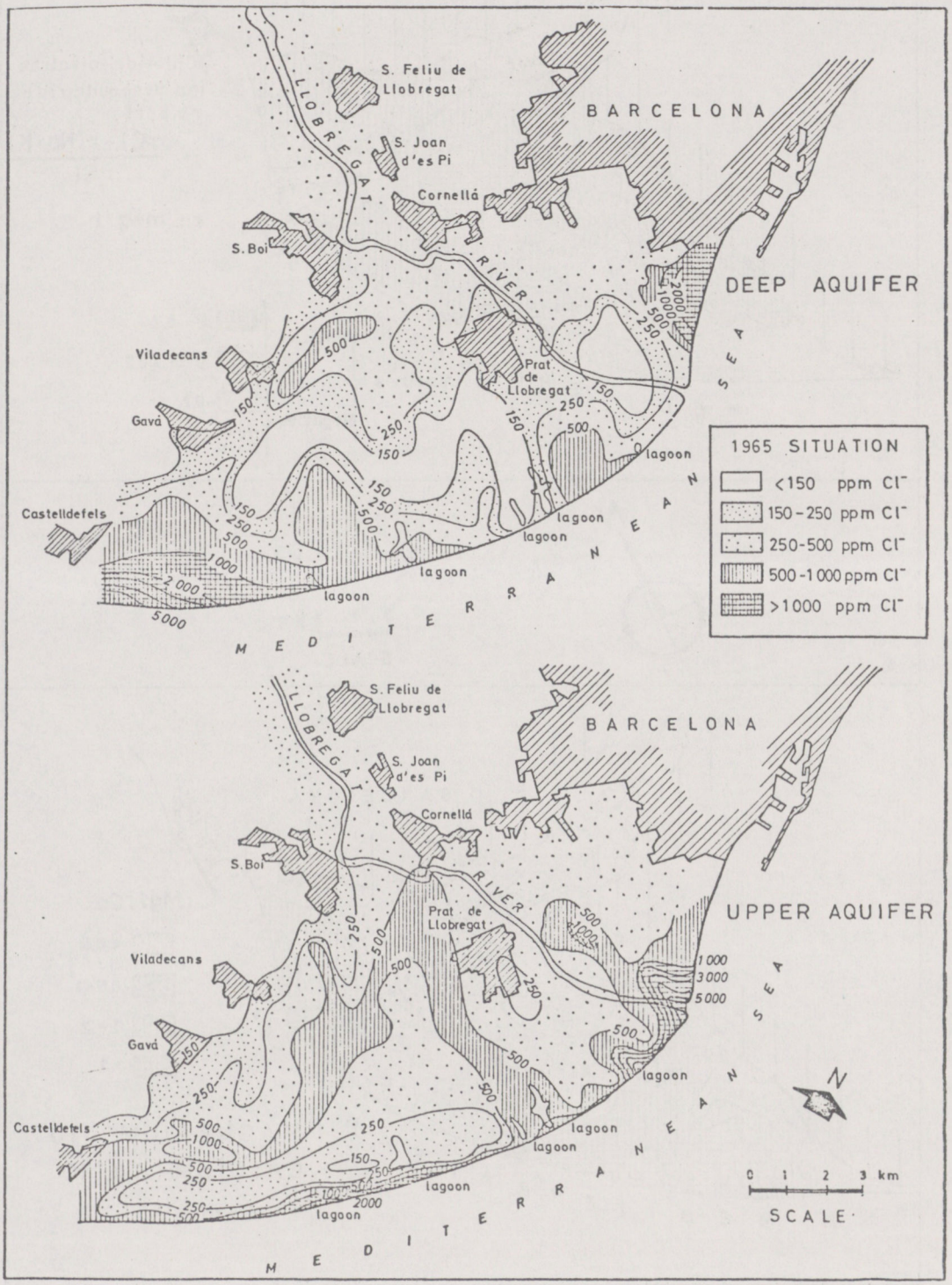

Fig. 12.

Chloride content in ground water in the upper and deep Llobregat delta aquifer, in 1965 (4) (17).

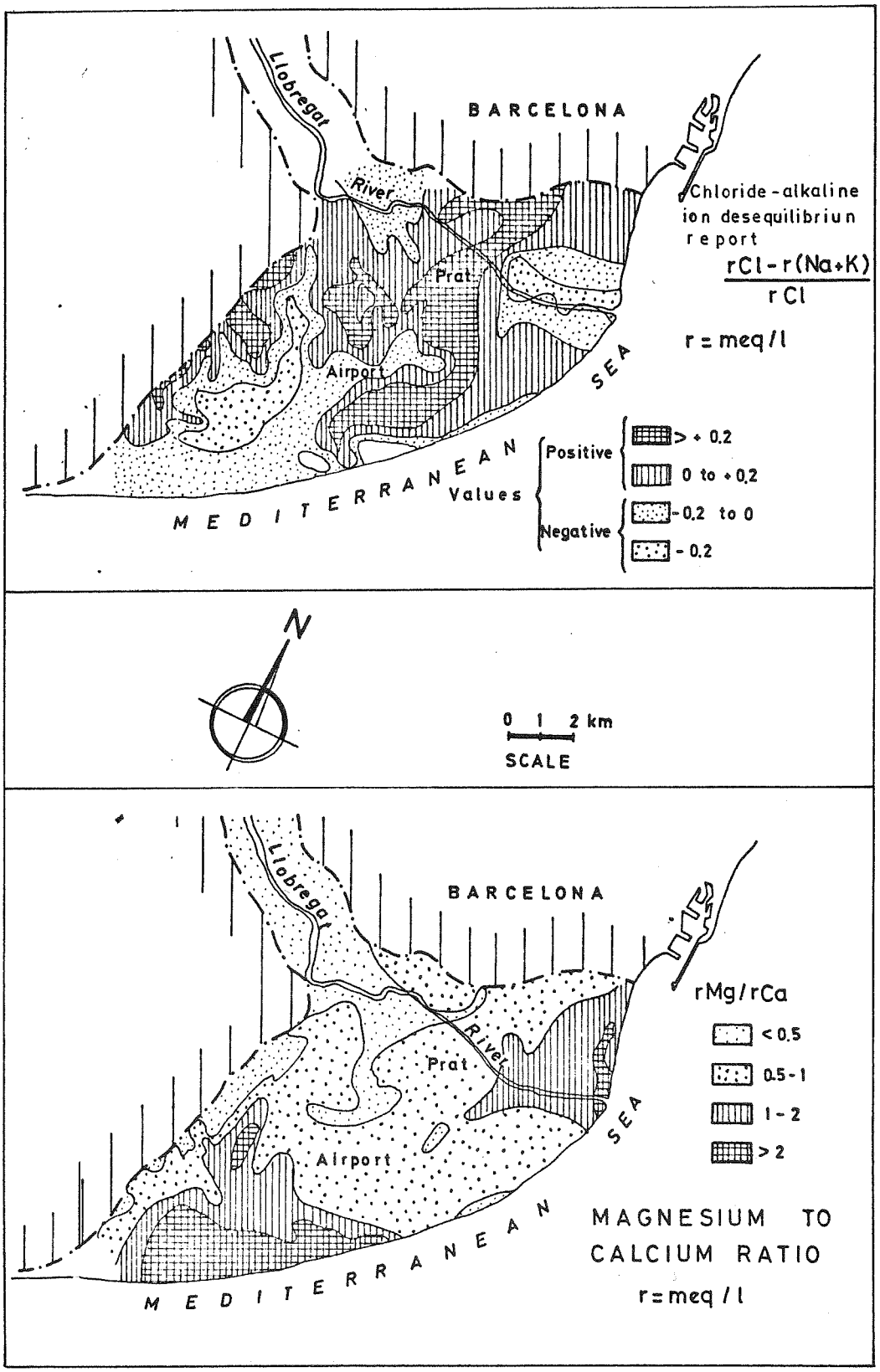

Fig. 13. Geochemical characteristics of the Llobregat delta confined aquifer in 1966. Chloride-alkaline desequilibrium report and magnesium to calcium ratio. (4) (5).

a non-finished process of sea water wash-out. It is clearly shown by the chloride-alkaline ion unbalance and the magnesium to calcium report (fig. 13), since ion exchange occur. When fresh water dilutes existing salty water, an excess of Na develops and Mg increases relative to calcium. The contrary is true when intruding sea water mixes with existing fresh water.

The two other saline zones will be discussed below, because they represent modern sea water-intrusion. Higher values landward represent river related young water or water concentrated in evaporation through irrigation, and reaching the deep upper aquifer leakage.

In the upper aquifer, the chemical composition of the water is variable, since different water sources exist: rainfall infiltration, lateral run-off infiltration, canal water infiltration, irrigation return flows and used water infiltration. Some saline penetrations along the coast represent the effect of coastal lagoons and marshes. The low chlorinity water existing near the coast in the central western part is mainly rainwater infiltrated in the dune belt there existing, floating on saline water.

In the Besós area the situation is similar, but a deep seated sea water encroachment masks most of the chemical aspects. Along the coast the ground salt water is a reducing one, with loss of sulfate and high iron content. It is assumed that sea water get contaminated with organic matter deposited on the sea bottom, when it infiltrates (11) (14).

6.- Sea water encroachment

The present water flow pattern in the central and eastern part of the deep aquifer of the Llobregat delta, as deduced from the piezometric maps and hydrochemical studies, is as shown in figure 14. Water coming from the valley is distributed to the main pumping centers, and some recharge is added from the upper aquifer along the east boundary. At the same time, sea water encroaches through two main areas.

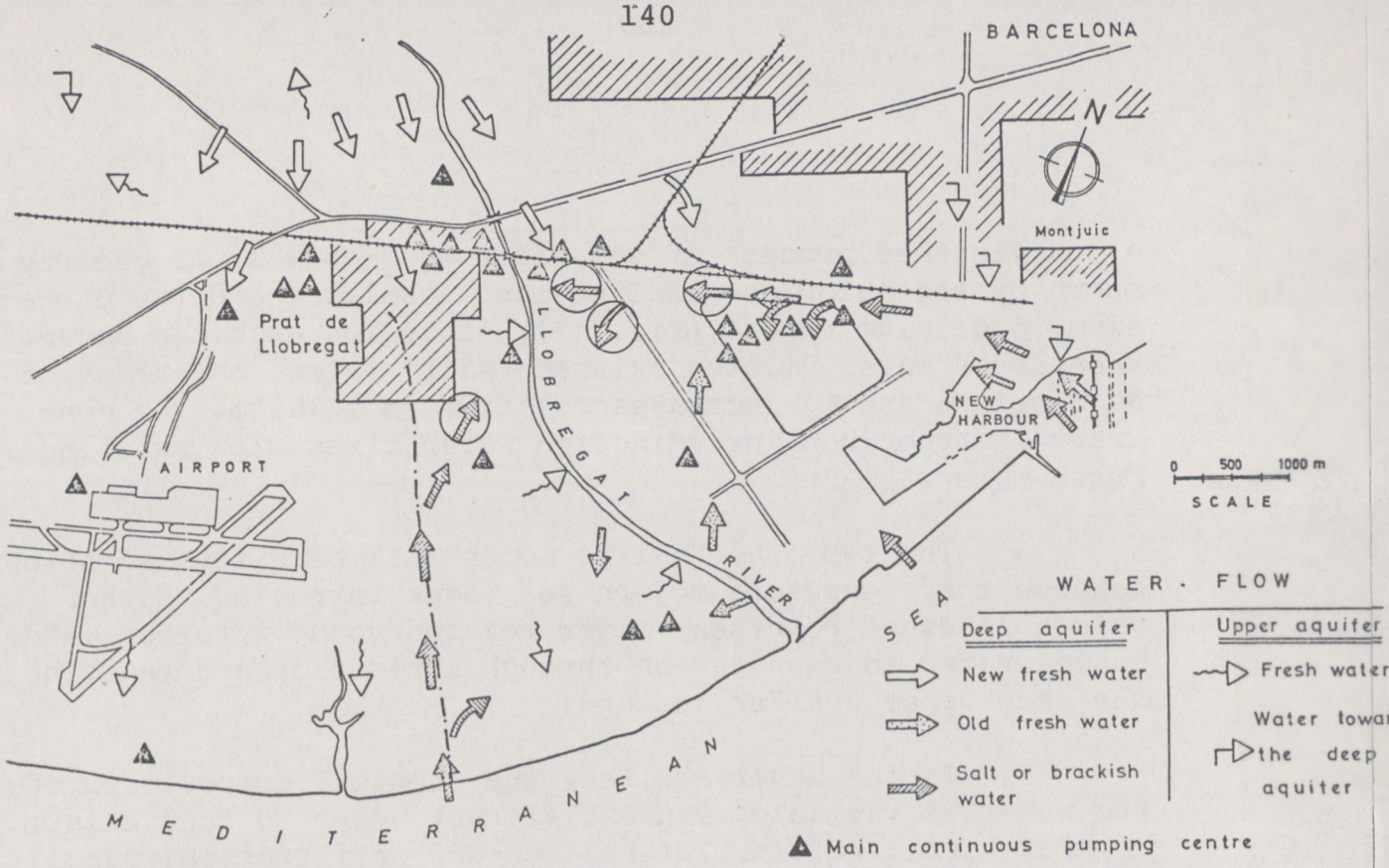

Fig. 14. Groundwater movement as deduced from the piezometric map and the hydrogeochemical studies. Old fresh water movement is very sluggish

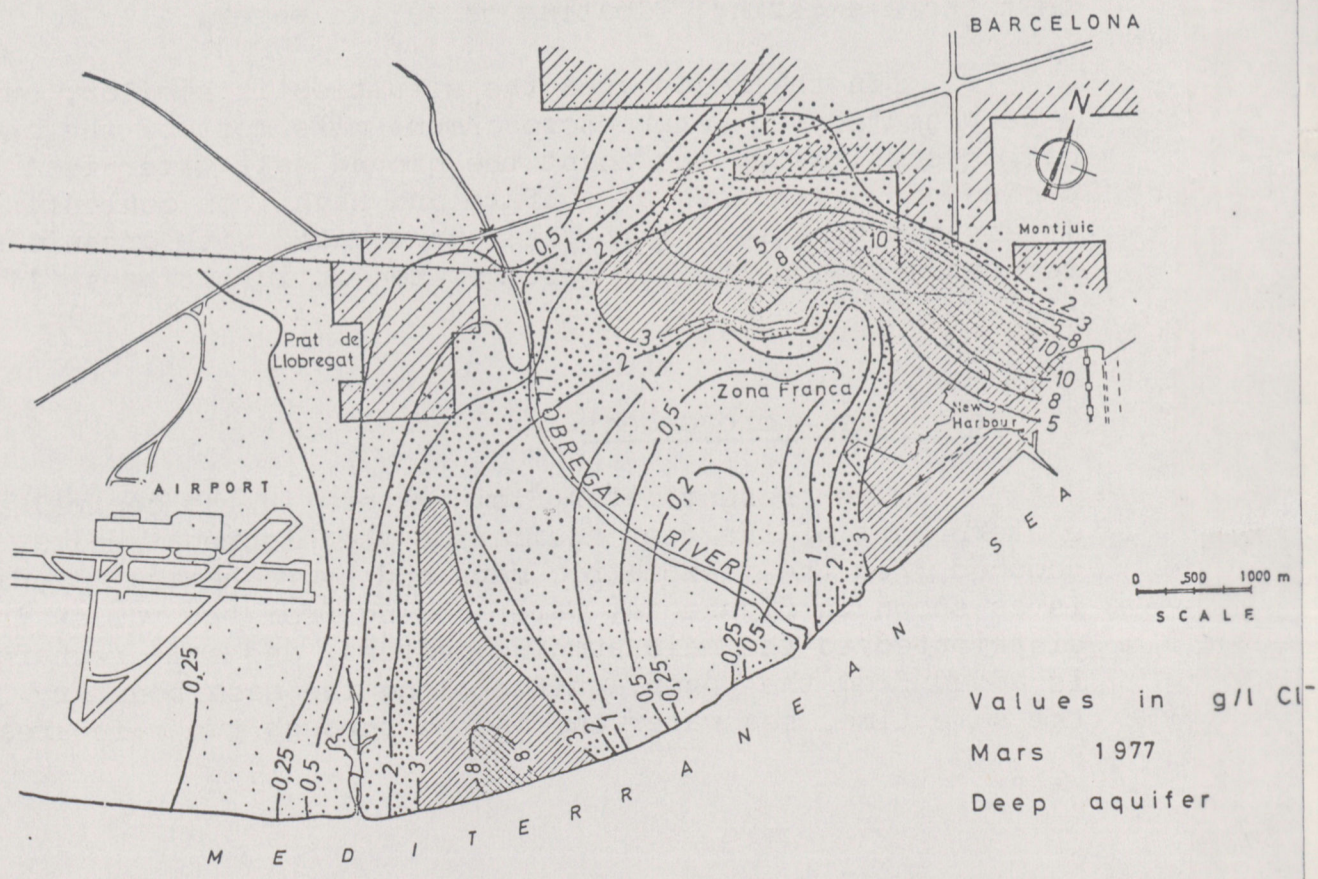

Fig. 15. Chloride-ion content in groundwater in the Llobregat's deep delta aquifer in 1977.

The first one is placed between the river mouth and the airport, coinciding with the old deep river channel (formed under a precedent estuarine situation), which is assumed the main nutcrop of the deep aquifer on the sea floor. This penetration was suggested in the 1965 chloride map of figure 12, and was fully developed in 1977, as shown in figure 15. A mixture of fresh water with some sea water advances through an elongated narrow area devoided of deep-wells, and thus the direct observation is not satisfactory. Dispersion is very high, as expected in such a unsteady process. At present this salty water tongue is quasi-stationary but its apex diffuses slowly toward the main pumping centres.

The second one is placed in the eastern boundary. Salinitation proceeds mainly downward near the coast, since in this area a thick sequence of sand formations prevail, and sea water is only hindered by a scarcely effective muddy sea floor. The flow follows the shorter way to the main industrial wells in the Zona Franca area, and towards the highly permeable deep aquifer lobe extending from the valley mouth until those wells. This sea water is diluted by the inflow from the upper aquifer and by mixing with fresh water. The penetration was clearly iniciated in 1965 (fig. 12) and well developed in 1977 (fig. 15).

Figure 16 reproduces the piezometric level trends in some of the observation bore-holes in the central-eastern area, and shows also the salinity evolution in some of the wells near the main pumping center in the Zona Franca. Well BS is in the path of the sea water encroachment and the salinitation is clearly developed. Wells 15 and 8 show the same trend but less developed, in spite of beeing in the main pumping center; they are placed in the boundary of the above mentioned more permeable lobe and an intense hydrodynamic mixing takes place in them between diluted sea water and river infiltrated water coming from the valley. Well 11 only shows a slight salinization because it is placed outside the lobe, and extracts preferently old water, slowly draining from a deep sandy formation arround the river mouth. Salinization was iniciated in 1965, but was accelerated in 1972 when the groundwater levels went down during the predominantly dry period that followed.

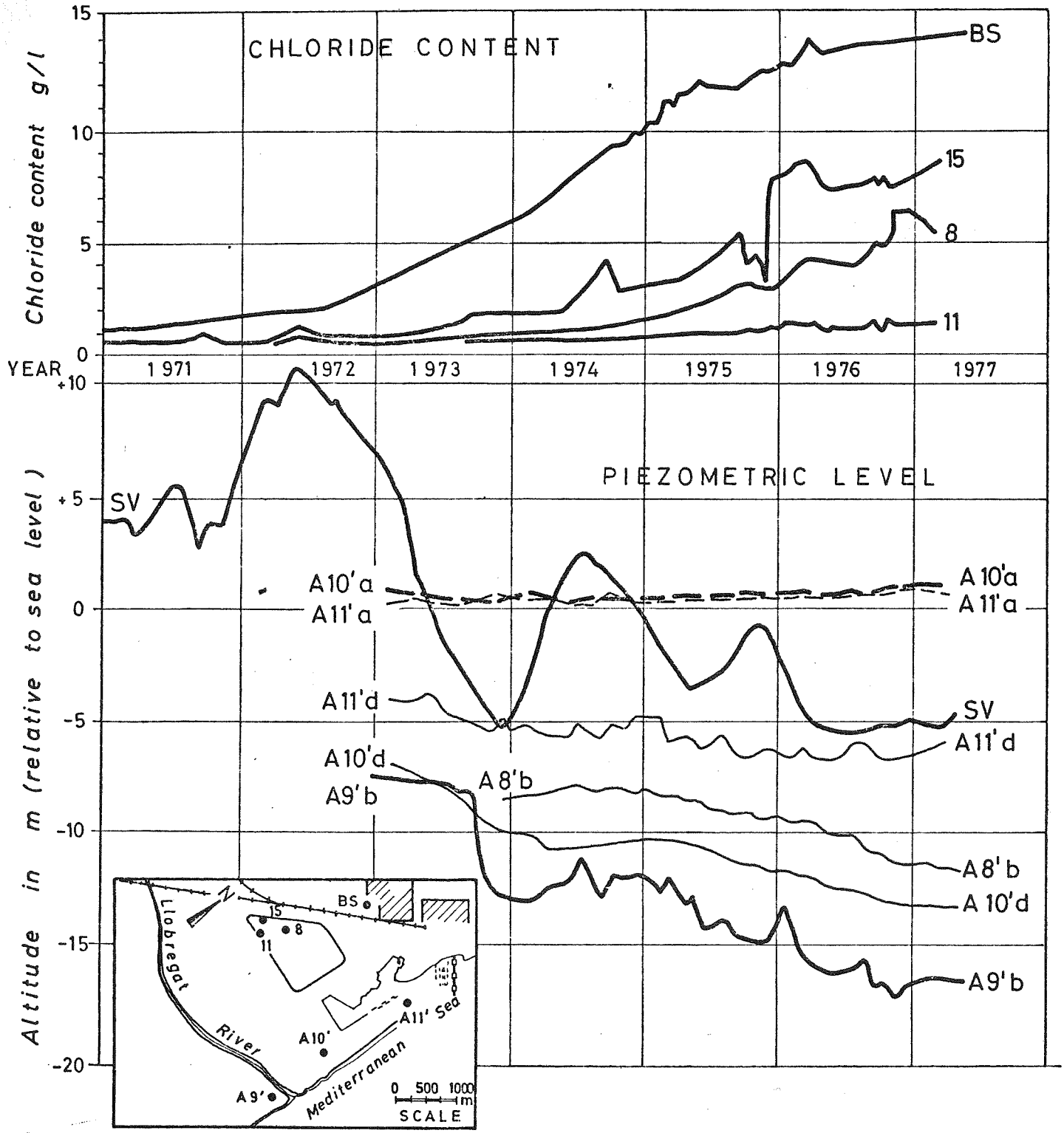

Fig. 16. Trens in ground water piezometric level and chloride content in the Zona Franca (Eastern portion of the Llobregat delta).

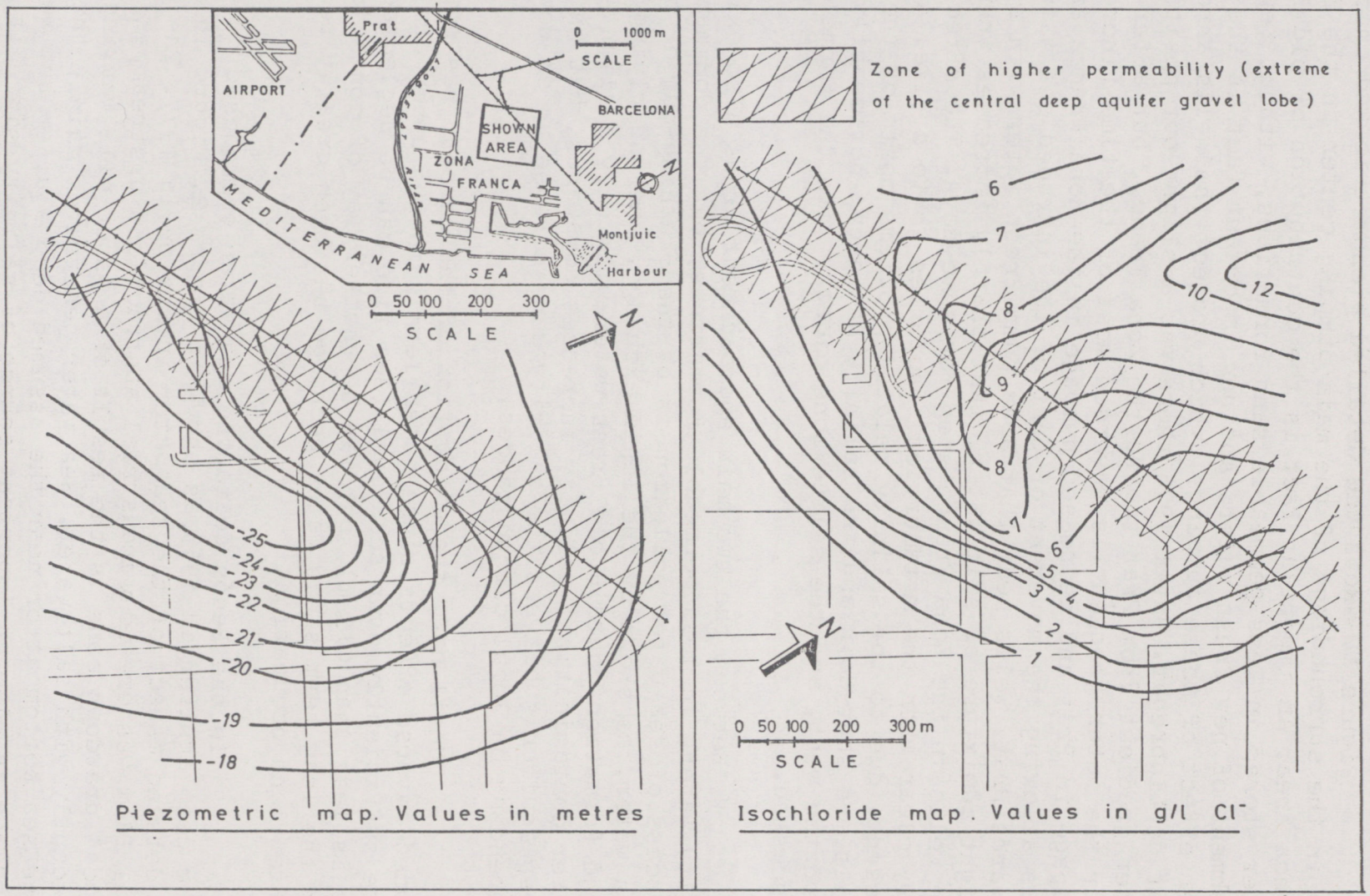

Fig. 17. Detailed piezometric and isochloride maps near the main pumping center in the eastern part of the Llobregat delta.

Figure 17 shows the detailed piezometric and chloride map in the surroundings of the main pumping center in the Zona Franca area. An important fact is the contribution of old fresh water stored in less permeable sand formations, without replenishment of new fresh water. Another important fact is the barrier effect of these pumpings, which protect the wells arround Prat de Llobregat. Until 1976 there was a slight but positive water gradient from Prat toward the Zona Franca, but after that year the water gradient is reversed most of the time (increased pumpage in Prat and decreased pumpage in the Zona Franca area), thus allowing the movement of part of the intruded sea water towards Prat. Wells near that area pump fresh water with a slight admixture of salt water, that sharply increases when the wells in the Zona Franca stop for some days. The effect is very clear when the metallurgical factories go to strike. At present, due to the salinity related problems most of the wells are shut-down and the salinity has grown noticeabily toward the west, and the saline zone connects with the formely discussed.

Between the two main encroachment areas also exist salty water, in less permeable formations, but the dominant process of sea water pollution is by the flushing-out of old sea water, as shown by the ionic exchange. The underlying fine sand formation contain old fresh water with only a slight sea water contamination. Salinity increases downwards, but in the deep aquifer sublayers the salinity can increase up or downwards depending on local circunstances.

Until now, the excavation of the harbour has not produced adverse effects, since water levels in the upper aquifer are sufficiently over sea level (sea tide can be neglected since is less than 0'3 m), but may be another way of contamination in the long term since water recharge has been greatly reduced after land occupation.

In the Besós delta, the sea water encroached deeply the deep aquifer in 1965, as show in figure 2. Figure 18 shows a detailed aspect of the sea water intrusion in 1977. The general flow was created by industrial abstractions upstream, and the local drawdown cone is the result of a paper mill working until recently with salt water. Sea water penetrates mainly through the sea bottom at or near the assumed deep aquifer, but vertical leakage is also an important source of salinity since it contains salt water and there is a non negligeable vertical permeability and also some wells are ill-constructed (slotted or screened in both aquifers, or not cemented between them). The figure 18 shows how an upper aquifer chromium contamination rea-

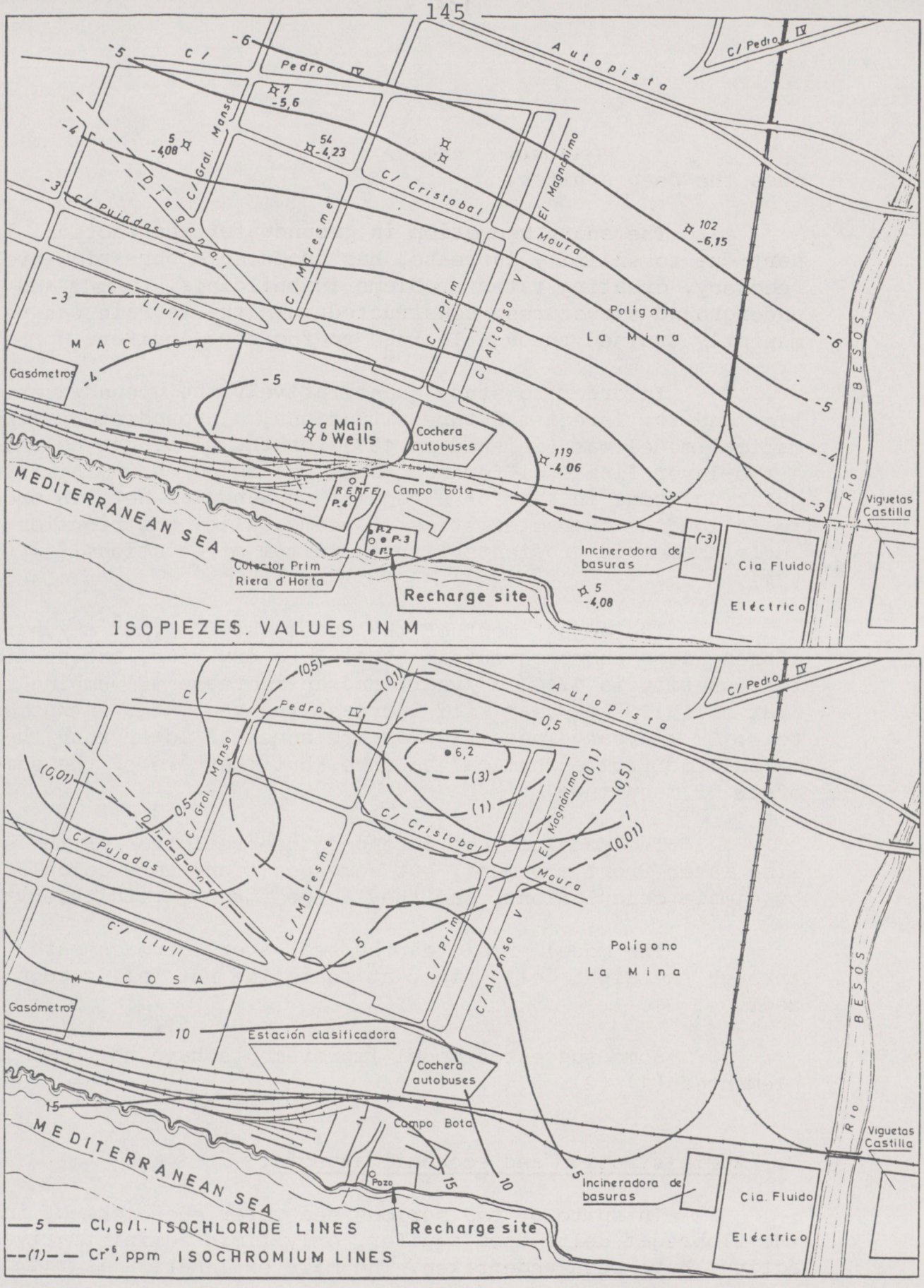

Fig. 18. Sea water intrusion in the besòs delta deep aquifer near an experimental recharge site. A: piezometric map. B: isochloride map showing also an hexavalent chromium contaminacion through vertical leakage. Situation in 1976 (11).

ches the deep aquifer.

The sharp reduction in groundwater abstraction, in part due to salinity increase, has produced a phreatic level recovery, creating water problems in buildings, tunnels and underground structures, constructed when the terrain was permanently drained through leakage to the deep aquifer.

In order to study quantitatively the groundwater flow and to forecast future situations, a groundwater mathematical model was finished in 1971 for the Llobregat delta. A two-layer finite difference model was selected, using irregular polygons in lieu of a regular rectangular net, in order to follow the characteristics of permeability distribution and piezometric gradients with a minimum of equations (3) (9) (21).

The adjustement of the model show $0'5.10^6$ m^3/year of sea water encroachment in 1971. In a dry year, the predicted quantity is 1.10^6 m^3/year, and an increase of pumping in Prat of 15.10^6 m^3/year will increase the sea water intrusion to 4.10^6 m^3/year. These figures are somewhat lower than that calculated from a chemical balance, but they are of the same order of magnitude.

The model do not intended the prediction of the saline water front movement, but some positions were calculated by simple calculations, and they agree roughly with observations.

The model predicted the sea water encroachment through the right delta side, at that time not well recognized.

Some suggested model improvements have not been implemented (1) (2).

7.- Administrative and economic aspects.

In spite of the serious sea water encroachment in the Llobregat delta deep aquifer, only timid administrative actions have been undertaken. The main explanation is the private character of groundwater according to the old Spanish Water Law (first writing in 1866, promulgated in 1878).

The water quality conservation is the matter of the wells owners, and they are protected only by ordinary civil courts after a suit has been presented. The owners are not aware of the problem and, in spite of their seriousness, do not fully understand nor realize the problem.

The Public Administration, through the Eastern Pyrenees Water Authority has carried out some detailed studies since 1965 and operates a dense observation network, but a shortage of manpower and economic resources, and an insufficient legal support, has prevented more complete studies and actions.

Only rough estimations of the direct cost of the salinization problem are available. The figures refer to the substitution of well water for treated river water, when available.

That substitution is bearable for the car and metallurgical factories, but seems excesive for artificial textiles and paper mills with present water use, although recent improvements in water use allows for a substancial cost increase.

But in a few years no enough regulated water in the river will be available and no place exist for a new reservoir to substitude the storage capacity of the salinizated aquifers. This compels to consider the construction of a new transportation water canal from the Ebro river—a 150 km long canal with a 300 m initial elevation, with acute political and administrative problems— or the sewage water reuse.

8.- Possible solutions

It is justified the interest in solving the sea water intrusion problem. Different actions must be taken simultaneously:

1) Reduction of total abstraction. It is not an easy task because the present plant layout impedes an efficient water recycling without major modifications. An 50 % reduction in water demand in previsible. The displacement of the industries to other areas will originate tremendous social problems. Some of the groundwater agricultural demand will be supplied with treated sewage water.

2) Temporarily pumping out salt water in the more effected areas, until other more definitive measures can be applied.

3) Improving river water recharge in the lower valley by surface artificial recharge practices in the river bed. This will be probably the most effective means to combat sea water intrusion through a general increase in groundwater levels. If the water table in the low valley aquifer can be maintained near the river bottom, piezometric levels in the delta deep aquifer will be increased by 10 m, thus reducing present sea water encroachment rate to 1/3, and to at least 1/10 if total abstraction is 50 % reduced. The cost is bearable but the feasibility must be demonstrated.

4) Correcting residual sea water encroachment problems, after previous steps have been completed, by means of some short injection barriers, using if possible excess water from the upper aquifer. Injection barriers are not foreseen at the inicial steps because the efficiency will be low with reasoble water flows, and at a very high cost in water and transport pipes.

From a regional water management point of view, the Llobregat lower valley and delta aquifers must be used in such a form that their regulation and transport capacity would be at maximum. Thus implies that a more important use of river water must be done to supply the industry, giving to the aquifers the role of providing water for human supply and supplying peack water for agriculture and when the river cannot be used.

9.- Acknoledgements

The author is indebted to the Eastern Pyrenees Water Agency, the Prat de Llobregat Municipality and the SEAT factory for permiting the use of the data obtained for them. The mentioned studies were iniciated in 1965 by M.R. Llamas and followed by F. Vilaró as chief of the working team on the Eastern Pyrenees. Some of the data come also from reports made by students of the International Groundwater Course. The exposed aseverations and conclusions do not represent any official statement, but the author's ideas.

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THE DISTRIBUTION OF FRESH AND SALT GROUNDWATER IN THE DUTCH COASTAL AREA AND THE QUATERNARY-GEOLOGICAL EVOLUTION

by J.J. de Vries*

ABSTRACT

Historical sequences of encroaching seawater and recharge of fresh water under different geological and geomorphological conditions during the Quaternary are responsible for a rather complicated distribution of fresh and salt groundwater in the Dutch coastal area. A qualitative genetic model is proposed based on the Holocene landscape evolution.

INTRODUCTION

The complicated spatial distribution of fresh and salt water in the subsurface of The Netherlands in relation to the hydrogeological evolution of the coastal area, has been under discussion for a long time now (e.g. Versluys, 1918; van Dam & Meulenkamp, 1967; Geirnaert, 1973; de Vries, 1974; Meinardi, 1975; ICW, 1976; Engelen, 1981). These authors deal with either a particular area or with a special aspect of the processes involved. Recently more information became available for the area between Den Helder and The Hague through systematic mapping by the Groundwater Survey (TNO), the Geological Survey and the Institute for Land and Water Management Research (ICW).

The present study is based on a combination of an overall view of the fresh-salt water** distribution, derived from the reports of the surveys mentioned above, and the hydrogeological history of the area during the Quaternary. A tentative genetic explanation is proposed for the present groundwater condition, especially based on the Holocene geomorphological evolution of the coastal district. For the Quaternary-geological history of the area, see e.g. de Jong, 1967 and Oele, Schüttenhelm & Wiggers, 1979.

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**In this study salt water is defined as water with a chloride content of more than 500 ppm.

GENERAL SCOPE

From a hydrogeological viewpoint, The Netherlands can be considered as a half-circular groundwater basin, which is continuous to the west and to the north. The basin is mainly filled with coarse Pleistocene fluvial deposits with a base of fine grained Tertiary sediments in the east and clayey Early-Pleistocene in the west. The Tertiary as well as the Early-Pleistocene deposits are of marine origin and contain connate salt water. The thickness of the Pleistocene aquifer increases north-westward and shows an average value of about 200 m in the coastal area. The Pleistocene aquifer is covered in the western part of the country with Holocene clays and peat, reaching a maximum thickness of over 20 m. These deposits were formed in a tidal area with alternating salt, brackish and fresh water character.

During the Quaternary, the western part of The Netherlands has been repeatedly invaded by the sea. Much of the Pleistocene aquifer is therefore filled with saline groundwater. The spatial distribution of fresh and salt water is caused by the historical sequences of encroaching sea water and discharge and recharge of fresh water. A relatively rapid redistribution of salt and fresh groundwater was initiated by the morphological changes that started about 1000 years ago. The lower older dune ridges on top of the coastal barriers were covered from that moment for a part by higher so-called Younger Dunes, reaching local elevations of over 30 m. The shape and location of the initial fresh water lenses below the dune area must have been modified by these morphological changes. About 700 years ago, the creation of the polder landscape started, initially by land drainage and later also by the reclamation of lakes. Due to the artificially maintained groundwater table in the polders and its lower elevation, a more or less continuous flow of groundwater from the higher regions to the lower ones was initiated, causing upward seepage of groundwater from the Pleistocene aquifer into the polder area. The loss of water from the Pleistocene aquifer was partly compensated for by subsurface intrusion of sea water.

Fig. 1 shows the location of former Quaternary coastlines and the present interior salt/fresh water line in the Pleistocene aquifer; table 1 lists the stratigraphy.

period		approx. age
HOLOCENE	x	10^4 yrs BP
PLEISTOCENE Weichselian		7×10^4 " "
Eemian	x	
Saalian		
Holsteinian	x	
Elsterian		
Cromerian	x	
		2.5×10^6 " "
TERTIARY		

Table 1 Schematic stratigraphy for the Dutch coastal area

x = marine influence

THE SPATIAL DISTRIBUTION OF FRESH AND SALT GROUNDWATER

General processes

- Saline groundwater is found in the Early-Pleistocene and Tertiary marine deposits of the lower boundary of the Pleistocene aquifer. The maximum depth of this lower salt/fresh interface is reached below the ice-pushed ridges in the central part of the country and in a large graben structure in the south. Deeply infiltrated rainwater in the high ridges (with an altitude up to 100 m), and a great thickness of fluvial deposits filling the graben may explain this distribution. The easternmost penetration of salt water in the Pleistocene aquifer is found in the large southeast-northwest oriented valleys of the rivers Rhine, Eem, and Hunze. The salt water tongues most probably are the result of dragging up of salt water from the marine deposits at greater depth by the upward flow of groundwater below these valleys (fig. 1). Other salt intrusions from the lower marine deposits into the Pleistocene aquifer are connected with : (i) Upward diffusion of salt into a stagnant groundwater body and dispersion through horizontal flow of groundwater at greater depth. (ii) Upwelling of salt water below deep polders through upward seepage.

- Invasion of the sea took place mainly in four periods after the Early-Pleistocene : In the Cromerian, in the Holsteinian, in the Eemian and in the Holocene (fig. 1 and table 1). The intruded salt from the Cromer-, the Holstein- and the Eem sea in general was removed from the coarse fluvial deposits by flushing during the subsequent continental period of the Weichselian, which lasted some 70,000 years. The presence of fresh water directly below marine Eemian clay in most areas supports this idea. Exceptions occur in the northern part of Friesland and near Amsterdam, where saline water from respectively the Cromerian/Holsteinian and Eemian are found in sandy lenses with clayey glacial deposits. Most probably a zone of brackish water with an eastward dipping interface was present in the lower part of the aquifer at the end of the Pleistocene, as a result of diffusion and dispersion of salt from below (Meinardi, 1975).

- The Holocene sequence started with the development of a thick peat layer on the Pleistocene surface, in front of the invading sea. Subsequently, the peat was covered by lagoonal clayey deposits. These less-pervious layers initially protected the Pleistocene aquifer from salt water intrusion by density currents. However, the transgression of the sea was more aggressive during the younger Holocene. Deep tidal inlets came into being, cutting partly into the Pleistocene subsurface. This caused a relatively high chlorite content in the larger part of the coastal area in both Holocene and Pleistocene deposits. But fresh water lenses are still found below clay layers of Early-Holocene, Eemian and Pleistocene-glacial origin. Furthermore, extensive fresh water lenses have been developed below the coastal dunes.

Regional conditions

- In the area of Westfriesland (north of Amsterdam, fig. 2 and 3), a large fresh water lens occurs in the upper Pleistocene layers, east of the brackish zone behind the dunes. Near the northern and western limits of this lens (Wieringermeer and Geestmerambacht, freshwater has only been preserved below confining layers layers, creating an inverse salinity distribution. The Westfrisian fresh water occurrence can be explained by the Holocene history of the area. The top of the Holocene sediments, which is slightly below mean sea level at present, was formed about 3000 years ago at an elevation of approximately 1.5 m above mean sea level at that time. Subsequently the area was covered by a peat layer, which has disappeared since, so that Westfriesland has been some metres above sea level during more than 1000 years. The clayey tidal flat deposits have been dissected in this area by many sand-filled channels, some of which exceed a width of 1000 m and penetrate deeply into the underlaying Pleistocene (fig. 4). The combination

of a relatively permeable soil and a high elevation may have caused a deep infiltration of rainwater and the forming of a fresh water lens , analogous to those below the coastal dunes. Evidently, this fresh water lens is not in hydraulic equilibrium with the present sea level, and will therefore have been submitted to an upward pressure from the surrounding salt water. However, a number of deep polders have been created in the area around Westfriesland, especially in this century through the reclamation of parts of the former Zuyder Sea. As a result the piezometric level in the Pleistocene aquifer has been lowered to more than 2 m below m.s.l., restoring the old equilibrium.

-East of Westfriesland, the former Zuyder Sea area shows neither strong Holocene salt intrusion, nor fresh water infiltration. The chloride content in the Pleistocene deposits is regularly increasing with depth, indicating a predominance of diffusion from below (fig. 3 and 5). Before the enclosure of the Zuyder Sea in 1932, the chloride content in the Holocene deposits decreased from about 6000 ppm at the top, to less than 1000 ppm at a depth of about 12 m. This vertical chloride distribution was quantitatively explained by Volker (1961) by downward salt diffusion from the Zuyder Sea, which used to have a chloride content of about 6000 ppm. This process started in Medieval times when the Zuyder Sea gradually came into being by destructive invasion of the sea through a tidal inlet into this former peat area. Evidently the Holocene clayey deposits prevented the salt water intrusion into the Pleistocene by density currents, for there were no deeply incised tidal gullies, opening up the aquifer from above.

- The morphology of the area between Amsterdam and The Hague is dominated by a central zone with deep polders, boundered by a coastal dune belt in the west and the Utrecht ice-pushed ridge in the east. Fresh groundwater from these ridges and salt groundwater from greater depth, discharge into the polder area as upward seepage. Furthermore, salt water intrusion is observed along the northern and southern limits of the area, where Young-Holocene tidal inlets and estuaria dissected the coastal barriers. Local groundwater flow systems connected with the polder topography are superimposed on the regional systems (fig. 2).

CONCLUSION

A reconstruction of the Quaternary geological and geomorphological history is the best approach towards understanding of the present spatial distribution of fresh and salt water in the subsurface of a coastal area. A geological survey will however yield only qualitative models of a hypothetic nature, which need to be tested subsequently by hydraulic and hydrochemical methods. Quantitative deductions ask for rather accurate morphological reconstructions, necessary to arrive at proper boundary conditions. Until now this state has not been reached in The Netherlands, in spite of recent very intensive Quaternary-geological research and groundwater survey. A combined effort is needed; let this be a first trial.

Acknowledgements

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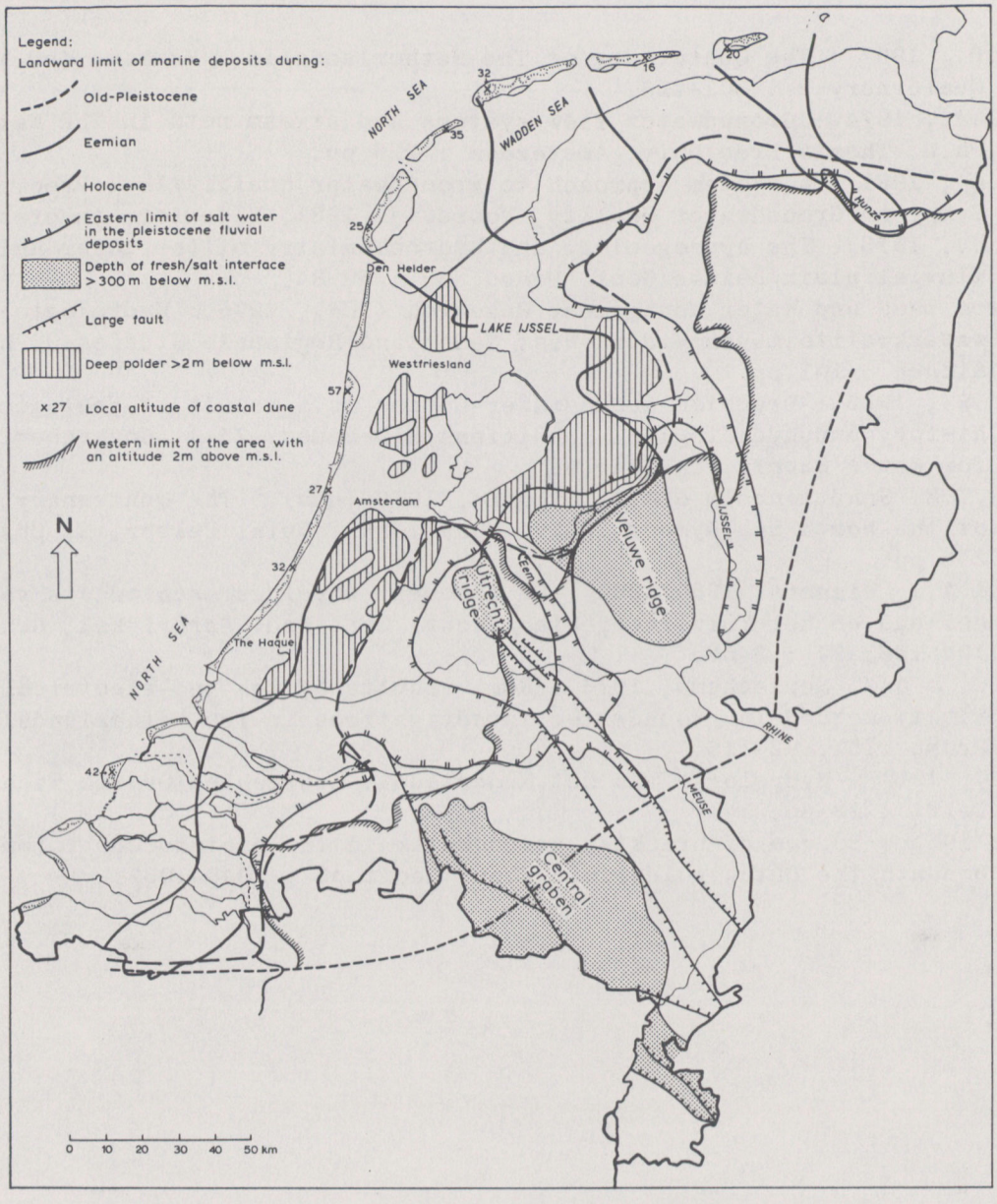

Fig. 1

Large scale topographic and geologic structures, former coastlines and the fresh-salt water interface at greater depth.

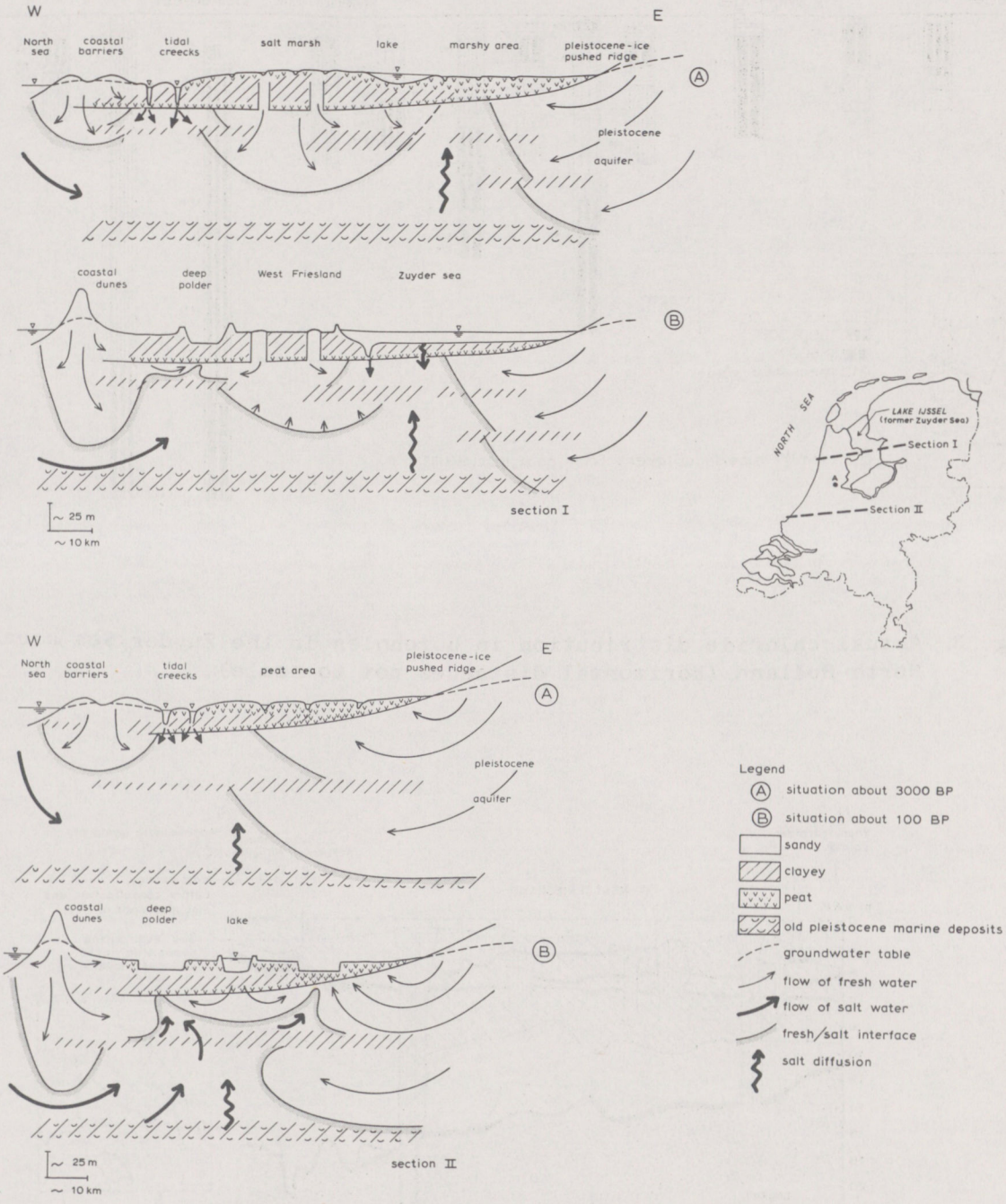

Fig. 2 Schematic hydrogeological sections through the Dutch coastal area, showing the movement of fresh and salt water in connection with the geomorphological conditions about 3000 years and 100 years before present.

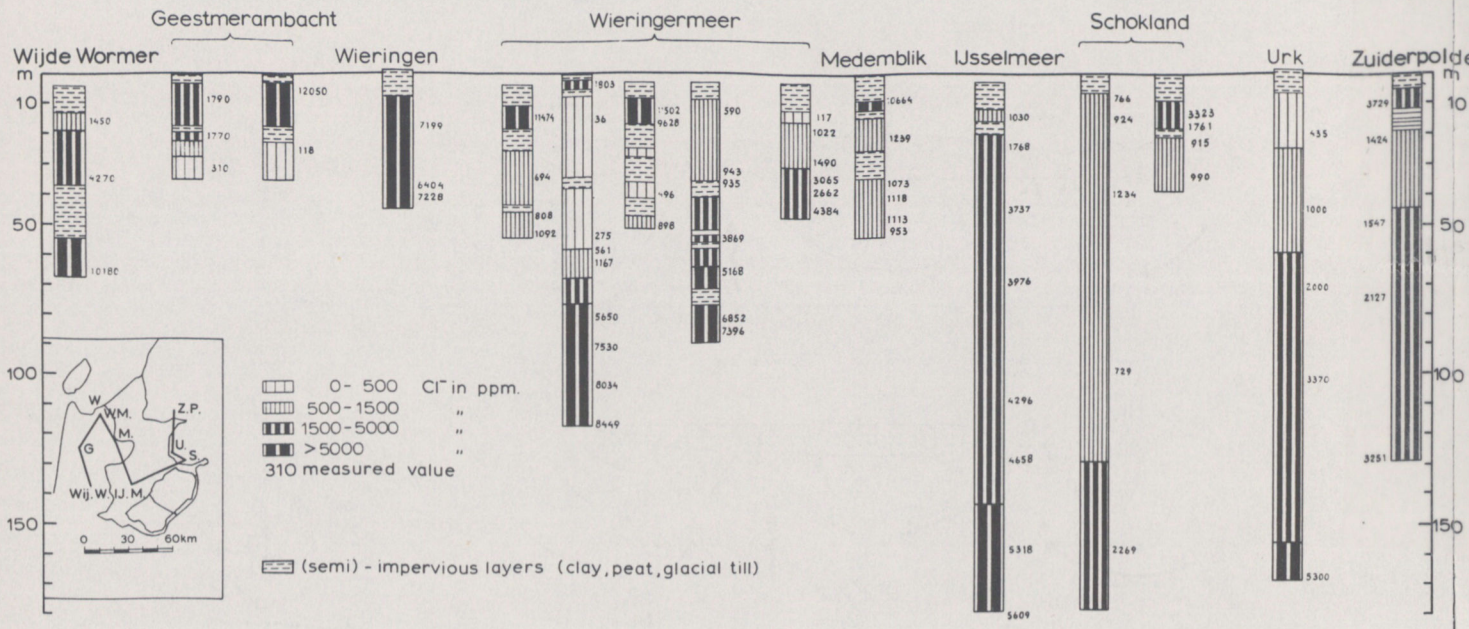

Fig. 3 Actual chloride distribution in boreholes in the Zuyder Sea area and North Holland (horizontal distances not to scale).

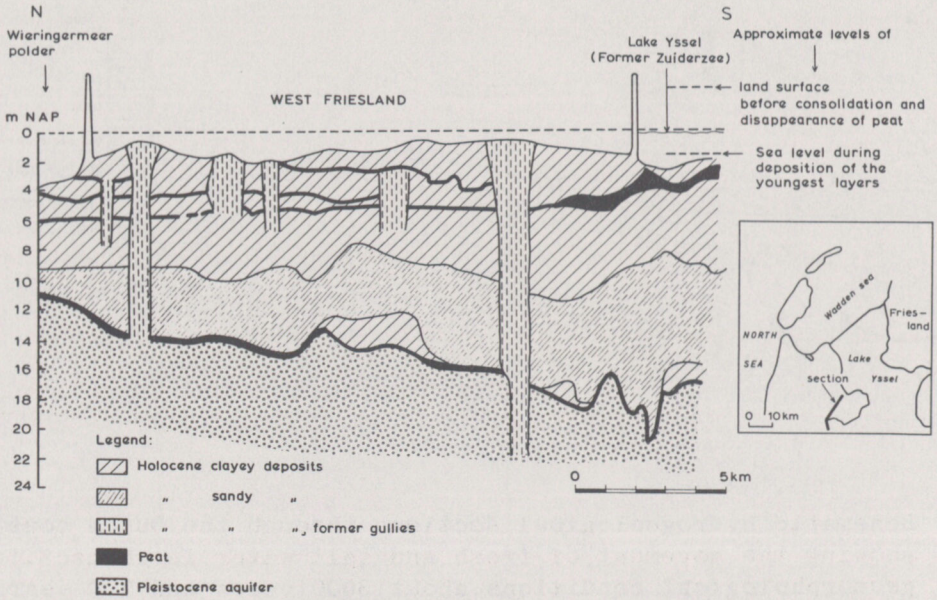

Fig. 4 Geological section showing wide and deep tidal inlets buried in the Holocene deposits of Westfriesland, and the land surface relative to former sea level (after Pons & Wiggers, 1959/1960).

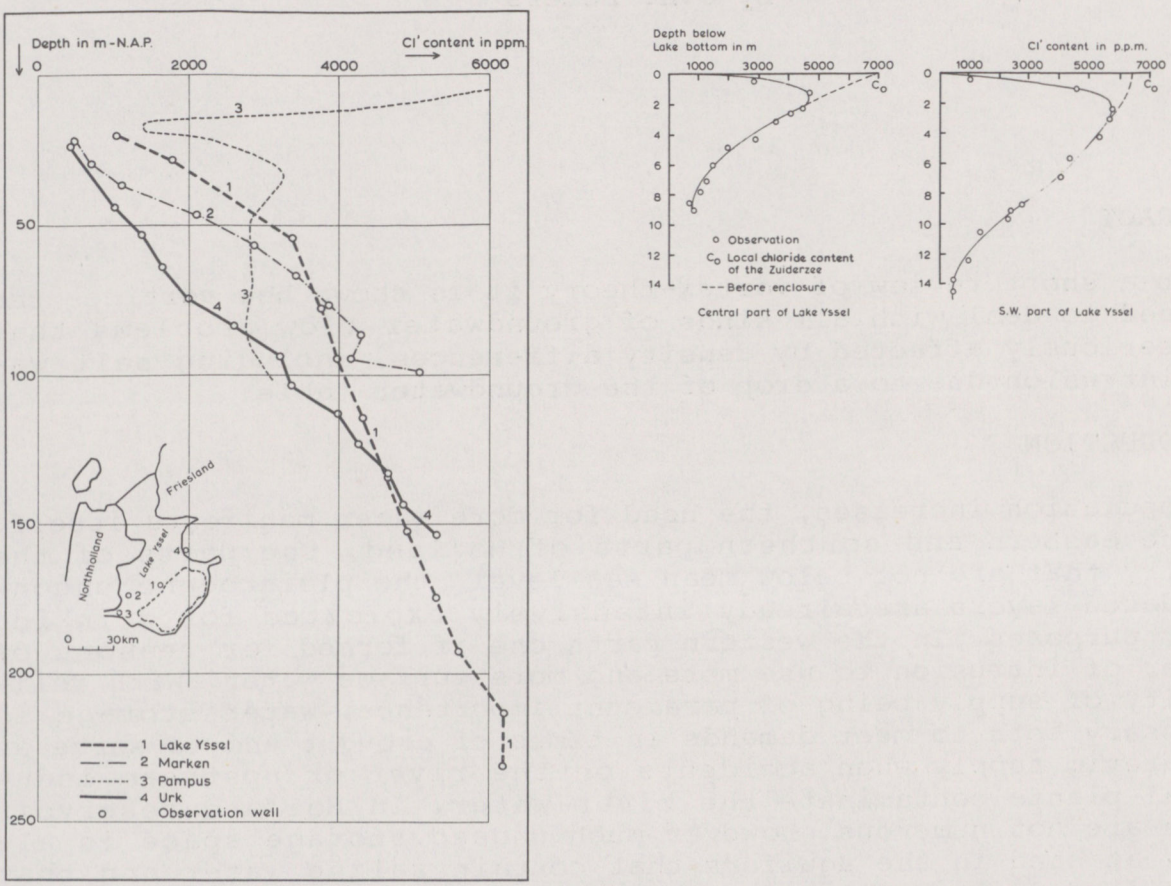

Fig. 5 Vertical chloride distribution in the subsurface of the former Zuyder Sea area (after Volker, 1961).

APPLICATION OF VORTEX DISTRIBUTIONS IN MODELLING THE STORAGE OF FRESH WATER IN SALINE AQUIFERS

by J.H. Peters[1]

ABSTRACT

After a short review of vortex-theory it is shown how vortices can be used to deal with all kinds of groundwater flow problems that are seriously affected by density differences, including salt water intrusion due to a drop of the groundwater table.

INTRODUCTION

As population increases, the need for more water manifests itself. In the eastern and southern parts of Holland, the parts of the country that are not below mean sea level, the pleistocene unconsolidated layers are already intensively exploited for drinking water purposes. In the western parts one if forced for reasons of danger of intrusion to use more and more surface water. With reliability of supply being of paramount importance, water storage is necessary both to meet demands in times of drought and to serve as an interim supply when accidents on the river or upstream industrial plants contaminate the river water. In Holland reservoir sites are not numerous. However much unused storage space is already at hand in the aquifers that contain saline water and that underlie densely populated as well as isolated areas that can be found along the coast. Once retrieved the water has good and fairly constant physical and chemical characteristics.
In case the aquifer is confined by less pervious deposits of great thickness, recharge can only be accomplished by injecting the water directly into the aquifer using wells. Various experiments have shown that severe clogging of those wells can be overcome by a sufficient pretreatment and a regular cleaning (Olsthoorn, 1981). Saline aquifers now are considered a natural resource because they are potential storage space for fresh water.

VARIABLE DENSITY FLOW

Considerable progress has been made recently in the analysis of fluid flow through porous media for those situations in which the flow is significantly affected by variations in the density of the fluid. In such problems an analysis in which small viscosity differences are ignored can be justified (de Josselin de Jong, 1960 and Verruijt, 1980).

When dealing with variable density flow the equation of motion is Darcy's Law, in terms of pressure. In the general three dimensional case it is written as follows:

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$$q_x = -\frac{K}{\mu}\frac{\partial p}{\partial x}$$
$$q_y = -\frac{K}{\mu}\frac{\partial p}{\partial y} \qquad (1)$$
$$q_z = -\frac{K}{\mu}\left(\frac{\partial p}{\partial z} + \varrho g\right)$$

where q_x, q_y and q_z are the components of the specific discharge vector $\vec{q}$ (*), ϱ denotes density and p is the fluid pressure. Gravity $\vec{g}$ is working in negative z direction, g is acceleration of gravity, K is intrinsic permeability and μ dynamic viscosity.

It was first recognised by de Josselin de Jong (1960) that density differences create rotation in the flow. This rotation is obvious when we take the curl of $\vec{q}$ (Wylie, 1975), using (1)

$$\nabla \times \vec{q} = \vec{i}\left(\frac{\partial q_z}{\partial y} - \frac{\partial q_y}{\partial z}\right) - \vec{j}\left(\frac{\partial q_z}{\partial x} - \frac{\partial q_x}{\partial z}\right) + \vec{k}\left(\frac{\partial q_y}{\partial x} - \frac{\partial q_x}{\partial y}\right) \qquad (2)$$

We define $\vec{i}$, $\vec{j}$ and $\vec{k}$ to be vectors of unit length directed, respectively, along positive x, y and z axes of the right-handed rectangular coordinate system.

Because the fluid pressure is a physical quantity it must be single valued. So (2) reduces to

$$\nabla \times \vec{q} = -\frac{Kg}{\mu}\left(\vec{i}\frac{\partial \varrho}{\partial y} - \vec{j}\frac{\partial \varrho}{\partial x}\right) \qquad (3)$$

Hence, when using the definition of a cross product of vectors (Wylie, 1975), we have

$$\nabla \times \vec{q} = \frac{Kg}{\mu}\left(\vec{k} \times \vec{\nabla\varrho}\right) = \frac{K}{\mu}\left(\vec{\nabla\varrho} \times \vec{g}\right) \qquad (4)$$

So the rotation or curl of $\vec{q}$ ($\nabla \times \vec{q}$) is a vector whose absolute value is the product of K/μ, the absolute values g and $|\vec{\nabla\varrho}|$ (the density gradient) and the sine of the angle between $\vec{g}$ and $\vec{\nabla\varrho}$ and whose direction is perpendicular to the plane determined by $\vec{g}$ and $\vec{\nabla\varrho}$ and so sensed that a righthanded screw turned from $\vec{\nabla\varrho}$ to $\vec{g}$ through the smaller of the angles would advance in the direction of $\nabla \times \vec{q}$, see figure 1 (**).
It shows that the rotation of $\vec{q}$ is a horizontal vector so senced that it rotates the interface towards a horizontal position when the lower liquid is the denser one (de Josselin de Jong, 1969).

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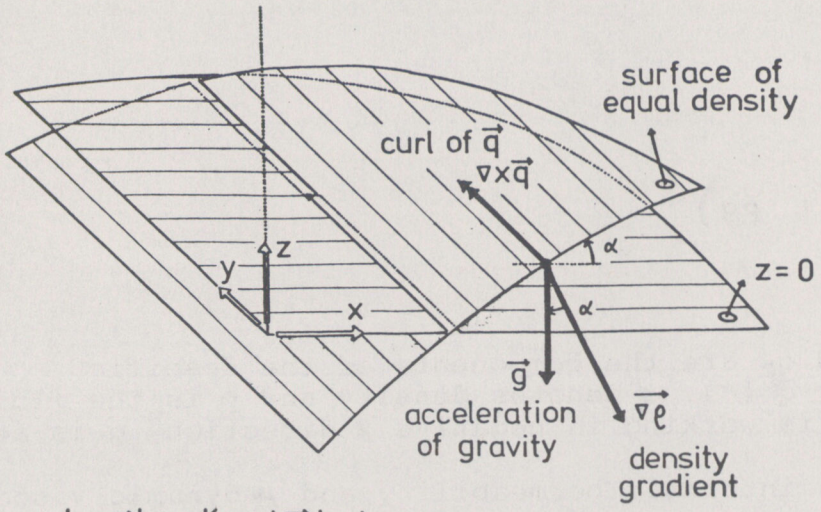
* A vector quantity is indicated by putting an arrow above the symbol.

** In all figures a vector will be drawn as an arrow. If it denotes rotation it will have a double point.



$$|\nabla \times \vec{q}| = \frac{K}{\mu} g \cdot |\nabla \vec{\rho}| \cdot \sin \alpha$$

Figure 1. The horizontal component of the density gradient creates rotation.

VORTEX-THEORY

A vortex can best be described as the total amount of rotation within a small element of soil. So a vortex possesses both magnitude (total rotation or vortex-strength) and direction and can be considered a discretization or singularity of rotation. The concept of replacing all fluids with different densities by one hypothetical fluid and then introducing singularities at those points where the densities of the actual fluids change is first mentioned by de Josselin de Jong (1960). The singularities generate the effects of varying density. By applying the principle of superposition the resulting flow can be computed. It consists of two parts. One of them is accounting for the effects of density differences, the other for the flow of the hypothetical fluid. The components of flow due to a vortex can be calculated, see next figure (*).

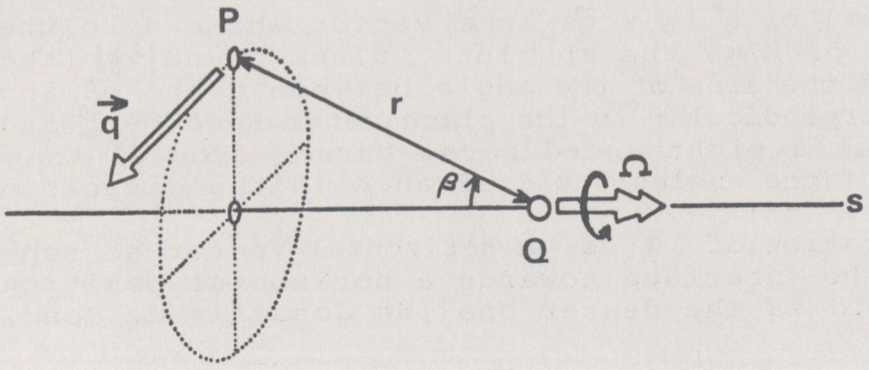


Figure 2. Flow due to a vortex.

(*) A vortex will be drawn as a white arrow with a double point.

The absolute value $|\vec{q}|$ at P due to a vortex with strength Ω at Q in the direction s equals (Lamb, 1932)

$$|\vec{q}| = \frac{\Omega}{4\pi} \frac{\sin\beta}{r^2}$$

$\vec{q}$ is perpendicular to the plane containing the direction of the vortex-line s at Q and the line r and its sense is that in which P would advance if it were attached to a rigid body rotating with the fluid element at Q. The components of flow due to all vortices can be calculated again by using the principle of superposition. Of course to get real velocities the specific discharge should be divided by the porosity with respect to flow.
Vortex distributions were first implemented in computerprograms for non-steady two-dimensional groundwater flow by Haitjema (1977 a,b).
Later he extended the vortex theory to axial symmetric flow (1980), other (three dimensional) programs are being developed (Luger, 1980). In the following paragraphs it will be shown how the axial symmetric model can be used to compute the amount of storage of fresh water that can be accomplished in aquifers containing saline water and to model the transient behaviour of the fresh water- salt water interface in polder areas due to a sudden lowering of the groundwater table.

RESULT OF MODELING STORAGE

Figure 3 shows a cross section through an essentially homogeneous aquifer which has negligible existing groundwater movement. It will be shown that the storage of fresh water in this formation is feasible.

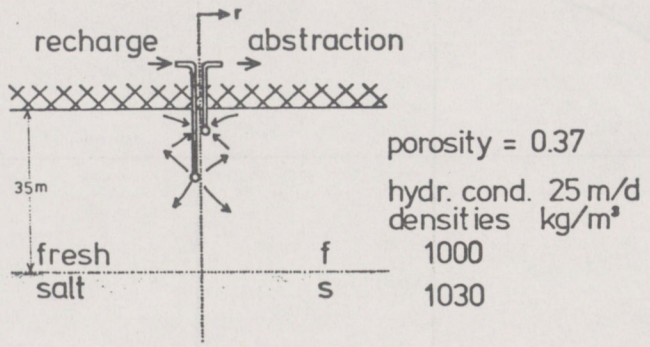


Figure 3. An aquifer containing salt water

The formation conforms to the usual assumptions of groundwater engineering. At the boundary between the regions of different density the pressure for the miscible fluids is continuous. Replacement of the fluids in the pores is complete and the aquifer has got no secondary permeability and porosity.
The volume of storage needed to tide over a period without recharge is difficult to calculate. Let us assume the intake of surface water must be interrupted for one month at regular intervals. The stock of fresh water must be built up again in one month after which the recharge stops again for one month etc. as is shown in the next diagram.

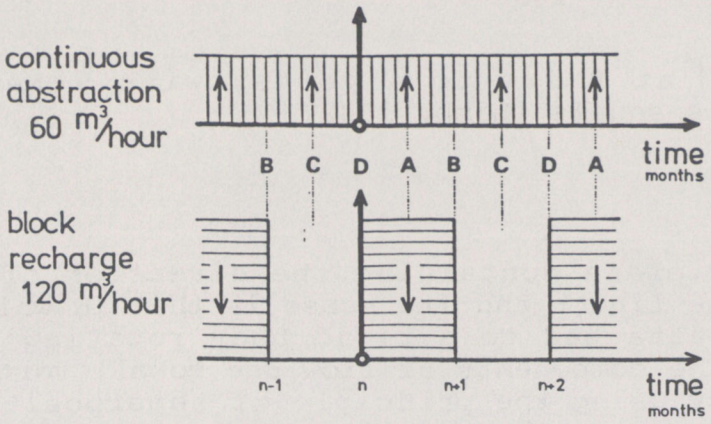


Figure 4. Abstraction-recharge scheme

On the long term total abstraction equals total recharge. In figure 5 the transient subsequent stages of the interface at times A, B, C and D and its initial position are shown.

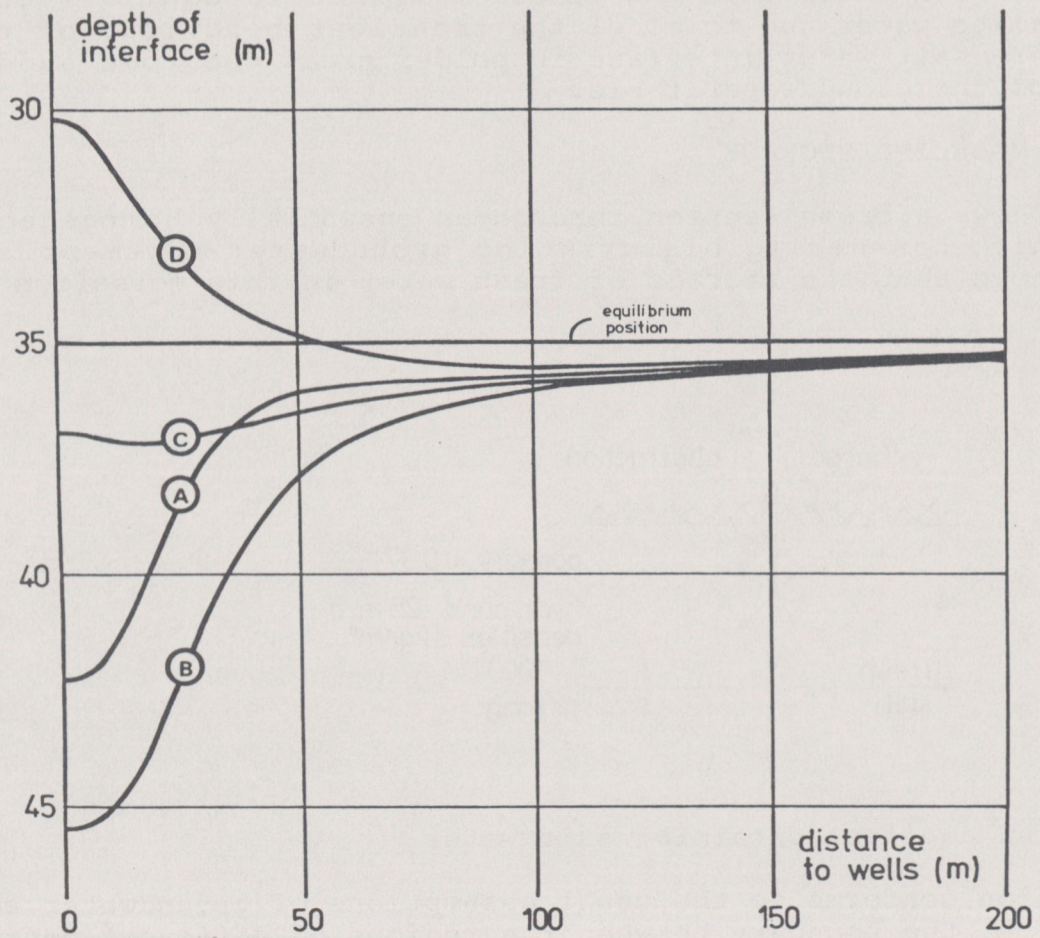


Figure 5. Initial and subsequent stages of interface

Mixing due to the fact that a part of the fluid is not displaced as readily as the rest will cause the growth of a gradual transition zone. This will be the case when there are dead end pores (into and out of which the solutes only move by diffusion) and less permeable parts or when a film of the displaced fluid is left behind on the sand grains. Adverse effects of the transition zone

can be overcome by injecting more water than will be abstracted, so that the brackish water will be flushed out. The mixing process can be reduced by preventing the interface to come up again when recharge must be interrupted. This can be pursued by positioning the abstraction wells in another aquifer at shallow depth or by abstracting salt water with other wells. Of course this is only possible if you can get rid of the salt water by injecting it at great depth, carrying it to the sea or desalting it by reverse osmosis.
Macroscopic mixing due to the randomness of streamlines is under study now at the Municipal Waterworks of Amsterdam with a recharge well positioned within a completely saline aquifer. In this and other field experiments electric conductivity and temperature are used as tracers.

RESULTS OF MODELING SALT WATER INTRUSION

The lower parts of Holland are divided in numerous separate polders with controlled shallow water tables. For reasons of agriculture (to get maximum crop) these water tables are lowered at regular intervals. This causes a continuous flow of groundwater to such a polder, the so-called seepage. This seepage is accompanied by a load of salt resulting in high chloride contents in the drainage system and the open water. For use in agriculture the chloride content of this water should not exceed certain values. Just to show the capabilities of a vortex computer program some results of modeling this form of salt water intrusion are shown in this paragraph.

Figure 6 shows a cross section of a geohydrologic profile of infinite extent. The formation has got negligible existing groundwater flow.

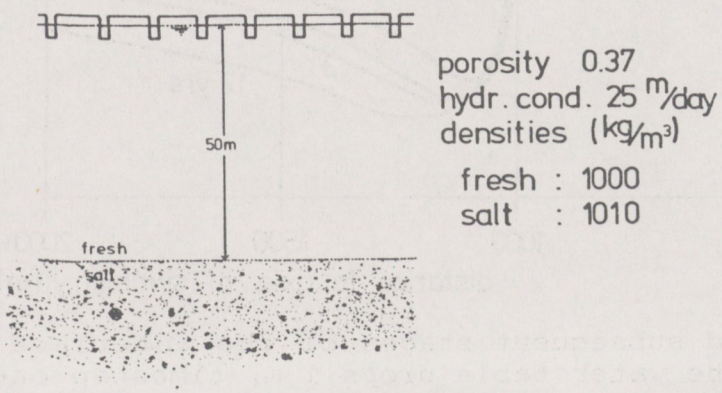


Figure 6. Geohydrologic profile of an aquifer containing salt water.

Because the groundwater table can be considered a special case of an interface (with the upper fluid being weightless) at the edge of a circular polder a vortex ring is created when the water table is lowered, figure 7.

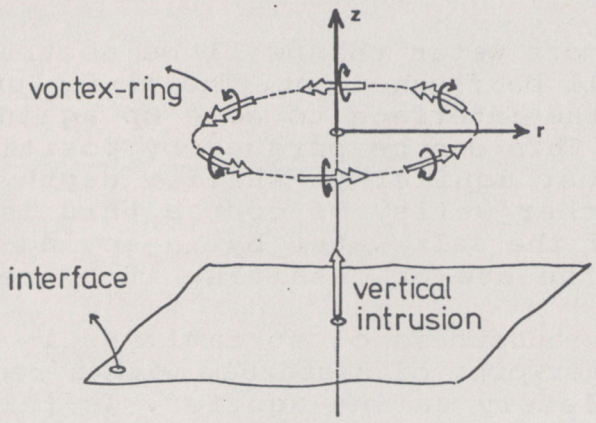


Figure 7. A lowering of the water table within a circular polder creates a vortex ring.

The components of the specific discharge (and actual velocity) due to the vortex ring can be implemented in the computerprogram. Subsequent stages can now be calculated.

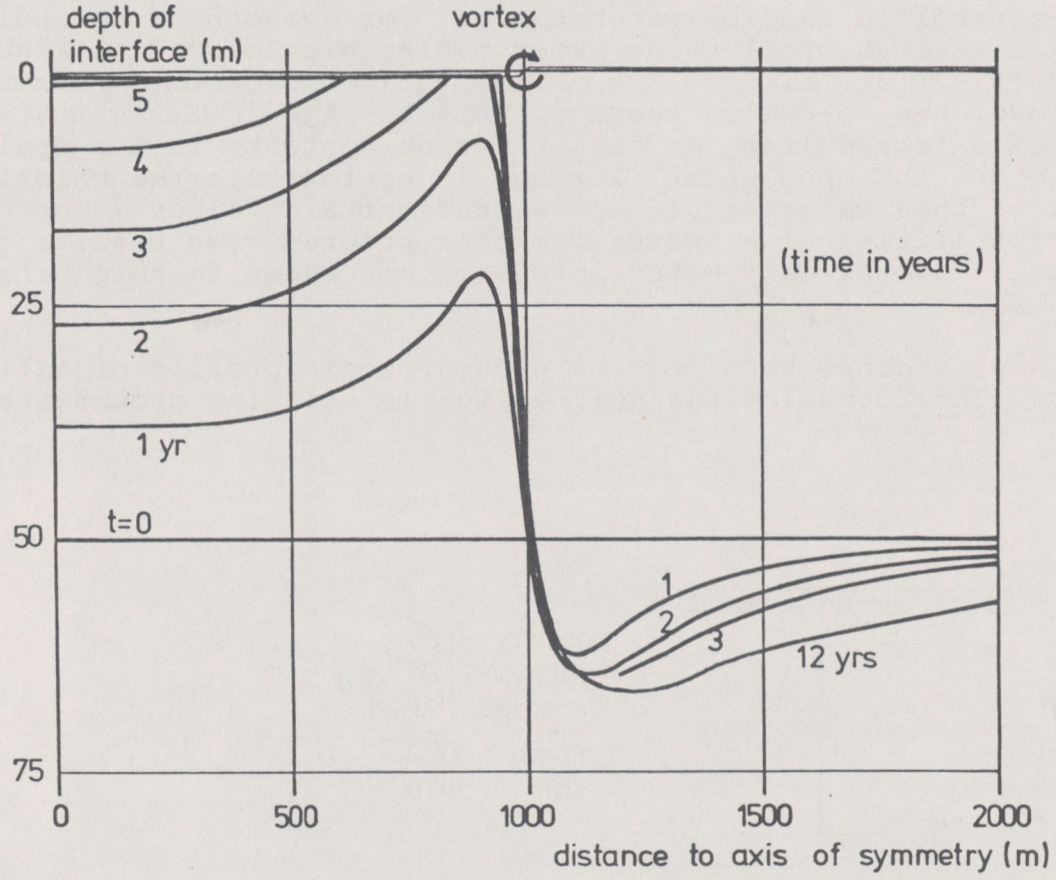


Figure 8. Initial and subsequent stages of the interface when at time = o the water table drops 1 m, timestep one month, radius of circular polder 1000 m

It is found out that saline seepage occurs first somewhere between the centre and the boundary of the polder. Awater also found this (Awater, 1980), however it can be shown that in isotropic aquifers it depends on whether there is a thick transition zone, on the initial depth of the interface in comparison with the dimensions of the polder, on the extent of the aquifer and on density differences, see next figure.

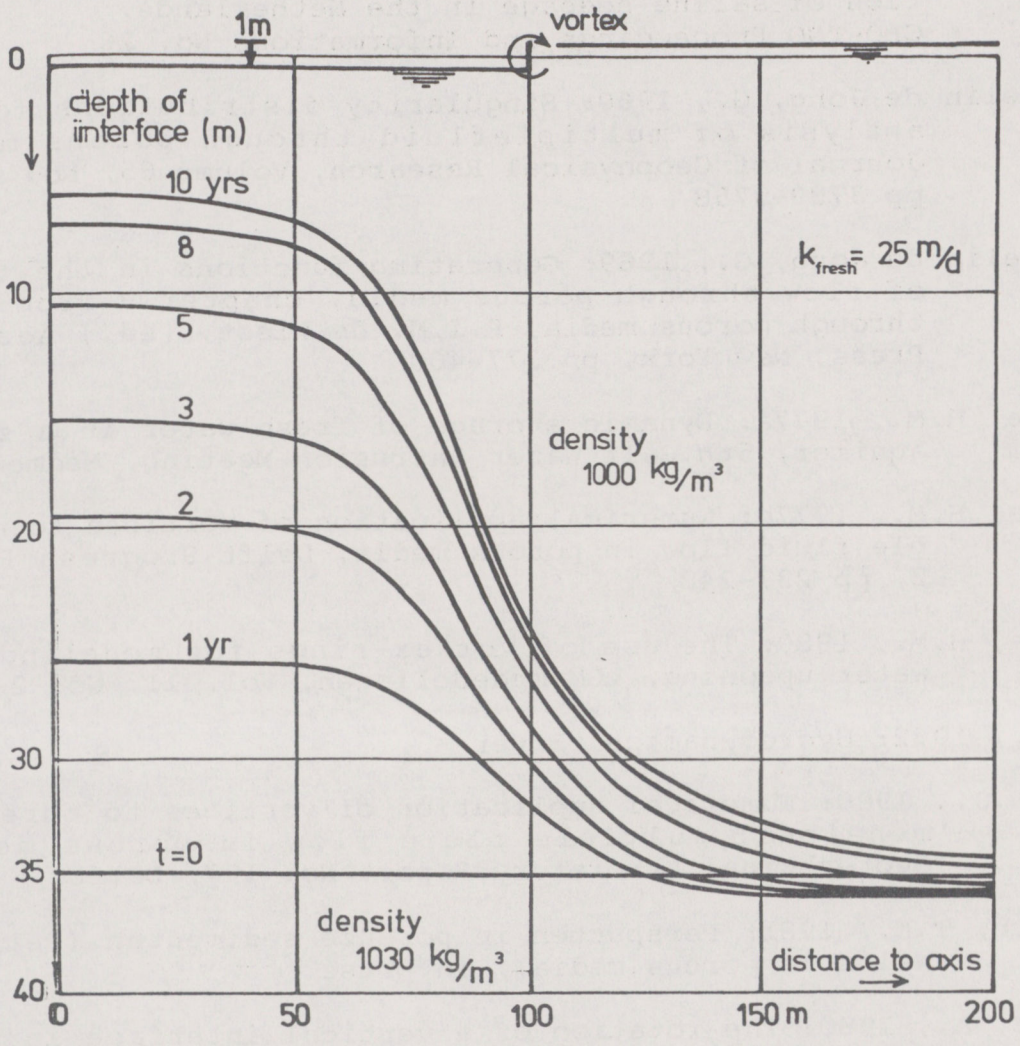


Figure 9. Initial and subsequent stages of the interface when at time = o the water table drops 1 m, timestep one month, radius of circular polder 100 m

CONCLUSION

Vortex-theory can be considered a valuable technique enabling us to simulate all kinds of ground water flow problems that are affected by density differences. However it will need further improvement to handle inhomogeneity as will be the case when dealing with actual problems.

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THE SIMULTANEOUS FLOW OF FRESH AND SALT GROUNDWATER IN HORIZONTALLY EXTENDED AQUIFERS

by C.J. VAN DUYN[1]

We shall discuss some fundamental properties of the simultaneous flow of fresh and salt groundwater in horizontally extended aquifers. When assuming a sharp interface (no dispersion) between the fluids, then it is possible to show that there exists a shearflow at every point of the interface, where it has a non-zero angle with the horizontal: see Dietz (1953) and de Josselin de Jong (1981).

Based on the shearflow concept, a nonlinear partial differential equation will be derived which describes the interface as a function of position and time. We shall study this differential equation, discuss a number of fundamental properties of its solutions and explain how these results relate to the physical problem. We consider both one-dimensional flows and axial-symmetric flows.

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ASSESSMENT OF THE FLOW REGIME OF FRESH AND SALINAR GROUND WATERS IN THE SEDIMENTARY COVER OF A SALT DOME - PROBLEMS AND METHODS

by G. Delisle, K. Schelkes[1]

ABSTRACT

The hydrogeology of an area of app. 400 km^2 in Lower Saxony, overlying a salt dome, is currently under intense investigation. Fresh and with depth increasingly salinar ground waters, which in some places interfinger, form a complex ground water flow pattern. Some investigational methods to measure vital data concerning the pressure and flow field of the ground water are discussed. Based on an idealised model, heat and mass transport by ground water flow is considered. Special attention is paid for natural convection phenomena within aquifers.

INTRODUCTION

The hydrogeological situation in a region, covering app. 400 km^2, located at the eastern border of Lower Saxony/W. Germany between the villages Lüchow, Gorleben and Schnackenburg is currently under intense investigation.[2] At more than 110 locations, about 300 piezometers (1 to 5 at any location depending on the number of penetrated aquifers) have been installed.
Regional ground water flow is influenced by several factors: An underlying salt dome causes the observed high salinities in deeper aquifers. Mass transport (salt) as well as heat transport (locally increased terrestrial heat flow due to the high thermal conductivity of the evaporites) effect to some extent the ground water flow pattern in the overlying aquifers. The development of a subglacial channel, now filled with permeable quaternary sandy material, during the Elster ice age has removed portions of the cap rock of the salt dome. Halokinetic processes have upwarped formerly horizontally layered tertiary sediments. The latter two processes have created a rather complex hydrogeological situation, which is characterised by a high variability of permeability of the sediments in lateral as well as vertical direction. The flow pattern of ground water is affected accordingly.

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STATEMENT OF PROBLEM

As the salt dome in question (Gorleben) is a candidate site for burial of nuclear wastes, it is of special interst to assess the velocity field of ground water as accurately as possible. In this context, not only fresh water aquifers but also horizons with salinar water are to be considered. Only after identification of all essential flow components an estimate on the consequence of potential failure of the salt dome as a natural barrier to migration of radioactive materials can be carried out.
It is, therefore, essential to understand the nature and the rate of mass transport (salt) from salinar into fresh ground waters. Diffusive, dispersive and convective processes need to be considered. The question, if and in what period of time a quasi - steady state between dissolution of salt at the interface salt dome - sedimentary cover and the mass transport by ground waters to the surface can be reached, is equally important. The paper below will concentrate on both questions. However, no final answers will be proposed, the points to be discussed resemble rather a selection of considerations, which have been discussed by those workers, which are involved in this project.

THE REGIONAL GEOLOGY

The major features, which might influence ground water flow in the area under study, will be discussed briefly below.

THE SALT DOME

The salt structure is trending in a southwest - northeasterly orientation across the study area. Its outline is shown in Figure 1. The length of the dome from the southwestern tip to the Elbe river totals 15 km, its greatest width amounts to 4 km. Today it is covered by a sedimentary layer ranging from 2oom to 300 m in thickness.
The high thermal conductivity of the evaporites cause a local increase of terrestrial heat flow over the dome. In interpreting thermal logs from observational bore holes, the influence of the salt structure on the regional temperature field in the subsurface has to be taken into account in appropriate fashion. A quantitative assessment of the thermal effects of the dome on the rock temperatures in the surrounding non-salt sediments can be found in Delisle (1980). In Figure 2, which shows rock temperatures at a depth of 150 m below mean sea level (MSL), the close interrelationship between the temperature increase and the shape of the dome is clearly demonstrated. The observed thermal anomalies at the center of the structure will be interpreted below in context with the discussion of regional ground water flow. The salt structure is surrounded by salinar ground waters. In general, only the uppermost aquifer carries fresh water.

THE SUBGLACIAL CHANNEL

A subglacial channel from the Elster ice age crosses in north - south direction the salt dome. At the central portion of the dome, the channel splits into two parts, whereby the less intensely

developed branch is diverted into an orientation parallel to the trend of the salt structure. The position of the channels with respect to the dome and the depth contour lines are shown in Figure 1 as well.
Over an area of more than 8 km^2 permeable quaternary sediments rest on the salt structure itself. A hydraulic contact between the ground water and the salt dome, therefore, exists.
Today, the distribution of fresh and salinar waters at shallow depth is apparently controlled by hydraulic interaction and the diffusive processes within or in the vicinity of the subglacial channel system.

THE SEDIMENTS

The salt dome is covered by tertiary and quaternary sediments. During the tertiary period, sediments of maritime origin, mainly clays and sands were deposited, At the same time, due to upward movement of the salt dome, erosion over the salt structure and development of elongated basins next to the dome flanks did occur. Several glacial stages since the lower pleistocene and intermittent warm epochs caused cycles of erosion and redeposition. Sediments are now mainly of glacial and fluviatile character. Sands, marls and loam, to a lesser extent clays, were deposited. Due to erosional and depositional cycles, no well defined aquifers bounded by low permeability layers could develop. Typically, clayey or marly layers of limited lateral extent are interspersed within the quaternary sands (see also Figure 3).
A detailed description of the sedimentary cover of the salt dome, from which the details above were extracted, can be found in Vierhuff et al. (1981).

METHODS TO ASSESS THE CHARACTER OF THE FLOW FIELD

Several methods, which have or will be tried to collect the necessary data for definition of the regional hydraulic potential, are discussed below.

SHALLOW GROUND WATER MOVEMENT

To obtain a first impression of the flow pattern of the shallow ground water, qualitative methods will be shortly discussed. Piezometric data from the uppermost aquifer down to a depth of 25 m below surface - in general sandy layers - were summarized in a map (Schelkes, 1981) as shown in Figure 4. The flow direction of these fresh - in a few cases slightly salinar - ground waters can be read off directly. The numbers on the lines shown indicate the height of the ground water table above MSL in m.
The flow potential is controlled by

a) the ground water regeneration rate, which amounts at the elevated regions (central portion of the map) to app. 100 mm/a
b) the ground water discharge area, defined by the surrounding lowlands.

For the two in the map indicated flow directions, making use of the obseved hydraulic gradient and an estimate of the permeability of 10^{-1} Darcy, pore velocity can be calculated to lie in the

range between 10 - 15 m/a.
Due to the complicated geologic, hydraulic and hydrochemical conditions, this type of consideration cannot be extended readily to deeper levels. Here, data concerning the thermal and salinar (density) field as well as age dating methods offer indirect paths toward a qualitative assessment of the actual flow patterns and velocities at depth.

THE TEMPERATURE FIELD

From all temperature logs measured in the observational bore holes, the temperature field of subsurface can be reconstructed with good accuracy. The thermal effects of the increased heat flow caused by the dome structure need to be subtracted from the observed temperature field. The remaining temperature field can then be analysed for anomalies, which are caused by convective heat transport.
The analysis can be carried out in a twofold fashion:
As shown in Figure 3, from a projection of the temperature field into a vertical cut through the subsurface, large scale effects of the ground water velocities on the subsurface temperatures, as indicated by the distortions of isotherms, can readily be recognised. Descending cold ground water originating from the nearby regeneration area cause a distinct cooling effect in the sediments over the subglacial channel. The areal extent of this negative thermal anomaly can be read off from Figure 2. From both figures it appears that this anomaly is closely connected with the general trend of the subglacial channel.
The positive anomaly at the western tip of the dome (Figure 2) has tentatively been interpreted as a result of upwelling waters within a tertiary formation. The second positive anomaly at the northern edge of the central portion of the salt dome is caused by upwelling waters in quaternary layers. Low rock temperatures immediately north of the dome are appearently associated with a ground water regeneration area (see also Delisle 1981 b).
To identify much slower flow components, which have only a minor effect on the trend of isotherms, another concept to interprete temperature data can be used. This method will briefly be discussed using again the information contained in Figure 3. As one might notice, the 18°C - 21°C Isotherms seem to be unaffected by ground water flow. However, their trend forms a slight angle with the horizontal plane suggesting a minor upward component of fluid flow toward the right hand side. A criterion is needed to distinguish between convective heat transport and small variations in the temperature field caused by lateral variations of the thermal conductivity distribution within the sediments. In Delisle (1981 a) a method is described, by which the predetermined terrestrial heat flow along each bore hole is divided by the measured temperature gradients of each sedimentary layer. One obtains for each subformation a set of apparent thermal conductivity values. If one of them differs significantly from the average value, a convective heat transport component is suggested. In the case mentioned here, it appears that the course of the isotherms is mainly controlled by the local thermal conductivity distribution and, additionally, by the thermal effect of the sloping, poorly conducting tertiary sediments at the channel flanks.

DENSITY DISTRIBUTION

A qualitative assessment of ground water flow can as well be derived from the fluid density distribution at depth. Of special interest is the area of the sedimentary fill of the subglacial channel, which represents the hydraulic contact between ground water and the salt structure. For this reason, this area has been subdivided into three sections(Figure 5). For each section, the density distribution in a vertical cut (indicated by broken lines in Figure 5) across the subglacial channel has been constructed (Figures 6 - 8).
Density values, which have been obtained by analysis of water samples drawn from the various filters, have been projected into these cross sections. The outline of the quaternary base is indicated as well. Whereas it reaches the salt structure at its deepest points, it is flanked at both sides by clays, which are interspersed locally with permeable sandy layers of tertiary age.

Figure 6 shows the profile in section I. The salinity in the deepest portion of the channel is relatively low as compared to the same depth level in Figure 7 (section II). 200 m below MSL, the density increases from a value of 1050 kg/m^3 (profile I) to 1150 kg/m^3 (profile II). The 1050kg/m^3 contour line raises, therefore, over a distance of 1,4 km by approximately 50 m.
Even more important, in profile II we find a local interfingering effect of fresh and salinar waters. Two interpretations seem to be applicable:

- A strong horizontal flow component of fresh water is bounded on the western flank by a weak component of upward flowing salinar waters, whereby the actual boundary between both flow regimes seems to be controlled mainly by the local permeability distribution.
- It appears possible that a salinar water body at shallow depth was maintained by a different flow regime at earlier times (conceivably during a time interval between the last two ice ages). This body will eventually decay due to diffusive and convective mass transport (non-steady state).

This shallow salinar water body seems to be rather old according to the results of age dating measurements (Moser, priv. comm. 1981).
The density distribution in section III, as seen in Figure 8, shows a similar situation as section II. No fresh water, however, is overlain by salt water bodies.
Finally, the density - depth relation (Figure 9) was deduced from the data plotted in Figure 7 and 8. All values, except the three points of the above mentioned abnormal shallow salinar fluid body, lie on one curve. The transition from fresh water to highly concentrated fluid appears to be rather sharp and seemingly independent from the geologic situation. Undoubtedly, however, does the inclusion of clayey and marly layers in the depth range from 100 m to 200 m below MSL influence the position of this interface.

QUANTITATIVE DERIVATION OF FLOW FIELD AT DEPTH FROM THE DENSITY DATA

A northward directed flow component of fresh or slightly salinar waters appearently reach at a depth interval from 0 - 200 m the boundary of the salt dome (section I). At depth, this flow is in

contact with a highly concentrated fluid body, which has been created by dissolution processes probably along the subglacial channel. In the area of segment II and III a salt water stratification was formed. The highly concentrated fluids appearently experience very little movement, as results of age dating measurements have shown. This high density body is overflown by low salinity waters.

PRESSURE FIELD

It is attempted to approximate fluid velocity values in close agreement to a concept presented by van Dam (1977) by
- use of Darcy's law
- pressure and density values measured in the observational bore holes
- hydrogeological data.

Samples collected in filters yield density data.
Pressure heads ought to be derived from piezometric measurements. This procedure is applicable, if by sufficient pumping from the filtered bore hole sections a constant density profile has been reached in each bore hole. For not well understood reasons, water leaks through pipes, therefore, a constant density or salinity profile cannot be maintained in most bore holes. Figure 10 shows an electrical conductivity profile, which was obtained in many wells in similar fashion at some time after the pumping procedure. For this reason, alternative methods to measure pressure at depth are being tried.

a) In cooperation with a private company, a sonde has been developed to measure conductivities up to app. 250 mS/cm. (close to saturation of water by salt). Values are measured with an accuracy of 1% - 2%. From the measured conductivity profile in the bore hole an average density will be calculated. The pressure acting in the filter area can then accordingly be determined.
b) A system is under development, which consists of the following components: two open-ended tubes filled with fluids of different specific density are lowered into the filter area. The other ends of the tubes are connected to a highly sensitive pressure transducer. Pressure in both tubes are measured at the surface. From both measurements and the known compensating effects of the fluid columns in the tubes both the absolute pressure and the depth of measurement can be determined.
c) A temperature compensated pressure transducer with an accuracy (absolute) of 5 mbar under a water column of 300 m will be employed.

MODEL CALCULATIONS

Models have been considered to investigate the influence of i.e. neglecting the true density profile or the use of k_f-values, even if density changes in the water column are known, on the calculated flow rate. Figure 11 shows a model profil with vertical permeabilities, an assumed density profile as well as pressure profiles, which are the result of an imposed vertical ground water movement of 1 cm/a or 5 cm/a. It was noted that the use of k_f-values can lead to errors of up to 20% by the determination of flow velocity, as long as the true density profile is used. However, if the

density profile is linearised between two filters, from which true density values are known, the error cannot only increase significantly but also might reverse the actual flow direction.
The implications of these deliberations have led to attempts by BGR, to determine the density profile of fluids in the neighbourhood of the wells as accurately as possible. The density of formation waters can be calculated from electrical conductivity. The conductivity depth function has to be derived first from geophysical well logs.
With the aid of a refined laminated shale model formulas to calculate the conductivity of ground waters using parameters from gamma-ray, focussed-electric, caliper and density logs were obtained. The equations contain rock specific parameters, which must be determined empirically. This is made possible by the use of i.e. the measured conductivity values of drawn water samples or the analysis of drill cores.
Figure 12 shows - conductivities have been converted into "densities" - as a first result the density profile in the subsurface around one piezometer group (3 piezometers). Fresh water overlies salt water, whereby a rather sharp transitional zone is located within a sandy layer from 70m to 120m below MSL. For this piezometer group the vertical flow component using the Darcy relation has been calculated. For the period of Nov. 1980 an upward movement of ≤ 1 cm/a in the upper portion, a downward motion of ≤ 1 cm/a in the lower portion of this profile was obtained. More accurate data will be presented as soon as the true distribution of k-values are available.

NUMERICAL MODELS

Independently from the actual field observations, it is considered to be worthwhile to simulate the naturally occurring mass and heat transport phenomena by computer codes. Boundary conditions and material parameters are chosen such that they represent to a first approximation the field conditions.
The models presented here serve a twofold purpose:

- The time dependence of processes occurring, as a salt - fresh water system is carried from one quasi - steady state into another (due to i.e. a changing ground water flow pattern), needs to be understood.
- Very slow movements of seemingly motionless fluids in aquifers can often not be detected by measurement of hydraulic heads. Lateral temperature gradients i.e. may, however, cause very low velocities, which might over time introduce mass transport capable to alter noticeably the salinity field. (These processes ought to be considered as well as i.e. age dating methods are applied.)

GOVERNING EQUATIONS AND BOUNDARY CONDITIONS

Boundary conditions and equations, which are common to all models presented below, are outlined in this section.
The simplified governing equations are:
equation of continuity

$$\mathrm{div}\,(\varrho \cdot v) = 0 \qquad (1)$$

Darcy's law

$$\nabla P - g\varrho + \left(\frac{\mu}{k}\right) \cdot v = 0 \qquad (2)$$

heat transport equation

$$\frac{\partial T}{\partial t} + \frac{\varrho \cdot c_{Water}}{\varrho \cdot c_{System}} \langle v, \operatorname{grad} T \rangle = a \, \Delta T \qquad (3)$$

mass transport equation

$$\frac{\partial C}{\partial t} + \frac{1}{\phi} \langle v, \operatorname{grad} C \rangle = d \, \Delta C \qquad (4)$$

P=pressure
T=temperature
C=concentration
d=hydrodynamic dispersion
ϕ=porosity
ϱ=density
a=heat conductivity
ϱc=specific heat capacity
v=velocity
g=gravitational constant
k=permeability
μ=viscosity

All model calculations are performed for a vertical cut (x-z) across a system of aquifers, which are bounded at the lower border by impermeable rocks (v_z=0). On the lateral boundaries, horizontal flow components v_x are set to zero. The pressure as a function of x at the upper boundary is maintained constant. Correspondingly, temperatures at the upper and lower boundary are constant, heat flow q across lateral boundaries is assumed to be q_x=0.
The temperature dependent density and viscosity of fresh water has been approximated by the following polynomials:

$$\varrho(T) = 1{,}000065 + 2{,}67707\ \text{E-5} \times T - 6{,}01562\ \text{E-6} \times T^2 \ (g/cm^3) \qquad (5)$$

$$\mu(T) = 1{,}002 + (20 - T) \times 0{,}0298 \ (\text{Bar s}) \qquad (6)$$

Density as a function of T and C is defined by:

$$\varrho(C,T) = \varrho(T) + (C \times 0{,}7688) \,/\, (\varrho(T) \times 1000 + C) \ (g/cm^3) \qquad (7)$$

THE SIMULTANEOUS HEAT AND MASS TRANSPORT IN A FRESH-SALT WATER SYSTEM

From the observed situation at the field site the question has arisen, with what speed salt waters might react to changes in the fresh water flow regime. Especially the situation in section II, where salt water overlies fresh water, implies the possibility of a very slow drift of the system from one quasi - steady state into a new one.
To assess to a first approximation the time dependence of mass transport in response to a changing flow pattern in an aquifer, intersperses with low permeability zones, the following simplified

model was considered (Figure 13):
A 200 m thick zone with a horizontal permeability of 1 Darcy ($k_{ver.}$=0,1 Darcy) is assumed to contain three lenses of low permeability($k_{hor.} = 10^{-3}D$, $k_{ver.} = 10^{-4}D$). This section is underlain by a 100 m thick layer with the same permeability as the lenses. The length of the cross section totals 9,5 km.
As indicated in Figure 13, the water table raises or dips under its mean elevation by a maximum of 5,6 m. While the upper 200 m of the system contain fresh water, the underlying sediment is filled initially with highly saline water. The concentration increases linearly from the fresh - saline water interface from 349 kg salt/ 1 t water to 359 kg salt/ 1 t water.
From the central portion of the cross section water is forced along hydraulic gradients onto the fresh-saline ground water interface, where it is contaminated by diffusive - dispersive mass transport mechanisms. The flow velocities in the deeper parts of this system are slow enough to allow the simplifying assumption of combining the diffusional and dispersional mass transport into one term in (4). A d-value of $3 \times 10^{-10} m^2/s$ was chosen. The desirability of the inclusion of a refined expression for the dispersive mass transport at a later time is recognized.
Equations (1) - (4) are evaluated step by step by an iterative process, whereby an explicite finite difference code is used.
Figure 14 shows the temperature effects of the ground water flow on the subsurface rock temperatures after a period of 9 000 a. In the center, the downward flowing groundwater induces a local cooling effect, somewhat more pronounced on the right hand side. Latter effect is due to the somewhat higher integral permeability of the subsurface (see also Figure 13) in the corresponding flow direction.
The salinity distribution 3 000 years and 9 000 years after initiation of flow are shown in Figure 15 - 16. On the left side, the continuous low permeability layer at a depth interval from 110 m to 170 m effectively slows down any upward movement of saline water by convective transport. On the far right hand side, vertical flow is unimpeded by low permeability zones. With time, an increasing contamination of shallow zones by salinar waters is observed. The buildup of the contaminated area in the upflow regions has by no means come to an end after 9 000 years.
This result - as well as other models, which have been calculated, but are not shown here - strongly imply that the salinity distribution in the subsurface of our study field is probably not in a steady state situation. The numerous clayey and marly layers obviously would in case of changing ground water flow patterns impede the response movement of salinar water bodies. One may speculate that during the last glaciation 15 000 years ago ground water flow was greatly reduced in the study area due to permafrost at the surface. The ground water system below probably responded by attempting to slide into a stable configuration, whereby salinar waters - formerly transported upward by ground water flow - sank to deeper levels, fresh water resting on top of them. Some salinar waters might have been trapped effectively in low permeability zones. The readjustment of the ground water system after the end of glaciation might not - as far as the salinity distribution is concerned - be complete today.

NATURAL CONVECTION

Saline water bodies at depth are frequently thought to have come to an almost complete rest. They are known to yield high "age dates". It is, however, in the context of safe burial of radioactive materials of interest, if small scale movements can be inherent in such bodies and which magnitudes of velocities can be expected.
One driving force of natural convection are lateral temperature gradients, which might be induced by lateral variations in the thermal conductivity distribution of the fluid-saturated rock matrix. Such a situation might arise, where i.e. clays (poor thermal conductor) and sands (good conductor) happen to rest next to each other. For example, at the flanks of a subglacial channel such a geologic setting can exist.
A simplified model, using the same governing equations and relations mentioned above, was considered. The following boundary conditions were imposed in addition:
In a 300 m deep vertical cut, spanning over a lateral distance of 675 m, an initial temperature field as depicted in Figure 17 is assumed. The lowermost 100 m of the permeable material contains salinar water. The salt concentration in this body increases from the top (342 kg salt/1 t water) to bottom (352 kg salt/1 t water). The diffusivity value was chosen to be $d = 10^{-10} m^2/s$. Dispersive effects were neglected due to the extremely low flow velocities obtained.
At a depth interval from 105 m to 155 m a zone of low permeability ($k_{hor} = 10^{-3} D$; $k_{ver} = 10^{-4} D$) was included in the otherwise permeable material ($k_{hor}/k_{ver}=10$; $k_{hor}=10$ D). Starting with no other than the by the thermal effects induced hydraulic gradients, heat and mass transport as a function of time were calculated. As in earlier published cases (Delisle 1981a) natural convection did develop.

After a period of t=20 a, two convection cells were observed. As by the temperature field reflected (Figure 18), the ground water in the uppermost layer responded by a clockwise rotation, whereby lateral velocities up to 6 mm/a, vertical velocities up to 1 mm/a on the left side and downward velocities of 0,5 mm/a were reached.

The second convection cell is located in the salinar water body, which shows clockwise rotation again. Since this cell is affected by higher lateral thermal gradients, the maximum vertical (0,45 cm/a) and horizontal (2,4 cm/a) velocity components are somewhat higher than in the cell mentioned above.
Movement in the low permeability layer is almost absent. The fresh water layer between this zone and the salinar water body tends to show at early times an upward directed velocity component on the left side, but appears to form in later stages its own convective cells. No stable state has yet been reached after t = 20 years.

The calculation was stopped after this point in time as the computer time needed to reach later stages appeared to be excessive. The flow pattern in the major convection cells, however, is not expected to change significantly, before a noticeable change in the salinity field due to diffusive processes.
The very low flow velocities at depth obtained from both models mentioned above seem to imply that in many cases mass transport in nearly stagnant salt water bodies is controlled by diffusion rather than by dispersion or convection.

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Figure 1 : Outline of salt dome Gorleben (continuous line) Broken lines indicate depth contours of the quaternary base (data from " Hydrogeologisches Untersuchungsprogramm Gorleben")

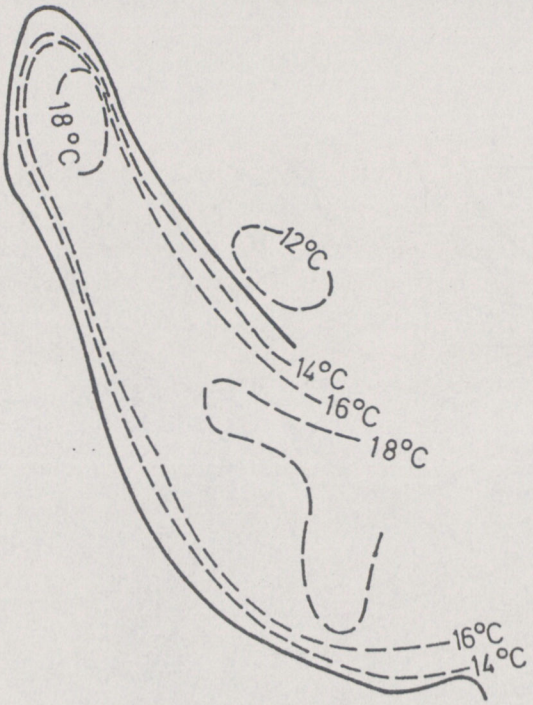

Figure 2 : Rock temperatures 150m below MSL
Continuous line indicates shape
of salt dome

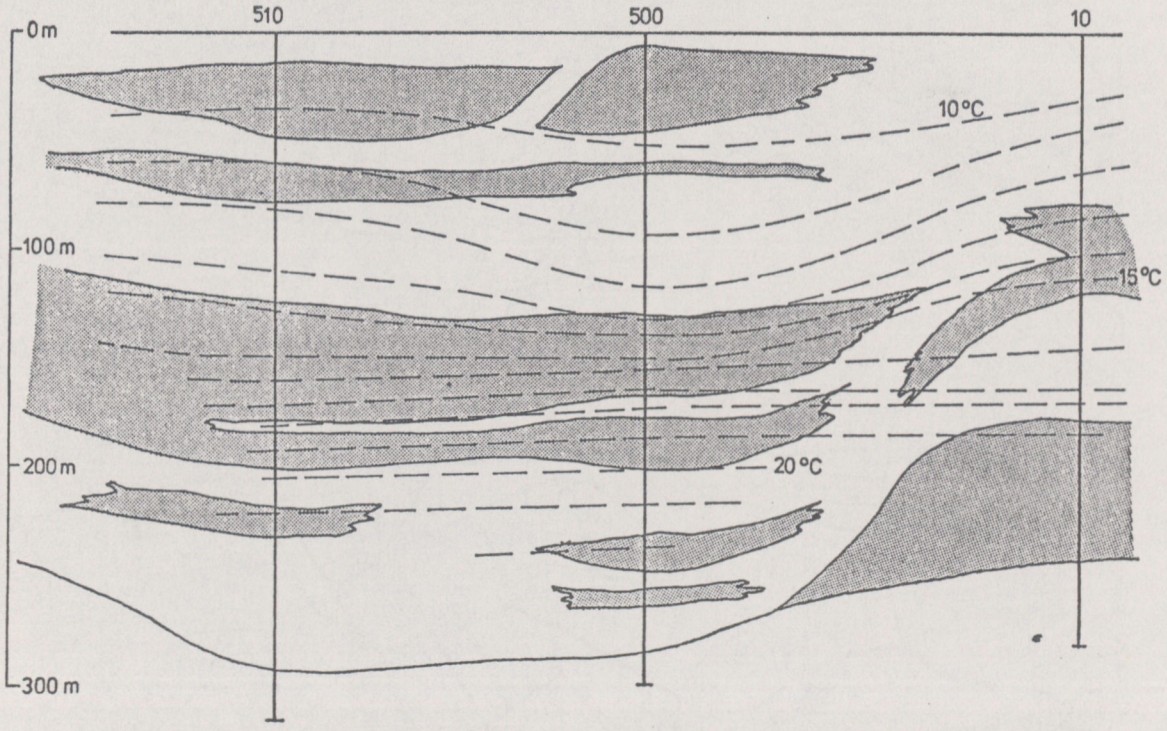

Figure 3 : Vertical cross section through subglacial channel
Low permeability zones dark, permeable zones white
Downward flowing ground water in a direction perpendicular to cross section depresses locally the isotherms at well 500

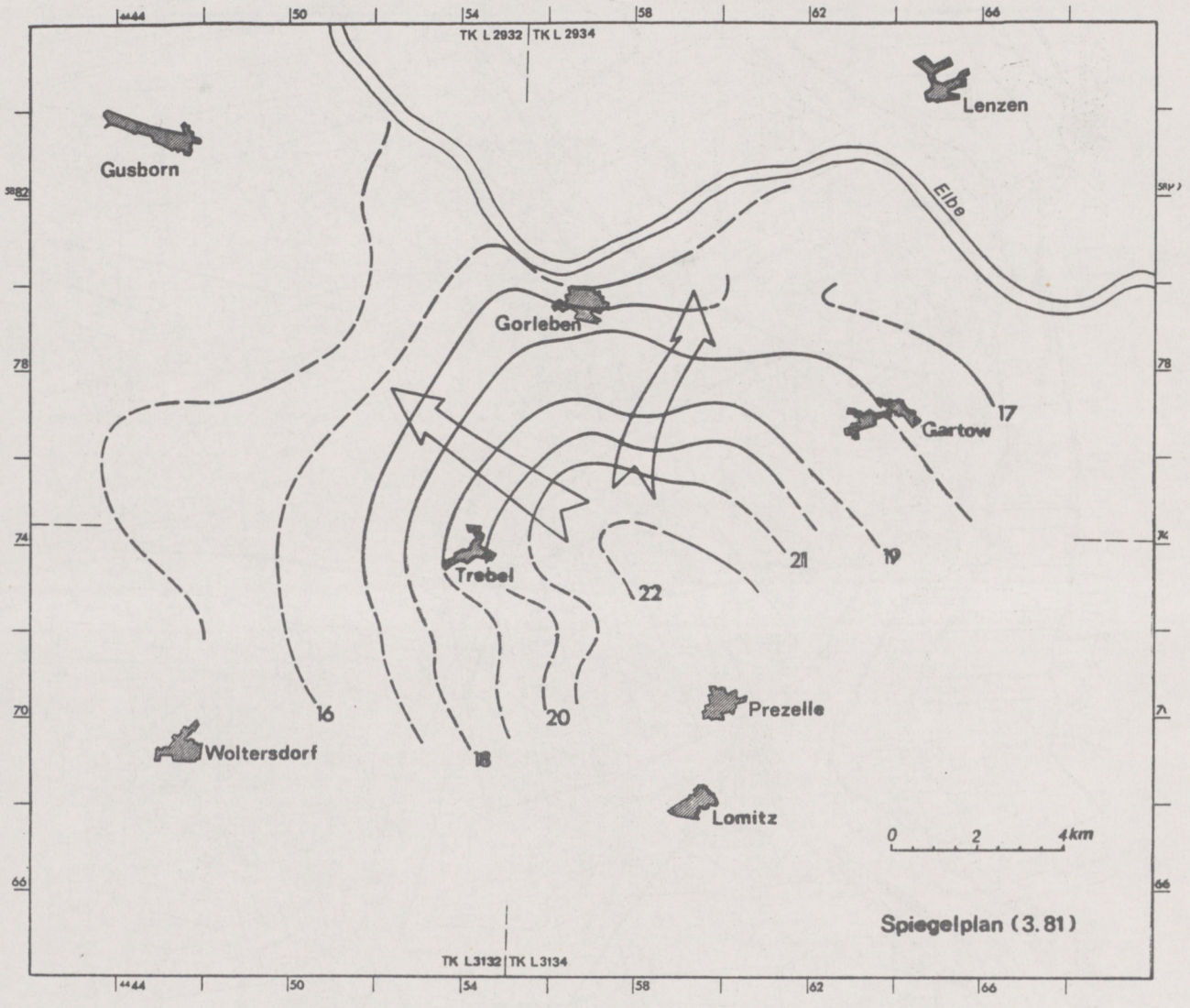

Figure 4: Elevation of water table in the study field above MSL (March 1981)

Figure 5: Positions of segments I, II, III

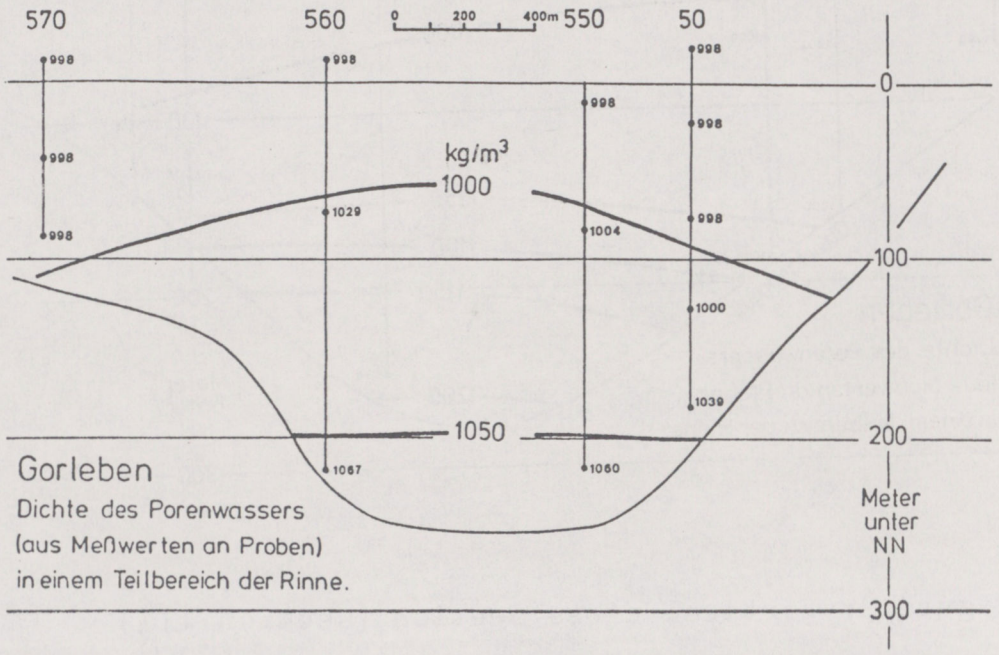

Figure 6: Vertical cross section (section I)

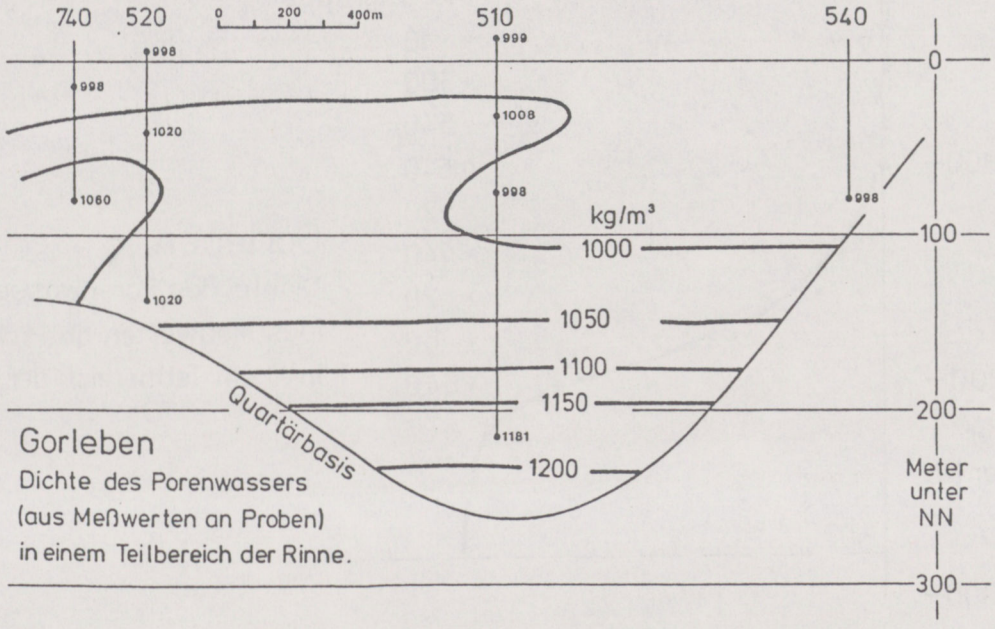

Figure 7: Vertical cross section (section II)

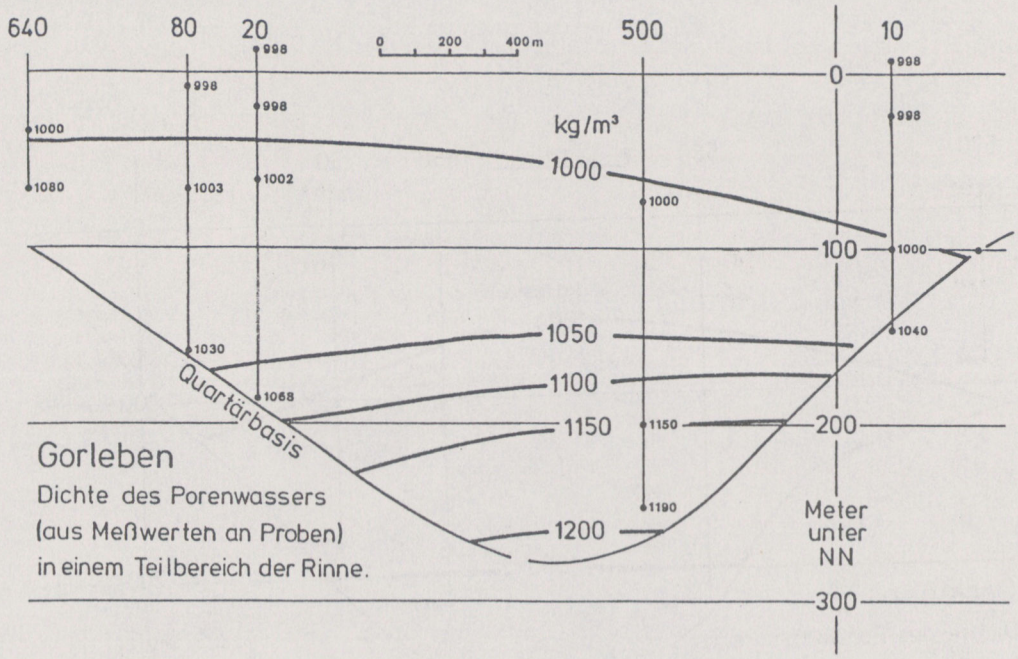

Figure 8: Vertical cross section (section III)

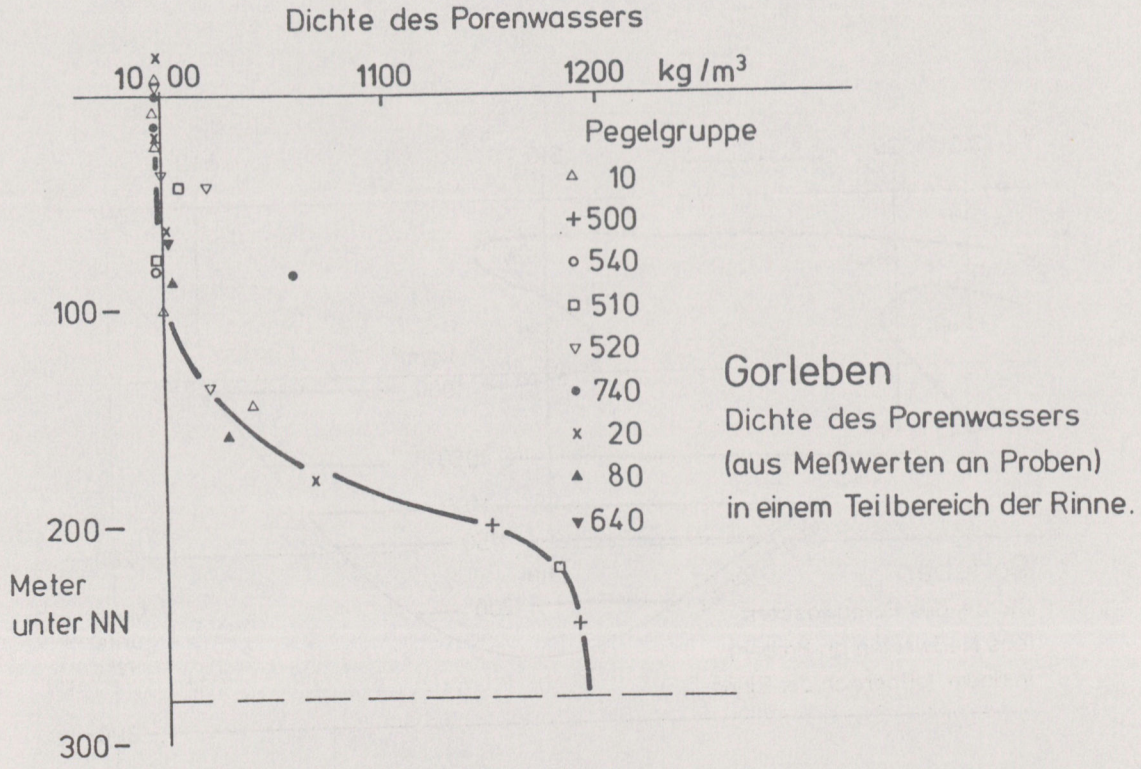

Figure 9: Density - depth relation as deduced from data plotted in Figures 7 and 8

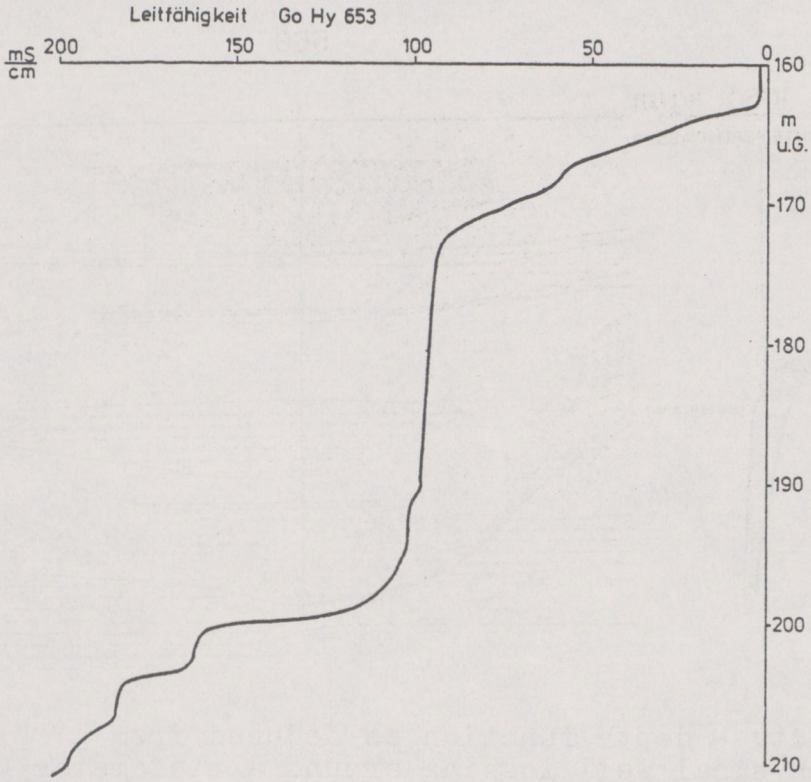

Figure 10: Electrical conductivity profile in a damaged piezometer

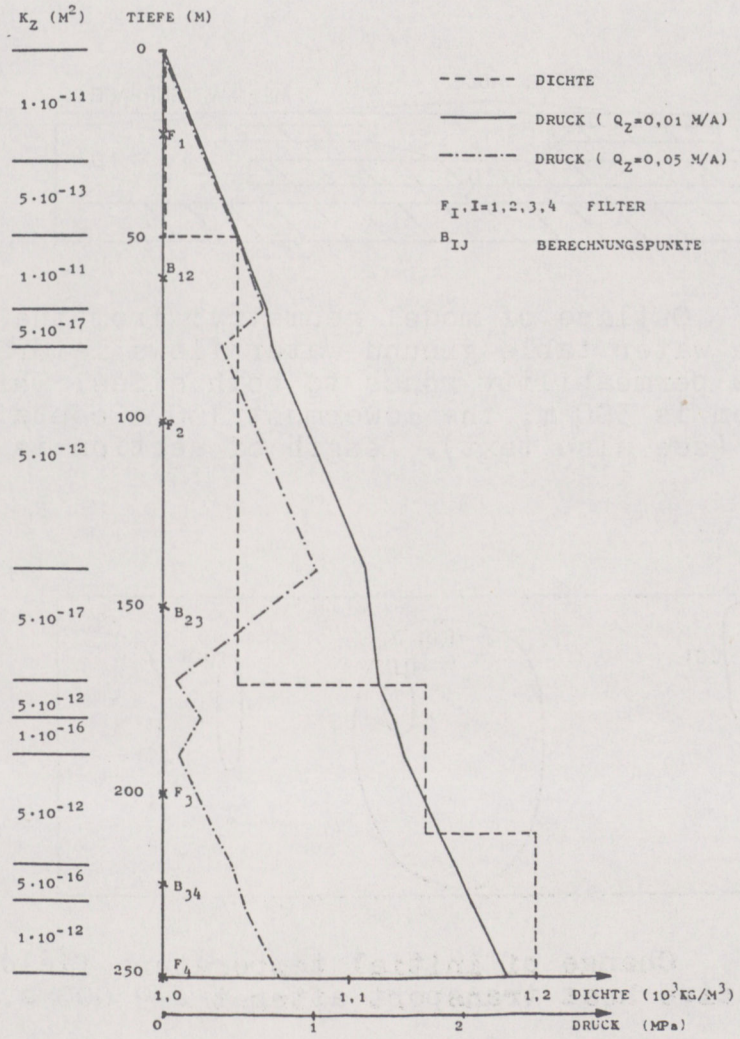

Figure 11: Model profile with vertical permeabilities, an assumed density profile and pressure profiles, resulting from an imposed vertical groundwater movement

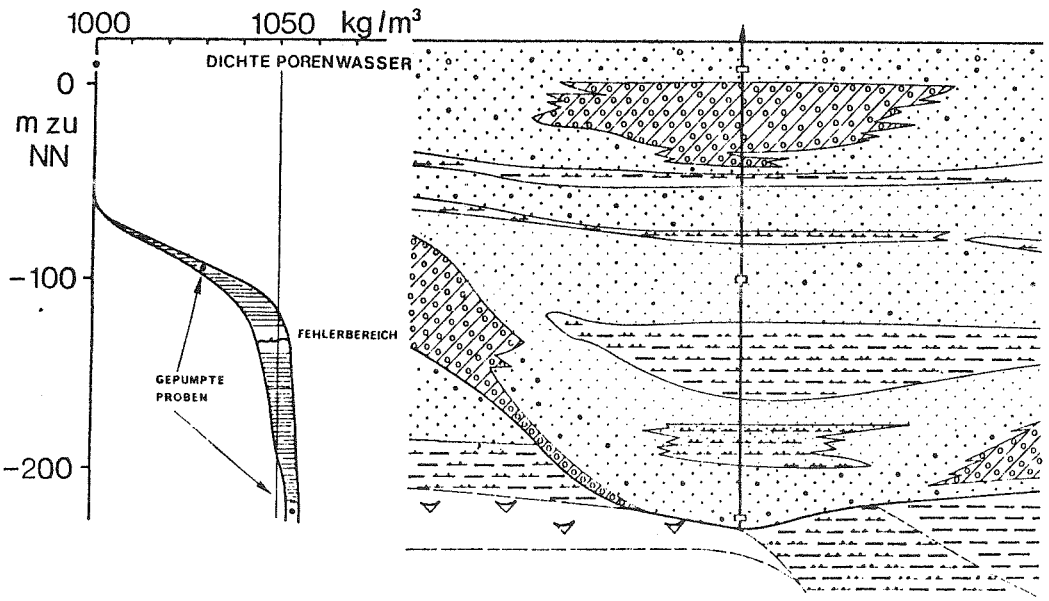

Figure 12: Density - depth function as deduced from geophysical well logging around a piezometer group

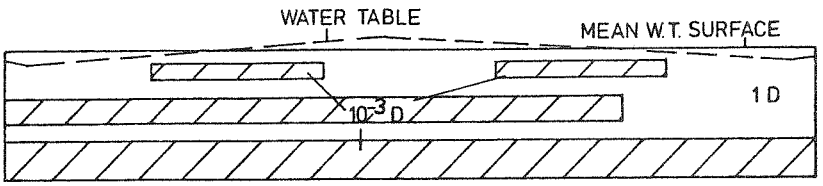

Figure 13: Outline of model geometry: from the high point of the water table ground water flows laterally, impeded by low permeability zones to both sides. Thickness of section is 300 m, the lowermost 100 m contain salt water (see also text). Length of section is 9,5 km.

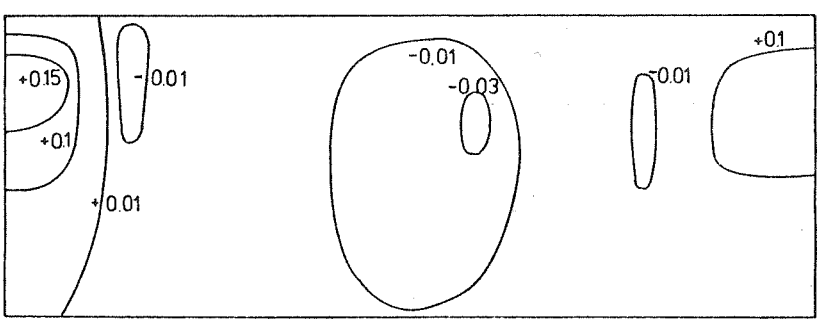

Figure 14: Change of initial temperature field due to convective heat transport after t = 9 000 a.

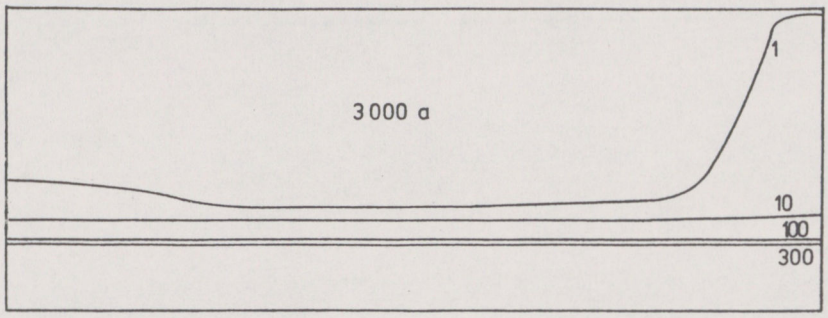

Figure 15 Salinity distribution after t = 3 000 a

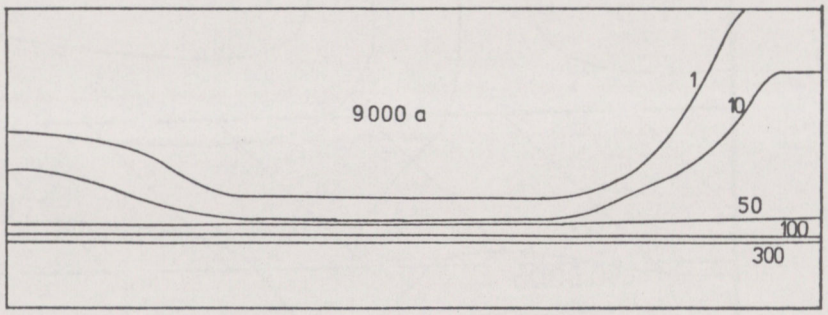

Figure 16 Salinity distribution after t = 9 000 a

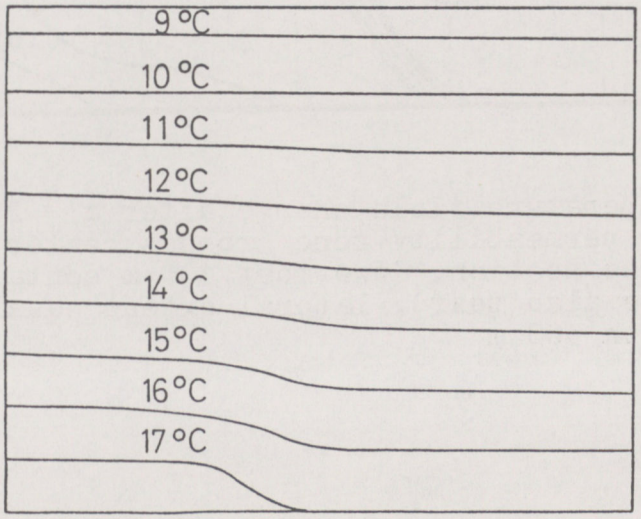

Figure 17: Initial temperature field, on which natural convection model is based

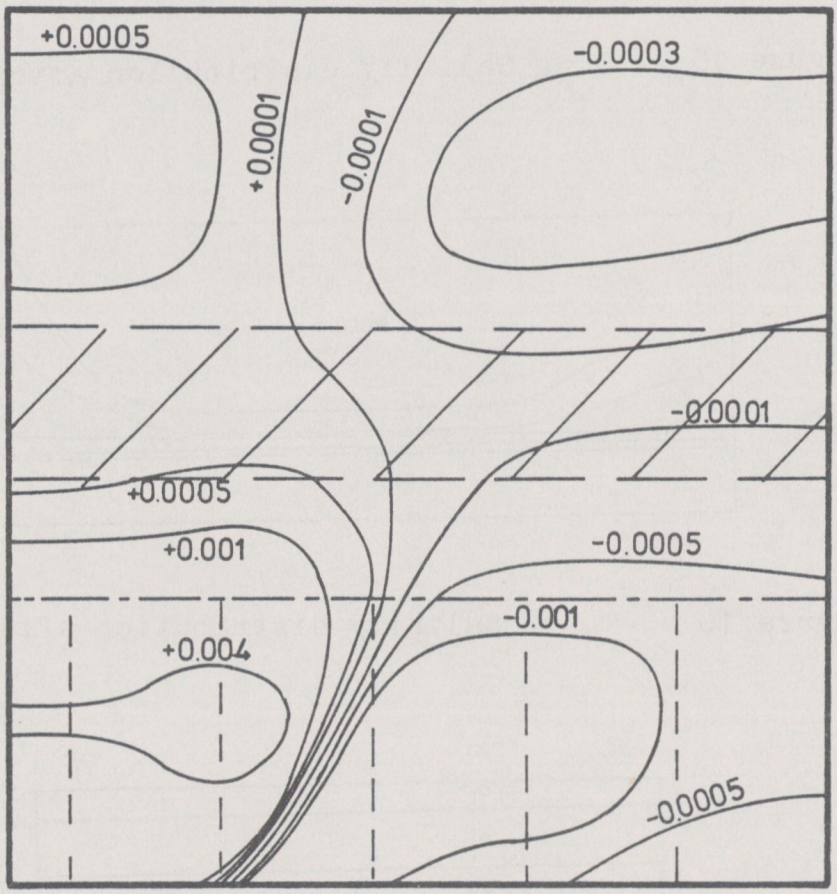

Figure 18: Temperature field change after t = 20 a
Low permeability zone crosses center portion of cross section, lowermost 100 m contain salt water (see also text). Lateral extent equals 675 m, depth 300 m

THE SUBTERRANEAN FLOW OF FRESH AND SALT WATER UNDERNEATH THE WESTERN BELGIAN BEACH

by L. LEBBE[1]

ABSTRACT

On the gently sloping sandy runnel and ridge beach thirty holes were drilled through the unconfined aquifer. The unconfined aquifer consists of a complex of permeable and semi-permeable layers with a thickness of about 30 m. In each of the boreholes a resistivity logging was performed. The resistivity logging thus obtained provide a fairly good idea of the fresh, brackish and salt water distribution underneath the beach. Five resistivity profiles perpendicular to the shore line were drawn. At one of these profiles the hydraulic head pattern has been measured continuously in the upper and the lower part of the aquifer. From these piezometers groundwater has been sampled for determining the mean anion- and kation concentrations. Based on these data a mathematical model, which describes the flow of fresh and salt water has been developed.

INTRODUCTION

In the western Belgian coastal plain a large part of the water demand is supplied by catchments in the dune belt. Because of two major reasons the demand of water has increased continuously. The first reason is the expansion of the tourism since W.W.II. This not only caused an increase in the water demand but the extension of the urbanized area also has reduced the natural infiltration in the dunes. The second reason is the connection of the farms and local communities in the polder area to the public water supply system.

Since 1958 several hydrogeological studies have been undertaken in the Belgian coastal area (J. DE PAEPE & W. DE BREUCK, 1958). The location of the fresh-/salt-water interface in the coastal plain has been mapped by resistivity soundings, resistivity traverses and borings (W. DE BREUCK & G. DE MOOR, 1974). This map shows that the fresh-water supplies in the polder area are limited to small domestic uses. The only parts in the coastal plain which can provide larger water quantities are the dune areas.

In the Westhoek area of the dunes (L. LEBBE, 1978), covering a surface of 9 km² (fig. 1), a hydrogeological study has been carried out. The larger part of the area is occupied by dunes

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bounded in the north by the beach and in the south by the polders. The western part of the area is a natural reserve. The eastern part comprises the urban area and the water catchment. More than eighty borings have provided a detailed picture of the lithology and the stratigraphy of the aquifer. By means of grain-size analyses, geo-electrical well loggings,pumping tests, measurements of tidal and seasonal fluctuations of the hydraulic head the hydraulic parameters have been estimated. They include the horizontal and the vertical permeabilities, the elastic storage coefficient, and the storage coefficient near the watertable.

By means of a mathematical model the hydraulic parameters of the pervious and semi-pervious layers were adjusted and the groundwater flow was calculated. From the hydraulic head configuration three zones of groundwater flow have been deduced (fig. 1), one where the groundwater flows towards the polder area, one where grounwater flows towards the sea. Near the French border the fresh groundwater flow towards the sea is a maximum while it is very much reduced in the eastern part of the area studied.

The chemistry of the groundwater was investigated by means of resistivity logging and chemical analyses of groundwater samples. The sandy deposits below the dunes resting on a clayey substratum only contain fresh water. In the polder area the distribution of salt, fresh and brackish water is complex. In the lower part of the aquifer salt water usually occurs. In the upper part the water is usually fresh and sometimes brackish. Between the salt and the fresh water there is a transition zone of brackish water of variable thickness. At the beach near the high-tide line the upper part of the aquifer contains salt water. The lower part contains fresh water.

Because of the rather unusual occurrence and flow of salt water above fresh water the study was further focused on the groundwater flow underneath the beach. Most of the studies about saltwater intrusion (TODD, D.K., 1980, BEAR, J., 1972) deal with the flow of fresh water on denser brackish and salt water. The Badon Ghyben-Herzberg formula gives the relation between the depth of a sharp fresh-/salt-water interface below mean sea level and the height of the water table. It assumes a static equilibrium and a hydrostatic pressure distribution in the fresh water region with stationary seawater. In the course of time many investigators have improved this formula by taking into consideration the flow, horizontal as well as vertical, in the fresh and/or salt water and the dispersion and diffusion processes. The configuration of the interface between fresh and salt water has also been investigated intensively in groundwater reservoirs with semi-pervious layers. The practical problem which has received most attention in this context is the upconing of brackish water under a pumped well.

CHARACTERISTICS OF THE BEACH AT DE PANNE

The Belgian coast is a sandy runnel and ridge beach. At De Panne the mean slope is 1,1 %. The beach is covered by semi-diurnal tides. The difference between the high and low tide is at spring tide approximately 5 m and at neap tide 3 m. The horizontal distance between the mean high-tide line and the mean low-tide line or the width of the beach ranges from 300 m to 450 m. The sealevel reaches its lowest point at 0 m T.A.W.[1] and its highest point at +5,5 m T.A.W. With a frequency of about eleven cases in ten years the sealevel overreaches +5,5 m T.A.W.level (R. COPPE & L. DE KEYSER, 1967).

From the hydrogeological profile (fig. 2) of the adjacent dune area (L. LEBBE, 1979) we may deduce the constitution of the aquifer. The substratum of the unconfined aquifer is formed by impermeable clay of Eocene age. The lower part of the aquifer is formed by medium to coarse medium sands of Eemian age, with a mean horizontal hydraulic conductivity of 24 m/d. The larger part of the aquifer consists of fine medium sands. Their mean horizontal hydraulic conductivity is 9 m/d. Lenses of silty fine sands with a small hydraulic resistance can occur. The top of the aquifer is formed by medium sands with a estimate horizontal hydraulic conductivity of 12 m/d. Before the urban area of De Panne on the back shore occurs a sandy layer with an accumulation of shells between the fine medium sands and the uppermost medium sands. On some places underneath the beach there can occur a silty fine sand layer between the lower medium to coarse medium sands and the upper fine medium sands which forms a semi-permeable layer.

RESISTIVITY LOGGING

In June, July and August 1980 five rows of borings were made on the beach at De Panne. These rows were situated perpendicular to the coast line, beginning at the French border K0 and continuing with K1, K2, K3 and K4 respectively 1, 2, 3 and 4 km to the north east (fig. 3). Drilling was done by the rotary method; the drilling mud consisted of a suspension of water and organic additives that degrade with time. The borehole had a diameter of approximately 100 mm. Due to this small diameter the weight of the drilling equipment was kept small involving only low costs and speed. Hence it was possible to drill very close to the low tide line. The small borehole diameter provides another advantage. The resistivity of the surrounding sediments, measured with a long normal device, can be compared with measurements in a screened well of diameter 80 mm (L. LEBBE, 1978).

According to the long normal method, a current and a potential electrode 1 m apart are lowered into the borehole. The other current and potential electrodes are placed at the surface at some distance from the borehole and from each other.

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1 T.A.W. Datum level of the second general leveling (National Geografical Institute)

The electrodes lowered into the borehole were ring electrodes with a diameter of 40 mm and a height of 10 mm.

The methods were compared as follows. A long normal log in a borehole Ø 100 mm filled with drilling mud was compared with one performed in screened well of Ø 80 mm at a very short distance.

The mean of 23 ratios between the resistivities in the borehole filled with drilling mud and those in the screened well was about 0,98, the standard deviation being 0,14. The largest value for the ratio was 1,27 and the smallest 0,74. As expected the two kinds of measurements did not differ much because of the ring electrodes and the rather small diameter of the boreholes. Both measured resistivities did not differ much from the true formation resistivity.

In a former hydrogeological study in an adjoining area chemical analyses of groundwater samples and resistivity logging of a screened well have shown that the ratio between the resistivity of the sediments, and the resistivity of the pore water in the sediments has an average value of approximately 2,7. This ratio varies between 2,3 for silty very fine sands and 4,5 for medium to coarse medium sands. According to the resistivity limits of fresh, brackish and salt water as proposed by G. DE MOOR & W. DE BREUCK (1969) (table 1) we have deduced the corresponding resistivity limits of the sediments filled with fresh, brackish and salt water, measured with the long normal device.

Tabel 1. Resistivity of fresh, brackish and salt water and resistivity of the sediments in the area filled with salt, brackish and fresh water measured with a long normal device (AM = 1 m)

Quality	Resistivity of the water ρ_w (in Ωm at 10°C)	Resistivity, ρ_m, measured with the long normal device (in Ωm)
Fresh	$\rho_w > 7,5$	$\rho_m > 20$
Brackish	$0,94 < \rho_w < 7,5$	$2,5 < \rho_m < 20$
Salt	$\rho_w < 0,94$	$\rho_m < 2,5$

The results of the resistivity logging on the beach are represented in five resistivity profiles perpendicular to the coastline (fig. 3). Three zones can be distinguished. A zone where the resistivity is larger than 20 Ωm, consisting of sediments filled with fresh water, a zone where the resisitivity varies between 2,5 and 20 Ωm, containing brackish water, and a zone where the resistivity is smaller than 2,5 Ωm with salt water.

In the resistivity profile KO at the French border salt water rests upon fresh water. There is a brackish transition zone of varying thickness. At the limit of the dune and the beach the transition zone between the fresh and the salt water is nearly vertical. Further under the beach in the direction of the sea the transition zone is nearly horizontal,gently rising towards the sea. Hence the salt-water lens decreases in thickness towards the sea. The resistivity of the sediments with fresh water decreases towards the sea, which indicates that the water becomes more mineralised. This rather simple configuration of fresh, brackish and salt water under the beach is found where the groundwater divide is situated the farthermost inland. This is where a large part of the fresh water underneath the dune drains towards the beach.

From the French border in the direction of the agglomeration of De Panne the water divide approaches the beach so that the fresh groundwater flow towards the sea gradually diminuish.

At 1 km from the French border the resistivity profile K1 shows underneath the back shore the same picture of the resistivity profile, at the French border. Underneath the fore shore the fresh water gradually passes into brackish water. Underneath the fore shore the salt-/fresh-water interface first rises and then descends towards the low-tide line where salt water flows above brackish water.

At 2 km from the French border the resistivity profile K2 shows underneath the back shore approximately the same picture of K0 and K1. Underneath the fore shore it shows a quite different picture. Here the thickness of the salt-water layer is very much reduced while brackish water appears under the fresh water. The resistivity of the fresh water reduces in the direction of the sea.

At 3 km from the French border before the agglomeration of De Panne where the water divide approaches the beach the resistivity profile K3 shows a completely different picture. In the lower part of the aquifer the extension of fresh water is strongly reduced. The fresh water passes into brackish and salt water over a short distance. The largest part of the lower aquifer is here occupied by salt water with a slightly lower salt content than the water in the upper part. At the low-tide line a lens of brackish water appears halfway the aquifer.

At 4 km from the French border at the NE corner of the urban area of De Panne the resistivity profile K4 shows similar to K3. The occurrence of fresh water in the lower part of the aquifer is limited to a small zone under the back shore and the transition from fresh to brackish and salt water occurs within a short distance. A lens of brackish water appears under the low-tide line.

HYDRAULIC HEAD MEASUREMENTS

A first series of hydraulic head measurements was made to determine the mean hydraulic head just underneath the surface of the beach. Because the hydraulic head fluctuates with the tidal movement it was necassary to measure it continuously at different distances from the high-tide line. Six piezometers were installed in a plane perpendicular to the shore line, the same plane of resistivity profile K0. The screens of the piezometers were placed at the -2 m T.A.W. level, in the upper part of the aquifer. They had a diameter of 63 mm and a screen length of 0,5 m. The piezometers P1, P2, P3, P4, and P5 were placed at respectively 360, 245, 165, 90, and 5 m from the landmark (stone marking the border between Belgium and France) in the direction of the sea. The piezometer 6 was placed in the dune area at 70 m from the landmark in the direction of the land. This landmark almost coincides with the high-high-tide line.

In the piezometers the pressure was measured by means of a pressure transducer of the VEGA-TYPE 137 series.By means of a data logger LOGMASTER MDL 1000 of MESS + SYSTEM TECHNIK the pressure was directly converted into a fresh-water head[1]. The results were printed and recorded on a magnetic tape. These measurements were made every ten minutes on the six piezometers. In fig. 4 the fresh-water head is plotted versus time. In table 2 the mean fresh-water head of the six days period of measurements are indicated together with the elevation of the surface.

Table 2 - The mean fresh-water head deduced from the first series of measurements during a six days period

Piezometer	Distance in m from landmark positive in the sea-direction	Surface elevation m T.A.W.	Mean fresh-water head at -2 m T.A.W.
P1	360	+0,6	+2,96
P2	245	+1,8	+3,15
P3	165	+2,6	+3,46
P4	90	+3,1	+3,74
P5	5	+5,3	+4,59
P6	-70	+8	+4,55

These data show that the hydraulic head at the groundlevel is induced by the tidal mouvement of the seawater on the beach. Below the low-low-tide line the salt-water head at the surface of the seafloor equals the mean sea-level. At the fore shore the salt-water head equals the sealevel when that part of the beach is

1 The reference level of the fresh-water heads and the salt-water heads used in this paper is the 0 m T.A.W. level

inundated. When the sea retreats the salt-water head is approximately equal to the elevation of the surface above the reference plane. The mean salt-water head at the surface of the beach can be deduced from the elevation of the surface and the tidal mouvement of the seawater. Because of the depth of the screen of the piezometer below the beach surface there is a slight difference between the hydraulic head measured and the one at the surface. At the seaward part of the fore shore at piezometer P1, the salt-water head at 2,5 m below the surface stood a little above the surface when the sea retreats, which indicates a seepage of saline water. In the piezometer P2, P3 and P4 the salt water head is about equal to or a little smaller than the elevation of the surface. When the sea inundates the beach the salt-water head in the piezometers is a little smaller than the sealevel. This difference increases towards the landward part of the fore shore. The difference is larger in piezometer P4 than in piezometer P2. The difference causes an infiltration of seawater at the upper part of the beach. Under the uppermost part of the beach, between the high-tide line of neap tide and the highest sealevel elevation at spring tide during western or northern storms, the hydraulic head is almost stable and is not influenced by the tidal mouvement . Only in periods when the sea inundates that part of the beach seawater percolated through the unsaturated zone, of which the thickness varies between 0 and 1,5 m;this causes a rise of the hydraulic head under this part of the beach. This is confirmed by the hydraulic head measurements in piezometer P5 during the first two days of measurements with stormy weather. After the retreat of the sea, the hydraulic head decreases very slowly. Under the dune area the hydraulic head just below the watertable does not fluctuate under the unfluence of the tides.

The mean fresh-water head (fig. 4 , tab. 2) indicates that the horizontal hydraulic head gradient is very small underneath the dune. There is even a small landward hydraulic gradient resulting from the high waters of the stormy weather. There is a steep gradient underneath the back shore and the landward part of the fore shore. Further to the lower part of the beach the horizontal hydraulic gradient drops to zero below the low tide line. The pattern of the mean hydraulic head at the surface of the beach, which is deduced here, is a boundary condition which defines the mean flow of salt, brackish and fresh water underneath the beach. The pattern of the mean hydraulic head depends on the amplitude of the tidal movements of the sea and the topography of the beach. The other boundary conditions are the seaward flow of the fresh water underneath the dunes and the hydraulic head pattern in the aquifer underneath the sea. These boundary conditions, and the horizontal and vertical permeability are the input data of the mathematical model which describes the flow of fresh and salt water underneath the beach.

A second series of hydraulic head measurements was made to deduce the mean hydraulic head in the lower part of the aquifer underneath the beach. In the same plane of the resistivity profile K0 two piezometers, P2' and P4', were placed at respectively 265 and 160 m from the landmark in the direction of the sea. The diameter of the piezometer was 63 mm and the length 2 m. During a period

of three days the pressure was measured every ten minutes and converted into a fresh-water head which was printed and recorded on a magnetic tape. During the same period these measurements were also made in the piezometers P2, P5 and P6 of the first series. In fig. 5 the fresh-water head is plotted versus time, table 3 shows the mean fresh-water head of the three days period and the elevation of the surface.

Table 3 - The mean fresh-water head deduced from the second series of measurements

Piezometer	Distance in m landmark positive towards the sea	Surface elevation in m T.A.W.	Mean fresh-water head
			at -22 m T.A.W.
P2'	265	+1,7	+3,5
P4'	95	+2,6	+4,1
			at -2 m T.A.W.
P2	245	+1,8	+2,78
P5	5	+5,3	+4,39
P6	-70	+8	+4,47

With this series of measurements it was possible to follow the hydraulic head underneath the seaward part of fore shore in the upper part of the aquifer (piezometer P2) and in the lower part of the aquifer (piezometer P2'). The mean fresh-water head is about 0,7 m higher in piezometers P2' than in piezometer P2. The maxima of the hydraulic head are reached at approximately the same time. The minima in piezometer P2' are reached earlier than in piezometer P2. The difference between the maximum and minimum is larger in piezometer P2 than in piezometer P2'. During the measurements this difference varied between 2 and 2,35 m in piezometer P2', in piezometer P2 between 2,40 and 2,70 m.

The comparison of the measurements in piezometer P4 of the first series and those in piezometer P4' of the second series shows that the mean hydraulic head is about the same in the upper part and in the lower part of the aquifer. The maxima are reached a little later. The minimum in piezometer P4' is reached earlier than in piezometer P4. The difference between the maximum and minimum varies during the measurement between 1,0 and 1,5 m. The results of this second series of measurements can be a helpful tool in the calibration of the mathematical model.

CHEMICAL ANALYSES OF GROUNDWATER

After the hydraulic head measurements groundwater has been sampled from the piezometers. Three samples were taken at the -2 m T.A.W.-level in piezometer P2, P4 and P5 and two at the -22 m T.A.W.-level in the piezometer P2' and P4'. The results of the analyses have been indicated on the resistivity profile K0 (fig. 6) : total salt content, anions (Cl^-, SO_4^{--}, HCO_3^-) and kations (Na^+, K^+, Mg^{++}) are

given in mg/l.

The mean concentration of the anions and kations of the fresh water underneath the dune belt was deduced from the chemical analyses of 27 groundwater samples (L. LEBBE, 1978). The mean Cl^- concentration of the fresh water underneath the dune belt is 32 mg/l; the mean Cl^- concentration of the infiltrating seawater on the beach can be estimated at 15.700 mg/l.

From these data and supposing that there is no exchange of Cl^--ions with the deposition one can deduce the mixture of these two primary waters in every piezometer.

The Cl^- concentration in P4' being approximately 91 mg/l one may deduce that the groundwater sample represents a mixture of 99,62 % fresh water and 0,38 % seawater. The Cl^- concentration in P2' is about 512 mg/l so that this sample is a mixture of 96,9 % fresh water and 3,1 % seawater. The Cl^- concentration in P5 is about 3 650 mg/l so that this groundwater sample is a mixture of 77 % of fresh water and 23 % of seawater. According to the well logging the piezometers P2' and P4' are both located in fresh water but one can see that the salinity increases towards the sea. The brackish water of piezometer P5 confirms the well logging data.

The groundwater samples taken in piezometers P2 and P4 at small depth underneath the beach have approximately the same kation and anion composition as the infiltrating seawater.

MATHEMATICAL MODEL

Theory

The flow of groundwater in an aquifer filled with water of different densities is defined by the generalized Darcy equation in vector form :

$$q_i = -\frac{k'_i}{\mu_i} (\nabla p_i + \rho_i g) \qquad (1)$$

where i is an index which defines any point in the groundwater reservoir

q	the seepage velocity	(LT^{-1})
k'	the intrinsic permeability	(L^2)
μ	the dynamic viscosity	$(ML^{-1}T^{-1})$
ρ	the density	(ML^{-3})
∇p	the pressure gradient	$(ML^{-2}T^{-2})$
g	gravitational acceleration	(LT^{-2})

The relation between the hydraulic conductivity and the intrinsic permeability can be written as :

$$k_i = \frac{k'_i \rho_i g}{\mu_i} \qquad (2)$$

where k is the hydraulic conductivity (LT^{-1})

The relation between the pressure and the fresh-water head can be represented as follows :

$$\varphi_{if} = z_i + p_i/\rho_f g \qquad (3)$$

where φ_{if} is the fresh-water head (L)
z_i the elevation of point i above the reference plane (L)
p_i the fluid pressure in point i ($ML^{-}T^{-2}$)
ρ_f the density of the fresh water (ML^{-3})

With the relations (1), (2) and (3) one can determine the horizontal and the vertical seepage velocity in a point i of the groundwater reservoir containing waters of different densities, if the fresh-water head and the density in that point are given.

$$q_{hi} = k_{hi}\frac{\rho_f}{\rho_i}\frac{\partial \varphi_{if}}{\partial x} \qquad (4)$$

$$q_{vi} = k_{vi}\frac{\rho_f}{\rho_i}\left(\frac{\partial \varphi_{if}}{\partial z} + \frac{(\rho_i - \rho_f)}{\rho_f}\right) \qquad (5)$$

where q_{hi} is the horizontal seepage velocity (LT^{-1})
q_{vi} the vertical seepage velocity (LT^{-1})
k_{hi} the horizontal hydraulic conductivity (LT^{-1})
k_{vi} the vertical hydraulic conductivity (LT^{-})

Because of the lack of accuracy by which the horizontal and vertical hydraulic conductivity can be determined the factor ρ_f/ρ_i, which is small, is ignored in the development of the model in this paper.

The law of continuity for steady state flow can be written as :

$$\nabla q = 0 \qquad (6)$$

Finite-difference approximation

The groundwater flow is considered to be two-dimensional in a vertical plane. On this plane a grid is superposed (fig. 7). The intersections are called "nodes". To each node a value for the fresh-water head $\varphi(I,J)$ is assigned. The node I,J is numbered in such a way that the coordinates are given : $x_{I,J} = I \times L$ and $y_{I,J} = J \times H$. The mean difference of the density of the water between the nodes I,J+1 and I,J and the density of fresh water divided by the fresh-water density is given by the symbol VD(I,J) and can also be called the mean areal buoyancy.
By means of a central finite-difference approximation one can determine the horizontal and the vertical seepage velocity.

$$q_h(I,J) = k(J)(\varphi_f(I+1,J) - \varphi_f(I,J))/L \qquad (7)$$

$$q_v(I,J) = (\varphi_f(I,J+1) - \varphi_f(I,J) + VD(I,J)H)/c(J) \qquad (8)$$

where the density of fresh water is supposed to equal 1 g/cm³

$q_h(I,J)$	is the horizontal seepage velocity from the node I+1,J to the node I,J	(LT^{-1})
k(J)	the horizontal hydraulic conductivity of layer J	(LT^{-1})
$\Psi_f(I,J)$	the fresh-water head in node I,J	(L)
L	the horizontal distance between two adjacent nodes of the same layer	(L)
$q_v(I,J)$	the vertical seepage velocity from the node I,J+1 to the node I,J	(LT^{-1})
c(J)	the hydraulic resistance between the nodes of the J+1-layer and the nodes of the J-layer	(T)
VD(I,J)	the mean difference of the density of the water between the nodes I,J+1 and I,J and of the fresh water density divided by the fresh-water density or mean areal buoyancy (dimensionless)	
H	the vertical distance between two adjacent nodes of the same column	(L)

With the equation (7) and (8) one can write the continuity law at the node I,J.

$$K(J)H(\Psi_f(I+1,J)-2\Psi_f(I,J)+\Psi(I-1,J))/L$$
$$+ L(\Psi_f(I,J+1)-\Psi_f(I,J)+VD(I,J)H)/c(J)$$
$$-L(\Psi_f(I,J)-\Psi_f(I,J-1)+VD(I,J-1)H)/c(J-1) = 0 \qquad (9)$$

Boundary conditions

The groundwater reservoir is bounded below by an impermeable layer. Because there is no flow through this boundary there is no need for a water quality boundary condition. At the landward vertical boundary we assume a constant horizontal flow of fresh water with a density of 1 g/cm³.

The upper boundary of the groundwater reservoir is partially situated underneath the dunes and partially underneath the beach and the sea. Underneath the dunes we assume a constant vertical flow boundary. The vertical flow equals the mean annual infiltration rate of fresh water in the dune area. Beneath the beach and the sea one assumes a constant hydraulic head boundary. These hydraulic heads have been deduced from measurements on the beach. Where on the beach a downward vertical flow occurs one assumes that there is an infiltration of salt water with a density of 1,025 g/cm³.

The seaward vertical boundary is a constant hydraulic head boundary. On this boundary the salt-water head is held the same over the whole depth. The density of the water which flows through this boundary depends on the flow direction. When the flow is landward through the boundary into the considered area we assume that this inflowing water has the density of seawater. When water flows from the considered area in the direction of the sea the density of the water is defined by the flow of salt and fresh water in the mathematically considered area.

Control of the iteration process

The finite-difference approximation leads to a system of linear algebraic equations. This system is solved by an alternating direction implicit iterative method. In the vertical section initially the water is considered to be of equal density. The accuracy of the calculation of the flow in the system is checked by establishing a water balance of the system. When the iteration process converges the difference between the inflow and the outflow of the system decreases. When the calculations of the hydraulic head and consequently of the flow have reached a sufficient accuracy the matrix VD is deduced. This is done by following the stream line which forms the limit between the flow of the fresh and salt water. The lower end of the fresh-/salt-water boundary is found by following the streamline which begins at the intersection of the reservoir floor and the landward or seaward vertical boundary. The upper salt-/fresh-water boundary can be found by following the streamline which begins at the high-high-tide line. At the landward part of this line fresh water infiltrates and at the seaward part salt water. When a fraction f_s of the area between two nodal points I+1,J and I,J is occupied by salt water and the complementary fraction $(1-f_s)$ is occupied by fresh water the value of the mean areal buoyancy VD(I,J) is calculated as follows :

$$VD(I,J) = f_s \, (\rho_s - \rho_f)/\rho_f \qquad (10)$$

The matrix VD being determined one now computes the fresh-water head configuration and consequently the flow anew. After reaching a sufficient accuracy the matrix VD is computed again. This algoritm is repeated until one instantly obtains a sufficiently accurate flow when applying the newly determined matrix VD. So we obtain a hydraulic head configuration where the flow is in agreement with the density distribution.

Input

The first series of input data determines the dimensions of the nodal grid and of the considered area. The dimensions of the grid are given by the number of colomn x and the number of layers y. The real dimensions of the considered area are defined by the horizontal distance between two adjacent nodes of the same layer, L, and by the vertical distance between two adjacent nodes of the same column, H. The number of the columns, situated underneath the dune area, x1, defines the position of the high-high-tide line.

The hydraulic conductivity of the aquifer is given by an array of horizontal hydraulic conductivities for each layer K(J), where J varies between 1 and y, and by an array of hydraulic resistances between each layer c(J) where J varies between 1 and y-1. The hydraulic resistance c(1) is the hydraulic resistance between layer 1 and 2.

For the boundary conditions one must define an array of horizontal seepages Q(J) where J varies between 1 and y-1. The infiltration N is provided by a constant vertical seepage through the upper boundary of the column underneath the dunes. The boundary condition on the beach is a constant hydraulic head pattern. This pattern of fresh-water heads is deduced from the hydraulic head measurements. This is also done for the vertical constant hydraulic head boundary under the seafloor.

Output

As a result of the calculation we obtain a hydraulic head distribution where the resulting flow is in agreement with the density distribution. For each nodal point one has a value for the fresh-water head and a value for the mean density of the water between the nodal points. These results are represented by lines of equal fresh-water head and stream lines on the vertical plane of the considered area. The horizontal axis is taken positive towards the sea and the origin is located at the high-tide line. The vertical axis is positive in the upward direction and the origin corresponds with the 0 m T.A.W.-level.

The lines of equal fresh-water head are deduced by an interpolation between the value of the fresh-water head of four nodal points within the area of these points.

The constant flow boundary conditions are indicated by a letter, Q, for the horizontal fresh water flow of the dunes in the direction of the sea and N for the vertical fresh water flow equal at the infiltration rate in the dune area. The constant head boundary conditions are represented by the value of the constant fresh water head. The inflow and outflow through the permeable boundary of the mathematically treated area are calculated and represented near these boundaries. The stream lines are drawn in such a way that they begin at the permeable boundary with an inflow of fresh or salt water. The arrows on the streamlines indicate the even years that a waterdrop is underway on the streamline from the beginning at the boundary of the considered area. For the last calculation a water-conducting porosity of 0,30 is taken into account.

Examples of fresh-/salt-water distribution as calculated by the mathematical model

The same nodal grid is used in the three examples. The grid consists of 25 columns and 18 layers (x = 25, y = 18). Each column is 30 m wide and each layer is 1,5 m thick (L = 30 m, H = 1,5 m). The lower boundary is an impermeable substratum. The first column contains the hydraulic heads resulting from the constant flow boundary condition. The next 4 columns have an upper constant flow boundary. The constant flow equals the infiltration rate in the dune area (N = 270 mm/year or N = $7{,}39.10^{-4}$ m/d). In this case a dune area of 120 m width is considered. In the mathematical model the beach has a width of 420 m. This is represented by 14 columns with an upper constant hydraulic head boundary. The fresh-water heads have been deduced from the pattern of measured hydraulic heads just below the beach surface. This pattern is a function

of the tidal fluctuation of the sealevel and the topography of the beach. Finally 150 m of offshore is considered. This is represented by 5 columns with an upper constant hydraulic head boundary. These constant fresh-water heads are calculated from the salt-water head on the seabottom based on the mean sealevel. The last column contains the constant hydraulic head boundary condition. The salt-water head being the same over the entire depth the fresh-water head consequently increases with a value of 0,025 m per metre increasing depth.

The horizontal hydraulic conductivities are the same for all layers, K(J) = 10 m/d for J = 1 to 18 as well as the hydraulic resistances between the layers, c(J) = 15 d for J = 1 to 17. The horizontal flow through the vertical boundary is the same for every layer. Only the sum of this horizontal flow differs in the three examples.

In the first example the sum of the horizontal flow of fresh water through the vertical boundary is 0,25 m^2/d (fig. 8). Because of the considerable seaward flow there is a horizontal hydraulic gradient towards the sea underneath the dune area. Underneath the back shore and the upper part of the fore shore salt water infiltrates. It forms a lens which flows upon the fresh dune water. On the largest part of the fore shore the salt-water flow is directed vertically upward resulting in an outflow at the beach surface. The fresh water only flows out at the seabottom.

In the second example the sum of the horizontal flow of fresh water through the vertical boundary is strongly reduced, namely 0,05 m^2/d (fig. 9). Consequently the horizontal seaward hydraulic gradient underneath the dunes. There the vertical hydraulic gradient becomes more important. The upper salt-water lens is enlarged in depth as well as in width. The fresh-water outflow is consequently further offshore on the seabottom. The zone of this fresh-water flow is narrower than in the first example. The lower salt-water tongue has retreated towards the sea. The flow pattern and the fresh-/salt-water distribution are in good agreement with the field data of fresh-water heads and the resistivity profiles K1, K2, K3 before the natural reserve. There is still a fresh water flow towards the sea.

In the third example the sum of the fresh-water flow through the vertical boundary is reversed, namely -0,09 m^2/d (fig. 10). Nearly all the infiltrating fresh water in the dune area is flowing towards this boundary. Only a small part of the infiltrating fresh water flows towards the sea. On the upper part of the beach seawater infiltrates and fills the whole aquifer under the beach. In spite of the small fresh-water flow towards the sea there is still a tongue of fresh water underneath the back shore. This is confirmed by the resistivity logging in front of the urban area of De Panne (K3 and K4).

This two-dimensional model can be used to describe the flow of fresh and salt water under a part of the dunes,the beach and a part of the offshore. Through sensitivity analyses one can deduce how the different parameters defines the flow and distribution of

the salt and fresh water under the beach. These parameters are the hydraulic head pattern just below the beach surface, the flow of fresh water towards the sea, the horizontal and vertical permeability distribution.

In the mathematical model the process of hydrodynamic dispersion was not incorporated until now. It is in our intention to do so in the future. The former field measurements and chemical analyses of the groundwater can be used to evaluate the longitudinal and transversal dispersion coefficients.

The seaward flow of fresh water and the hydraulic head pattern underneath the beach are time dependent. The seaward fresh-water flow is considerable larger after a period of successive wet years than after a prolonged period of dry years. For example at the end of the dry summer 1976. The horizontal hydraulic gradient was landward over the whole length of the fore dunes. This was only during a short period. At the end of spring 1981, after a number of wet years, there was a seaward hydraulic gradient over the whole length of the dunes although this gradient was many times larger before the natural reserve than before the urban area . The pattern of the hydraulic head underneath the beach can strongly change. North-western to northern storm causes a supplementary rise of the sealevel during the whole period of the storm. These results in the increase of salt-water infiltration on the back shore and the upper part of the fore shore. The upward outflow of salt water on the largest part of the fore shore diminished or reversed. So during these period the salt water amount of the upper salt-water lens can increase. The reverse can also take place during a south-eastern to southern storm which result in a drop of the sealevel during this period. This south-eastern to southern storms are at the Belgian coast restricted in occurrence.

Because of the possible large variation in time of the two important factors which define the distribution of fresh, brackish and salt water there is a need for mathematical models which describe the unsteady-state flow of water with waters of different qualities. Continuous measurements of the time dependent factors are needed for the calibration of such models.

Such studies will indicate the relative importance of the different factors which define the variation of the salt, brackish and fresh water distribution like the hydrodynamic dispersion, the change of the fresh-water flow and of the hydraulic head pattern underneath the beach. This knowledge is necessary for a good management of groundwater reservoirs with water of different qualities like the one in the Belgian coastal plain.

Practical considerations

It is clear that the danger for salt-water intrusion under the seaward part of the dunes comes first of all from the upper salt water lens. In this case it is not from the lower salt water tongue that the danger of salt water intrusion will arise. The location of the lower salt water tongue is defined by the mean sea level and the dispersion (COOPER et al., 1964). The salt water of the upper lens is close to the dunes. The salt-water head is there higher than the mean sealevel. This salt-water head depends on the height and the duration of the high waters and the permeabi-

lity of the subsurface of the back shore.

At slightly inclining coast with a permeable surface and subsurface and a meaningfull tidal fluctuation an upper salt-water lens can be expected. This can be a potential danger of salt water intrusion. Because of this salt-water infiltration on the beach the lower salt-water tongue in a thick aquifer will be more seaward than can be deduced from older studies which ignore the upper salt-water lens and flow.

In thick aquifers where fresh water wells are placed near the beach one must know the distribution and flow of salt, brackish and fresh water. The distribution consists of an upper salt-water lens an upper salt-/fresh-transition zone, a fresh-water zone, a lower fresh-/salt-transition zone and a lower salt-water tongue. The distribution of the horizontal and vertical permeabilities with the hydrodynamic dispersion coefficients define further the flow to the wells of the waters of different qualities.

CONCLUSION

Resistivity logging in the Western Belgian beach has revealed different salt-/fresh-water distributions. The unconfined aquifer with a thickness of about 30 m consists of permeable layers of fine to medium fine sand with discontinuous semi-permeable lenses of very silty fine sands. Above the fresh groundwater from the dunes which flows seawards, a salt water lens is present. The salt water infiltrates on the back shore and the upper part of the fore shore. Under the lower part of the fore shore the salt water flows upwards resulting in an outflow. The fresh water under the salt water gradually becomes more mineralised. Underneath the low-tide line it is nearly brackish.

In front of the urban area of De Panne, at K3 and K4, the fresh groundwater flow is very much reduced because of the decrease in infiltration and the influence of the water catchment. Only a small fresh-water tongue occurs in the lower part of the aquifer under the back shore. Under the fore shore the whole aquifer is occupied by salt water. The resistivity of the lower part of the aquifer being slightly higher than the upper part one can deduce that the salinity in the upper part of the aquifer is a little higher than in the lower part.

The occurrence of this salt-fresh water distribution underneath the Belgian beach can be explained by means of a two-dimensional mathematical model, taking the density differences into account. One of the boundary conditions which defines the flow of salt and fresh water is the pattern of the mean hydraulic head at the surface of the beach. Hydraulic-head measurements have shown that this pattern depends on the amplitude and the topography of the beach. Because the amplitude and the topography are approximately the same along the resistivity profiles the differences in fresh-/salt-water distribution are mainly due to the difference in seaward fresh-water flow from the dunes. A second, but minor reason for the changes of the salt-/fresh-water distribution under the beach is the occurrence of discontinuous semi-pervious lenses at different levels.

So far hydrodynamic dispersion has not been included in the model. To do so in the future, field measurements and chemical analyses will be used to evaluate the longitudinal and transversal dispersion coefficients.

Some of the parameters which define the salt-/fresh-water distribution under the beach are time dependent. The variability of these time-dependent parameters requires repeated observations and the construction of mathematical models for unsteady flow.

A better knwoledge of all parameters defining the flow will contribute to a better management of reservoirs containing groundwater of differing quality.

ACKNOWLEDGEMENT

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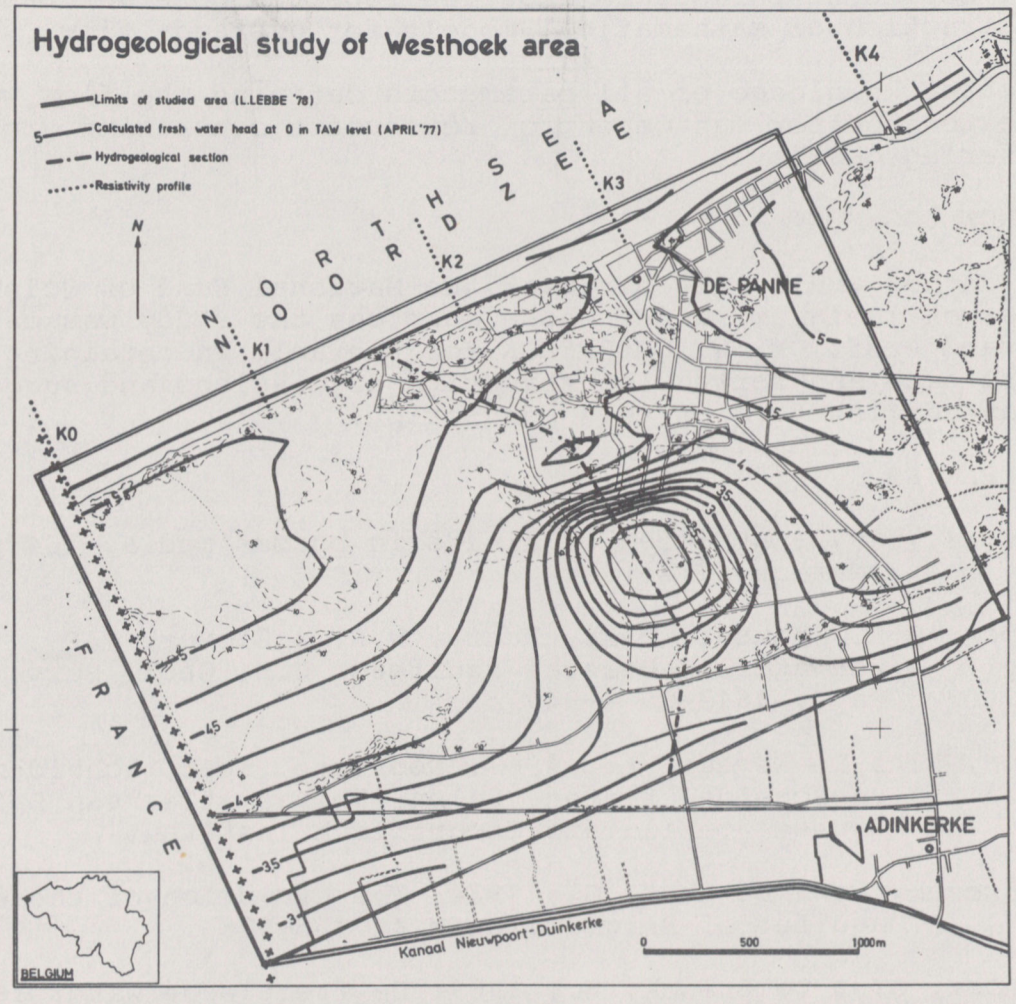


Fig. 1 - Location of the hydrogeological study in the Westhoek area with calculated lines of equal hydraulic head, of the hydrogeological section and of the resistivity logging profiles

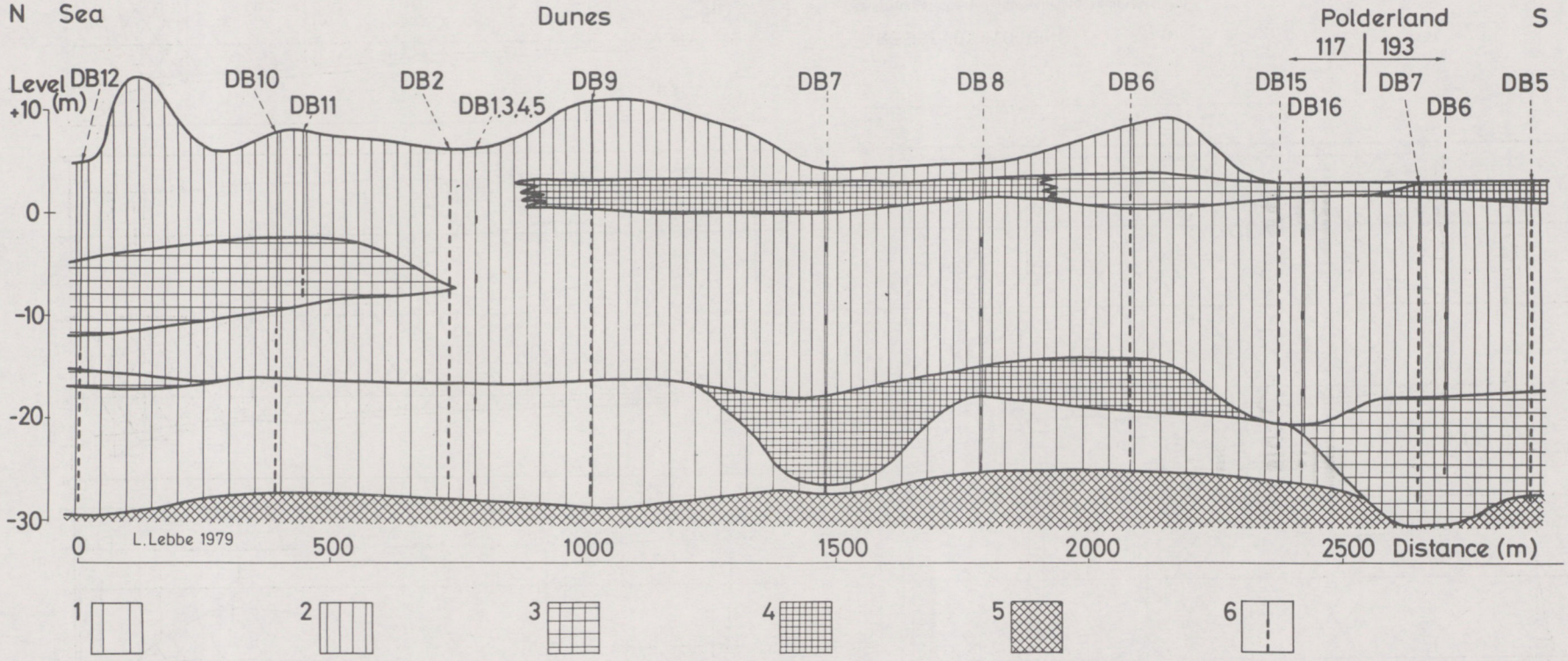


Fig. 2 - North-south hydrogeological section through the Westhoek area (L. LEBBE, 1979)

1. permeable layer with large horizontal hydraulic conductivity
2. permeable layer with small horizontal hydraulic conductivity
3. semi-permeable layer with small hydraulic resistance
4. semi-permeable layer with large hydraulic resistance
5. impermeable layer
6. well screen

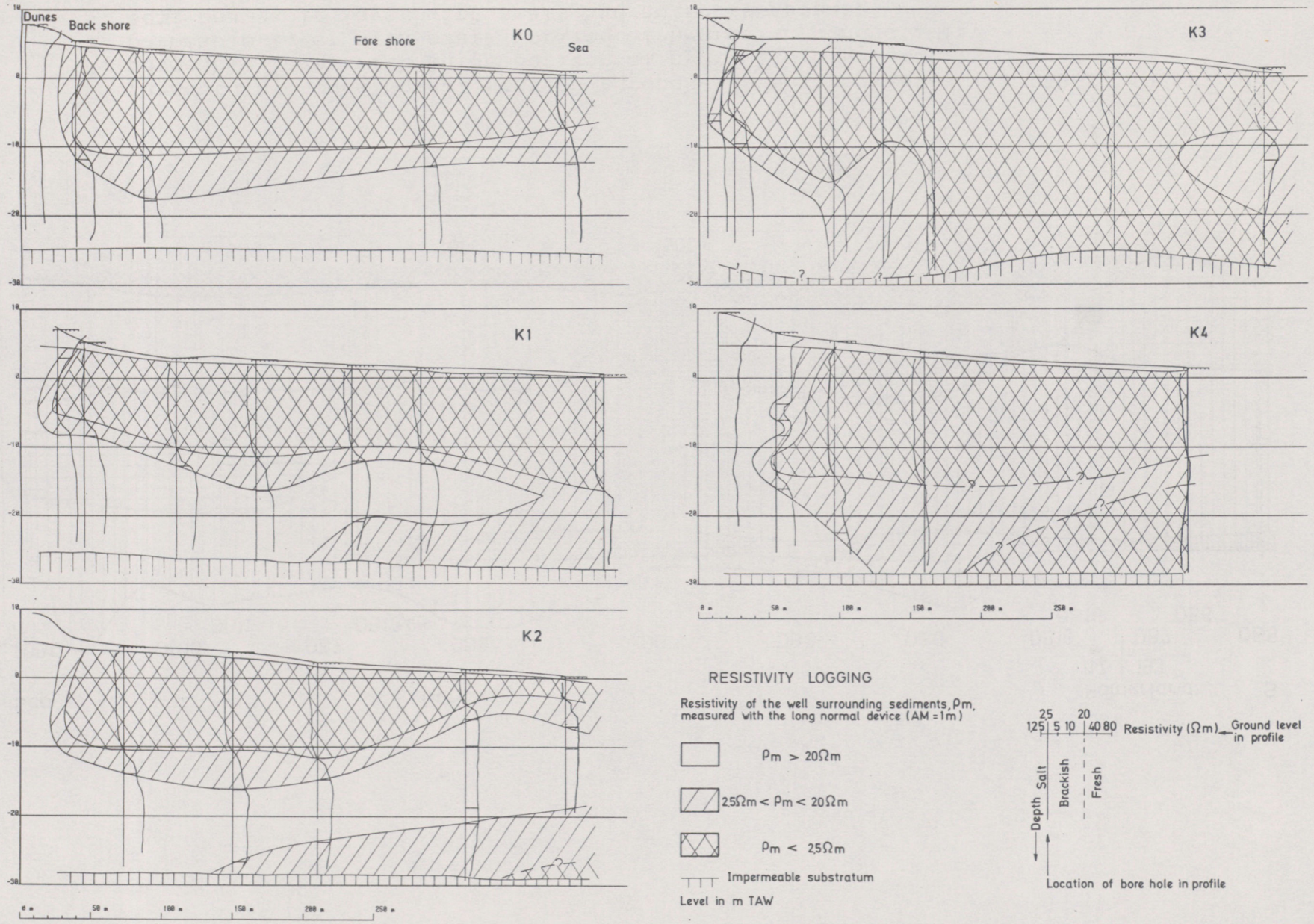


Fig. 3 - The resistivity profiles perpendicular to the shore line of the Westhoek area

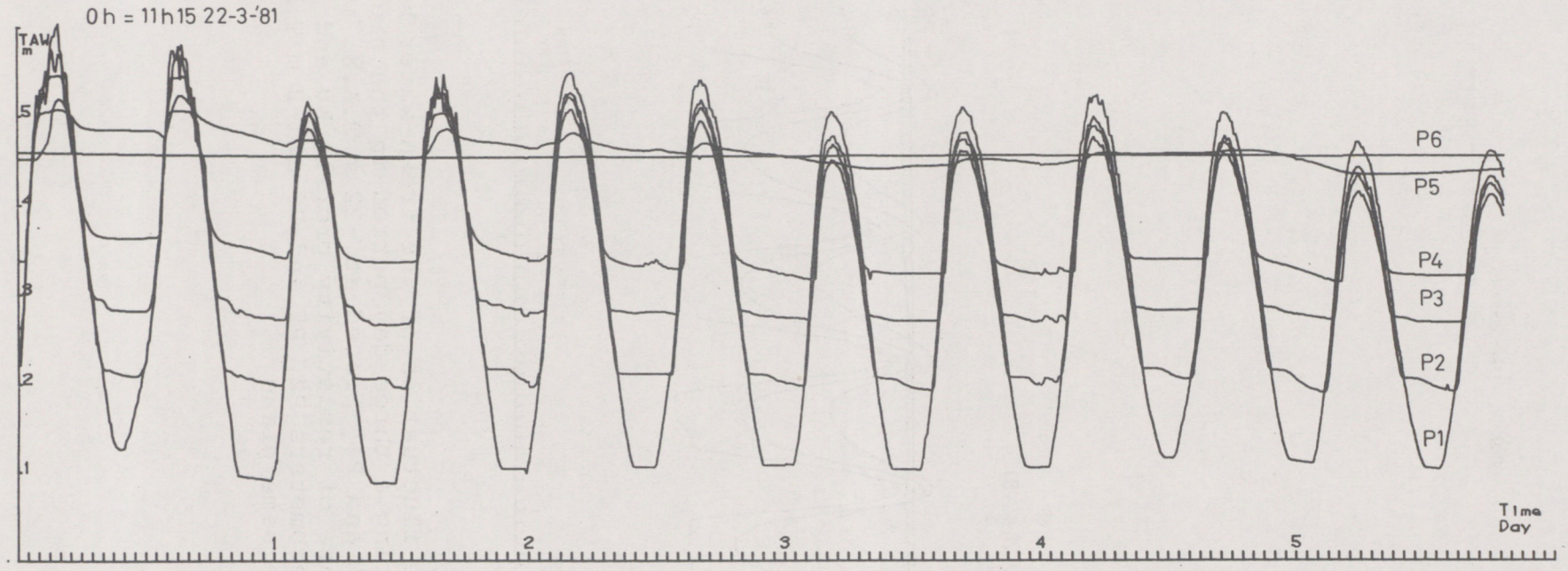


Fig. 4 - The fluctuation of the fresh-water head during a six-day period in the piezometers P1, P2, P3, P4, P5 and P6 placed at -2 m T.A.W. in the plane of resistivity profile K0

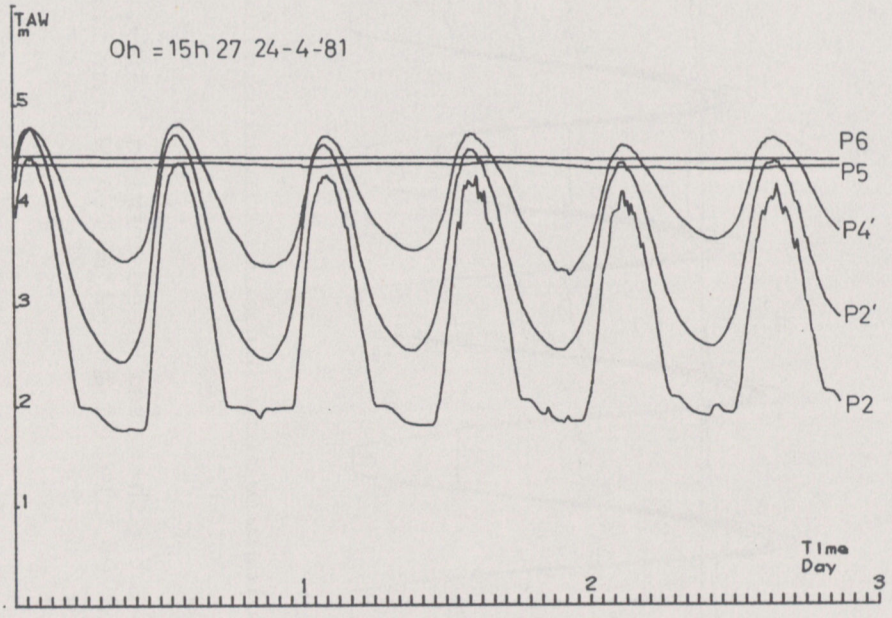
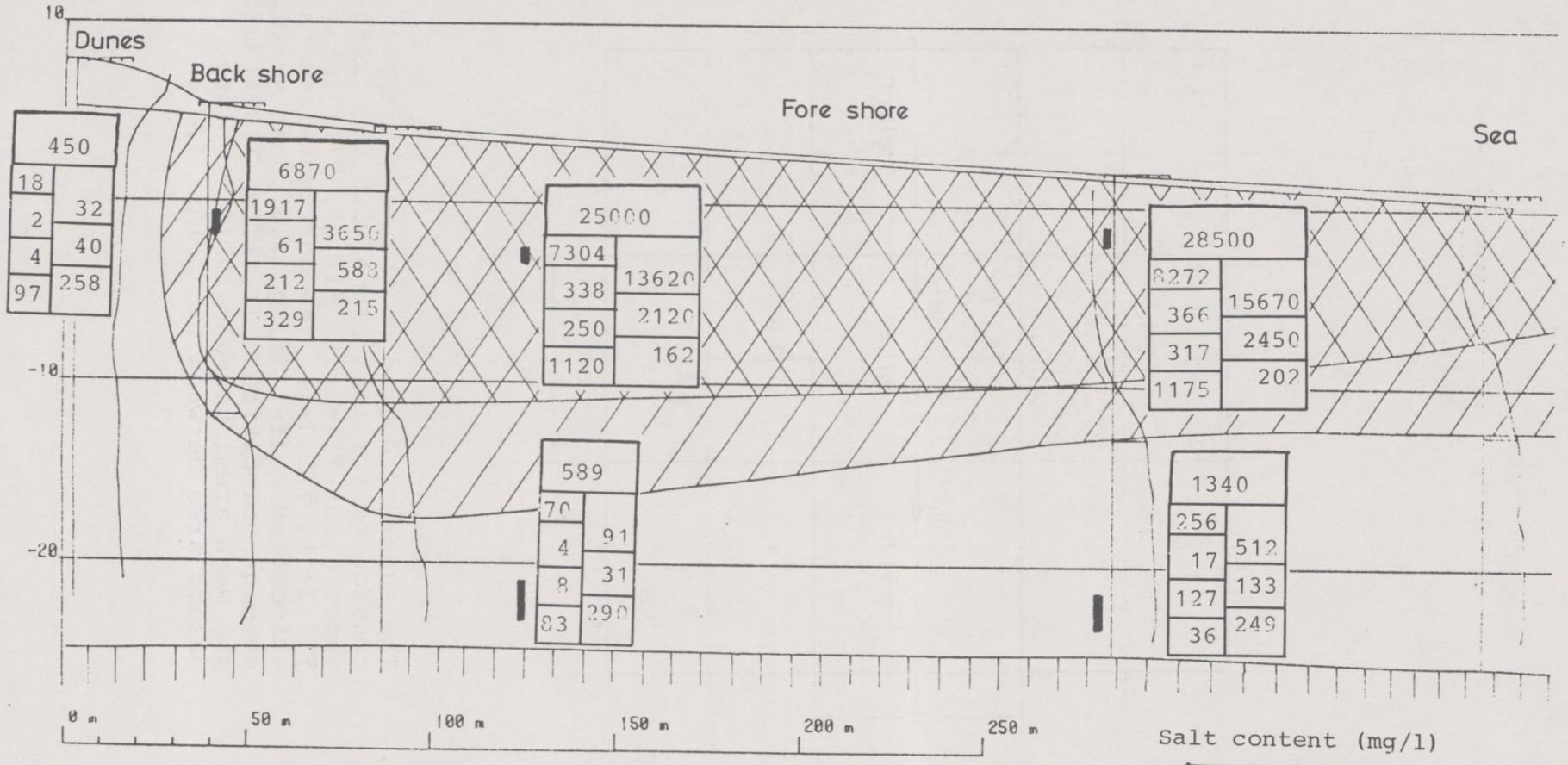


Fig. 5 - The fluctuation of the fresh-water head during a three-day period in the piezometers P2' and P4' placed at -22 m T.A.W. in the plane of resistivity profile K0 and in the piezometers P2, P5 and P6 at -2 m T.A.W. in the same plane



RESISTIVITY LOGGING

Resistivity of the well surrounding sediments, ρ m, measured with the long normal device (AM = 1 m)

- $\rho m > 20\ \Omega m$
- $2,5\ \Omega m < \rho m < 20\ \Omega m$
- $\rho m < 2,5\ \Omega m$

CHEMICAL ANALYSES

Location of screen pumped for collecting groundwater sample

Impermeable substratum

Level in m T.A.W.

Salt content (mg/l)

Total	
Na^+	Cl^-
K^+	SO_4^{--}
Mg^{++}	HCO_3^-
Ca^{++}	

Fig. 6 - The results of the chemical analyses of groundwater samples collected in the plane of resistivity profile K0

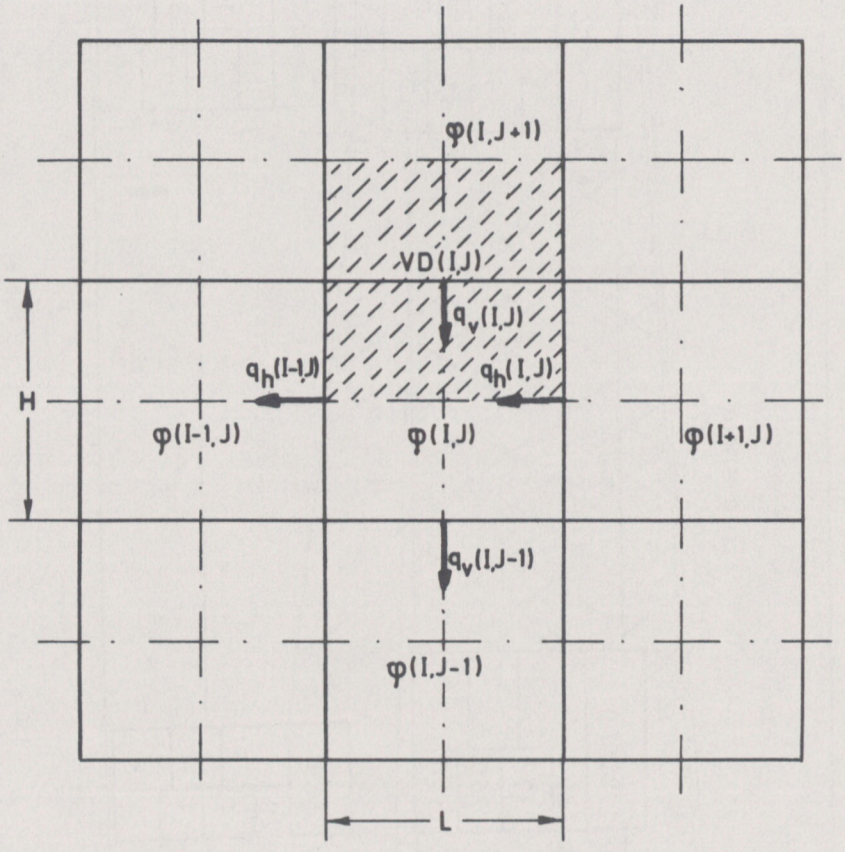


Fig. 7 - The superposed grid on the mathematically considered vertical plane with the freshwater heads $\Phi(I-1,J)$, $\Phi(I,J)$, $\Phi(I+1,J)$, $\Phi(I,J+1)$, $\Phi(I,J-1)$, the horizontal seepage velocities $q_h(I-1,J)$, $q_h(I,J)$, the vertical seepage velocities $q_v(I,J)$, $q_v(I,J-1)$ and the mean areal bouyancy VD(I,J) of the dashed area between the points I,J and I,J+1

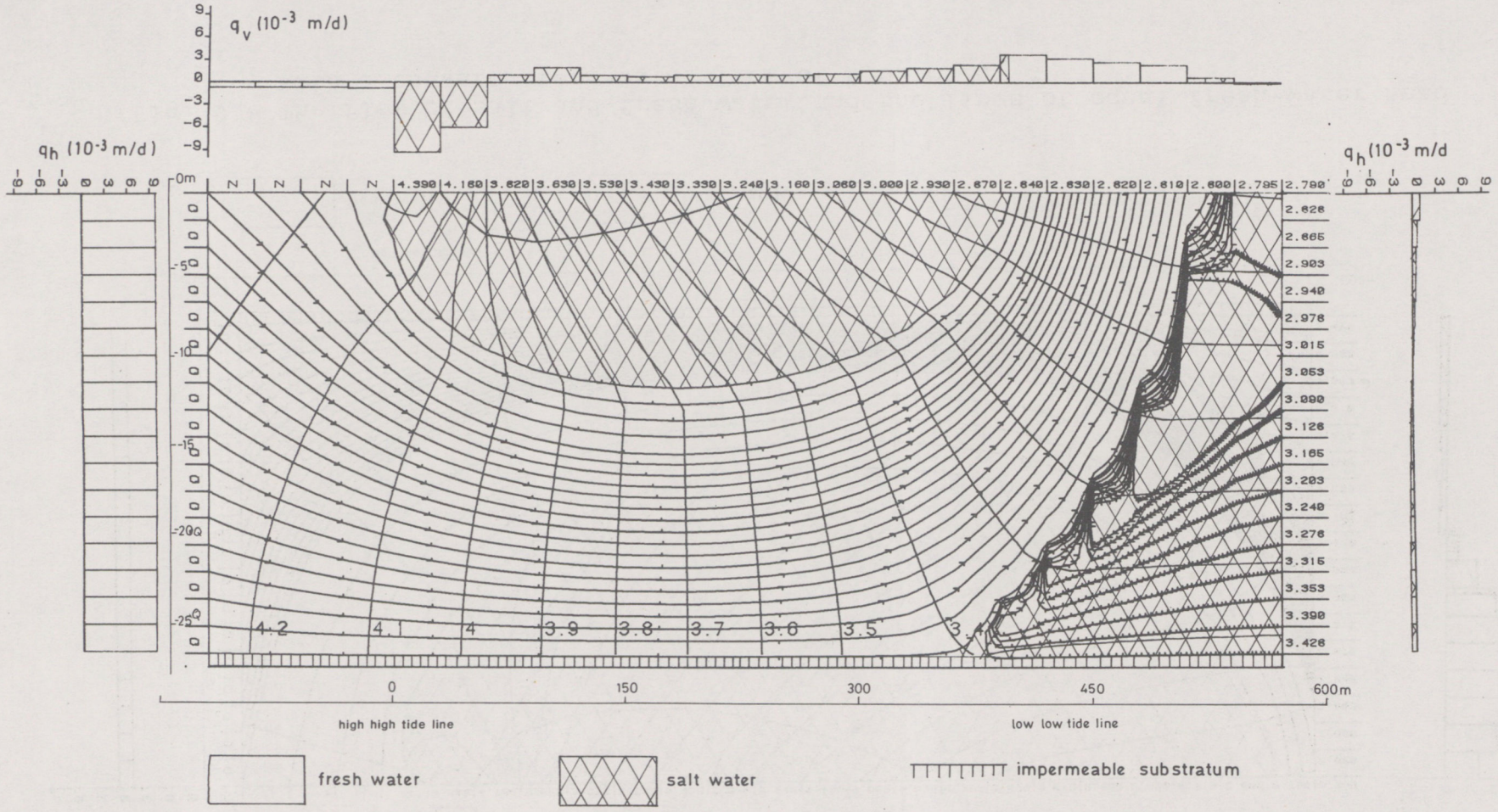


Fig. 8 - The flow of salt and fresh water and the lines of equal fresh-water head with a seaward fresh-water flow of 0,25 m²/d

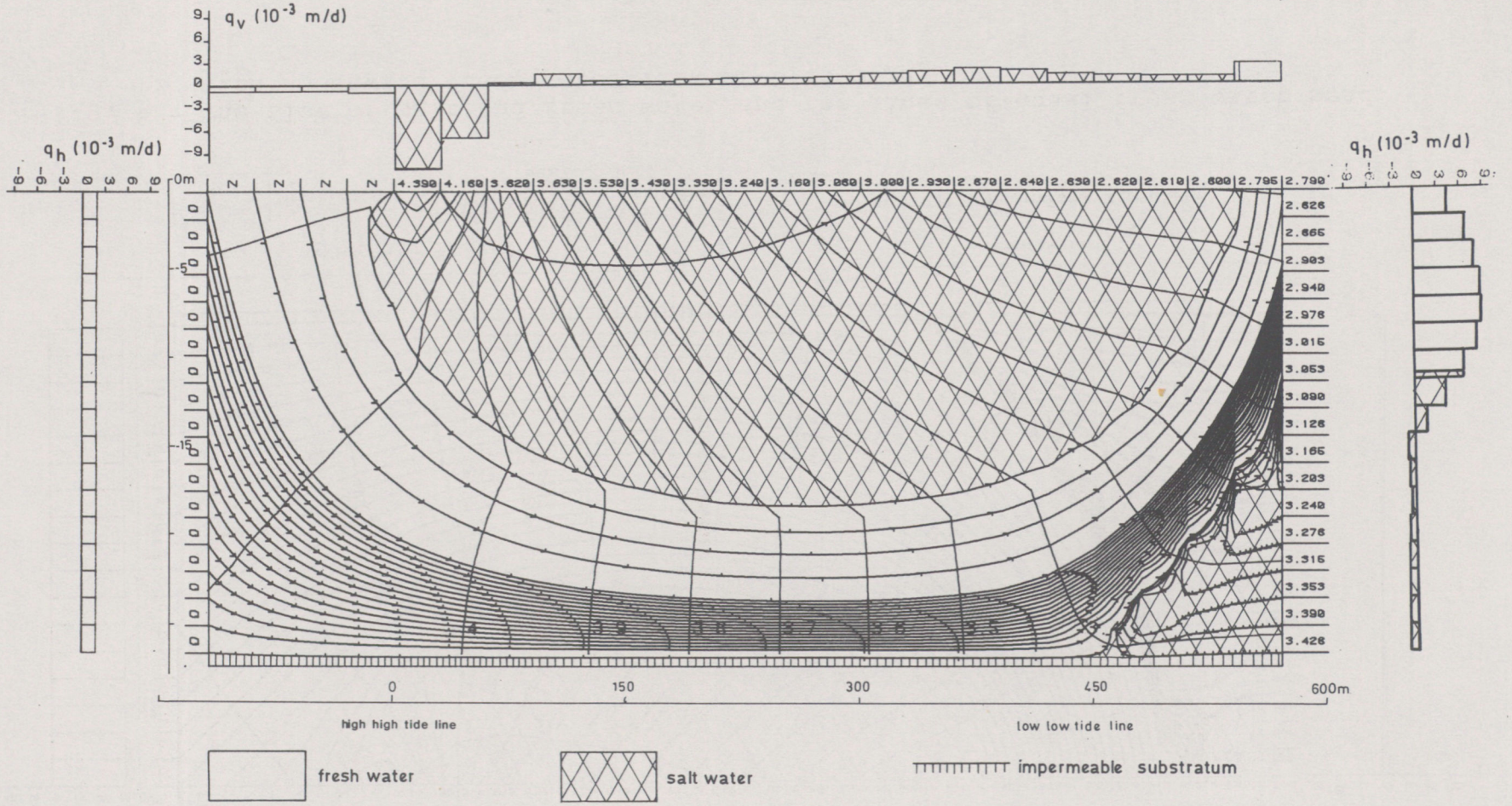


Fig. 9 - The flow of salt and fresh water and the lines of equal fresh-water head with a seaward fresh-water flow of 0,05 m²/d

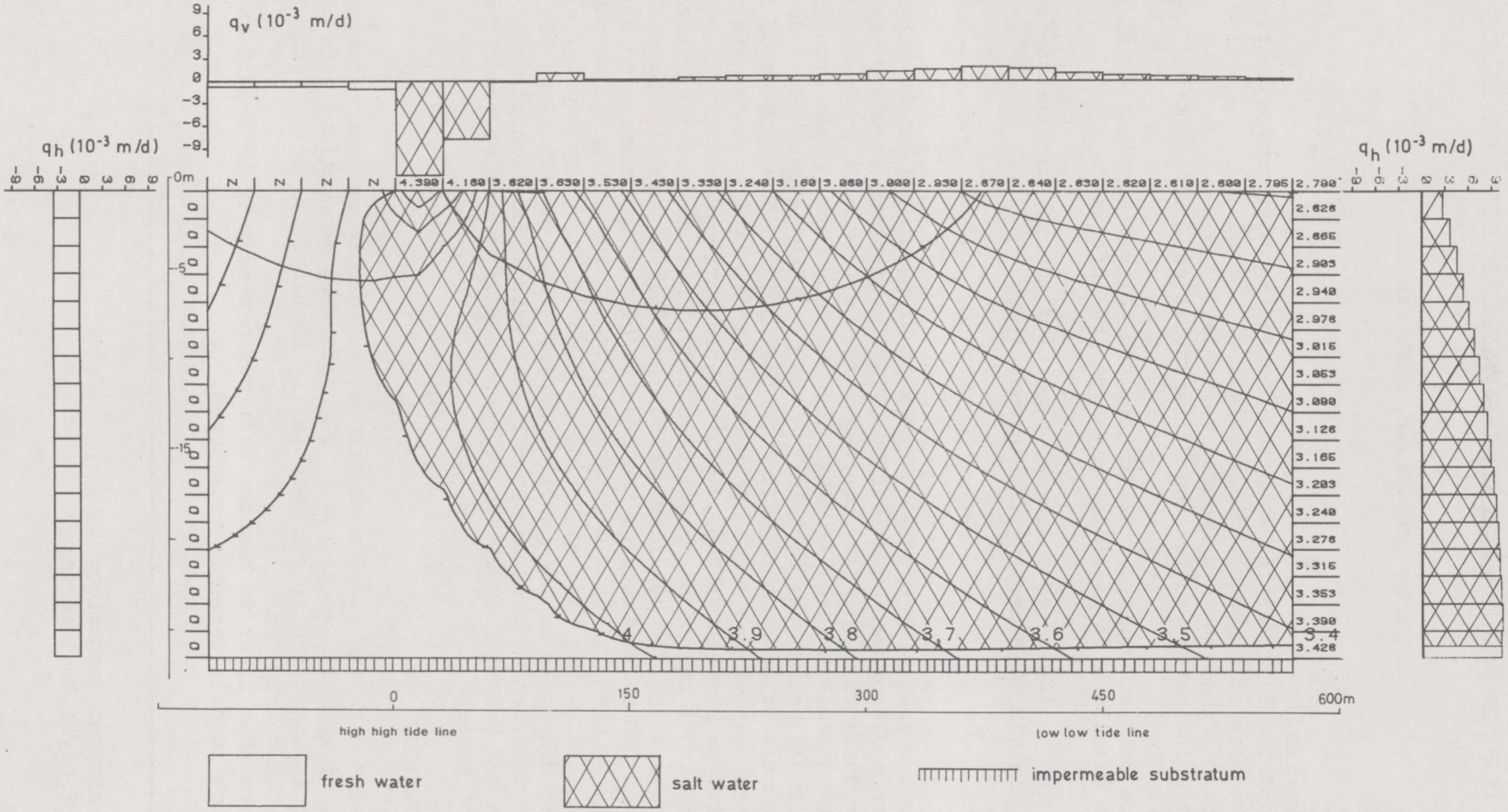


Fig. 10 - The flow of salt and fresh water and the lines of equal fresh-water head with a landward fresh-water flow of 0,09 m^2/d

ANALYSIS OF THE POSSIBLE SHAPES OF THE FRESH WATER-SALT WATER INTERFACE IN A SEMI-CONFINED AQUIFER WITH AXIAL-SYMMETRIC BOUNDARY CONDITIONS

by prof.dr.ir. J.C. van Dam[1]

ABSTRACT

The paper comprises:
- derivation of the differential equation describing the problem mentioned in the title in the case of stagnant saline water;
- introduction of dimensionless reduced variables;
- a numerical solution method for the differential equation in terms of the reduced variables;
- presentation of the complete results obtained for all possible axial-symmetric cases,i.e.a catalogue of possible shapes of the interface.

The paper is the follow-up of a similar paper for the one-dimensional case presented in the 6th SWIM-meeting in Hannover 1979.

INTRODUCTION

This paper is the follow-up of the author's paper presented at the 6th Salt Water Intrusion Meeting, Hannover 1979 (van Dam, 1981). Like the previous paper the present paper deals with the flow of fresh groundwater above stagnant saline groundwater in a semi-confined aquifer. The previous paper dealt with the case of one-dimensional horizontal flow in the aquifer. The present paper describes the case of horizontal axial-symmetric flow in the aquifer. The subject of the previous paper has since been worked out in more detail in two companion papers (Sikkema and van Dam, 1982)(van Dam and Sikkema, 1982) concentrating respectively on the analytical solutions and on an approximate solution method of the relevant differential equation. The paper by van Dam and Sikkema contains practical examples and an analysis of the accuracy of the results of the approximate solution method. The reader of the present paper may profit from reading the previous papers as well for two more reasons. The reduced variables used in the present paper were first introduced and discussed in the previous papers (van Dam, 1981)(van Dam and Sikkema, 1982). The 13 different curves obtained by analytical solution of the one-dimensional case (Sikkema and van Dam, 1982) can be seen as limit-cases of the curves described in the present paper.

PROBLEM FORMULATION

This paper deals with the axial-symmetric geohydrologic profile of figure 1. The vertical axis of rotation is at r = 0. The following assumptions were made:
- the aquifer is homogeneous; not necessarily isotropic as only horizontal flow in the aquifer is considered;
- the semi-permeable layer is homogeneous; not necessarily isotropic as only vertical flow in this layer is considered;

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- the interface between fresh and saline groundwater is sharp; diffusion and dispersion are ignored;
- both the fresh and the saline groundwater are homogeneous (i.e. of constant densities);
- the saline groundwater is stagnant; this implies that the fresh water flows in steady state;
- the phreatic groundwater table is horizontal.

The depth D is not relevant for the mathematical formulation of the problem, except as a boundary. Yet -for practical reasons- D is chosen equal to the aquifer thickness (van Dam, 1981)(van Dam and Sikkema, 1982).

The Dupuit approach of describing the flow in the aquifer as horizontal flow can be justified as long as the slope of the interface is not too steep (e.g. < 0.02 to 0.05). This is often the case in large scale problems. Similarly the flow in the semi-permeable toplayer is described as vertical flow, which is mostly justified because the permeability of this layer is often so much smaller than that of the aquifer.

The meaning and dimension of the symbols, most of which occur in figure 1, is:
- geohydrological constants:
  k = aquifer permeability $[LT^{-1}]$
  c = hydraulic resistance of the semi-permeable layer $[T]$
  D = thickness of the aquifer $[L]$
  $\alpha = (\rho_s - \rho_f)/\rho_f$ $[-]$
  ρ_s = density of the saline groundwater $[ML^{-3}]$
  ρ_f = density of the fresh groundwater $[ML^{-3}]$
- boundary conditions:
  p = elevation of the phreatic groundwater table, horizontal, above reference level*) $[L]$
  h_s = piezometric level of the stagnant saline groundwater, above reference level*) $[L]$
  E = equilibrium depth, see expression (5) $[L]$
- variables:
  h_f = piezometric level of the flowing fresh water in the aquifer, above reference level*) $[L]$
  H = thickness of the fresh water in the aquifer $[L]$
  Q = flow of fresh groundwater in the aquifer through a cylindrical surface of radius r $[L^3T^{-1}]$
  r = distance from the centre $[L]$

EQUATIONS

The problem is described by the following three equations:

Darcy's law: $$Q = -2\pi r H k \frac{dh_f}{dr} \qquad (1)$$

continuity equation: $$\frac{dQ}{dr} = -2\pi r \frac{h_f - p}{c} \qquad (2)$$

Badon-Ghijben/Herzberg's principle: $$h_f - h_s = \alpha(h_s + H) \qquad (3)$$

*) the reference level has been chosen at the top of the aquifer

The three basic equations (1), (2) and (3) contain three unknowns Q, H and h_f, variables of the independent variable r. When eliminating the variables Q and h_f one easily arrives at the following differential equation in H:

$$rH \frac{d^2H}{dr^2} + r\left(\frac{dH}{dr}\right)^2 + H \frac{dH}{dr} - r \frac{H-E}{kc} = 0 \qquad (4)$$

, where $E = \frac{p-(1+\alpha)h_s}{\alpha}$ (5)

E is a function of p, h_s and α. The constants p, h_s and α occur in the differential equation in the term E only. E is called the equilibrium depth because it is that particular value of H for which the corresponding value of the piezometric level of the fresh groundwater h_f equals p, so that no vertical flow occurs in the semi-permeable layers.

So far, the author knows only two particular solutions of differential equation (4) viz.:

$$H = E \qquad (6)$$

$$H = 2E + \frac{r^2}{8kc} \qquad (7)$$

A general solution has, to the best of his knowledge, not yet been found. In order to overcome this difficulty a good approximation can be formulated by replacing the continuity equation (2) by:

$$\frac{dQ}{dr} = -2\pi r \frac{\overline{h_f}-p}{c} \qquad (8)$$

, where $\overline{h_f}$ is the average value of the variable h_f over a small reach of r.

Integration of (8) yields:

$$Q = Q_o - \pi(r^2-r_o^2) \frac{\overline{h_f}-p}{c} \qquad (9)$$

The unknown Q can be eliminated by combination of the equations (1) and (9):

$$-2\pi rHk \frac{dh_f}{dr} = Q_o - \pi(r^2-r_o^2) \frac{h_f-p}{c} \qquad (10)$$

h_f and $\overline{h_f}$ can be eliminated from (10) when using equation (3) and the corresponding equation:

$$\overline{h_f} - h_s = \alpha(h_s+\overline{H}) \qquad (11)$$

, where $\overline{H}$ is the average value of the variable H over that same reach.

After this elimination and introduction of the equilibrium depth E the following differential equation is found:

$$rH\frac{dH}{dr} - (r^2 - r_o^2) \frac{\overline{H} - E}{2kc} + \frac{Q_o}{2\pi k\alpha} = 0 \qquad (12)$$

This differential equation describes the variable H as a function of r over a small reach of r starting from $r = r_o$. In $r = r_o$, $Q = Q_o$ and $\overline{H}$ is the estimated average value of H over the considered reach. The estimated value $\overline{H}$ is subject to revision after each calculation of H.

REDUCED DIMENSIONLESS VARIABLES

For convenience it is attractive to introduce the following reduced, dimensionless, variables

$$y = H/D \qquad (13)$$

$$u = r/\lambda \quad \text{, where } \lambda = \sqrt{kDc} \tag{14}$$

$$q = Q/\pi k \alpha D^2 \tag{15}$$

and the reduced, dimensionless, constant

$$\varepsilon = E/D \tag{16}$$

Differential equation (4) then reduces to:

$$uyy'' + u(y')^2 + yy' - u(y-\varepsilon) = 0 \tag{17}$$

Differential equation (12) then reduces to:

$$uyy' - \tfrac{1}{2}(u^2-u_o^2)(\bar{y}-\varepsilon) + \tfrac{1}{2}q_o = 0 \tag{18}$$

Equation (9) then reduces to:

$$q = q_o - (u^2 - u_o^2)(\bar{y}-\varepsilon) \tag{19}$$

SOLUTION

The solution of differential equation (18) was first found by van der Molen (van der Molen, about 1975). In a somewhat modified denotation his solution is:

$$y = \pm\sqrt{\tfrac{1}{2}(\bar{y}-\varepsilon)(u^2-u_o^2) - \{(\bar{y}-\varepsilon)u_o^2 + q_o\} \ln u/u_o + y_o^2} \tag{20}$$

, where y_o is the value of y at $u = u_o$.
The estimated value of $\bar{y}$ can now be checked by

$$\bar{y} = (y + y_o)/2 \tag{21}$$

and, if necessary, revised. This iterative procedure can be avoided when substituting (21) in expression (20):

$$y = \pm\sqrt{\frac{y+y_o-2\varepsilon}{4}(u^2-u_o^2) - \{(\frac{y+y_o-2\varepsilon}{2})u_o^2 + q_o\} \ln \frac{u}{u_o} + y_o^2} \tag{22}$$

or:

$$y^2 - (\frac{u^2-u_o^2}{4} - \frac{u_o^2}{2} \ln \frac{u}{u_o})y - \left[\frac{y_o-2\varepsilon}{4}(u^2-u_o^2) - \{u_o^2(\frac{y_o-2\varepsilon}{2}) + q_o\} \ln \frac{u}{u_o} + y_o^2\right] = 0 \tag{23}$$

This is a quadratic equation. The explicit solution for y is:

$$y = \frac{-B \pm\sqrt{B^2 - 4C}}{2} \tag{24}$$

where $B = -(\frac{u^2-u_o^2}{4} - \frac{u_o^2}{2} \ln \frac{u}{u_o})$ (25)

and $C = -\left[\frac{y_o-2\varepsilon}{4}(u^2-u_o^2) - \{u_o^2(\frac{y_o-2\varepsilon}{2}) + q_o\} \ln \frac{u}{u_o} + y_o^2\right]$ (26)

This enables us to proceed straight forward from $u = u_o$, where $y = y_o$ and $q = q_o$, to $u = u$. For the next step a new value of y_o and a new value of q_o is needed; y_o for the next step is the y found with (24) for the present step and q_o for the next step is the q at $u = u$ found with (19) after substitution of (21):

$$q = q_o - (u^2-u_o^2)(\frac{y+y_o}{2} - \varepsilon) \tag{27}$$

, where y_o and y respectively are the values at the beginning and the end of the present step.

RESULTS

Calculations have been made with a programmable pocket calculator HP67 using the expressions (24) and (27) for y and q. Some details of the programme will be dealt with later, together with the tricks applied to overcome some numerical problems for the calculator. After gaining some experience with the programme a survey was made of the possible shapes of the curves.

On account of the results obtained for the one-dimensional case (van Dam, 1981) (Sikkema and van Dam, 1982) (van Dam and Sikkema, 1982) and of understanding of the geohydrological problem it may be taken that each possible curve has q = 0 for at least one value of u. So one may expect to find all possible shapes when taking q = 0 respectively at:

u = 0
u = small, e.g. u = 1
u = great, e.g. u = 100

This holds for each of the three classes $\varepsilon > 0$, $\varepsilon = 0$ and $\varepsilon < 0$. So nine different cases occur as in the table below.

	q = 0 at		
	u = 0	u = 1	u = 100
$\varepsilon > 0$	figure 2	figure 3	see van Dam (1981) and Sikkema and van Dam (1982)
$\varepsilon = 0$	figure 4	figure 5	
$\varepsilon < 0$	figure 6	figure 7	

Three of them, namely those for q = 0 at u = 100 are practically the same as for the one-dimensional cases described in previous papers (van Dam, 1981) (Sikkema and van Dam, 1982). This is obvious because for the great values of u the curvature of the cylinder is negligibly small. The numerical results of the calculations confirmed this. Therefore in the present paper only the remaining 6 cases are dealt with. A survey of the possible shapes in each of these six cases is given in the six figures 2 through 7, the numbers of which have been indicated in the table.

A discussion of the curves in each of these six figures follows now. The description of the curve types is chosen in accordance with the terminology chosen for the one-dimensional case by van Dam and Sikkema (1982).

Figure 2. $\varepsilon > 0$ is chosen at $\varepsilon = + 0.4$.
q = 0 at u = 0 for respectively $y_{q=0} = + 1.0; + 0.8; + 0.5; + 0.35; - 0.2; - 0.5$ and - 1.0. The first three of these curves ($y_{q=0} > \varepsilon$) are of a mutually similar shape, the reverse U-type. The same holds for the last three curves ($y_{q=0} < 0$), the S-type. The curve with $y_{q=0} = + 0.35$ is one of the U-type which occurs for $0 < y_{q=0} < \varepsilon$. The curve with $y_{q=0} = + 0.8$ is the particular solution, the parabola.

Figure 3. $\varepsilon > 0$ is chosen at $\varepsilon = + 0.4$.
q = 0 at u = 1 for respectively $y_{q=0} = + 0.6; + 0.35; - 0.2; - 0.228$ and - 0.3. The curve with $y_{q=0} = + 0.6$ is one of the reverse U-type which is found for $y_{q=0} > \varepsilon$. The curves with $y_{q=0} = + 0.35$ and - 0.2 are both of the U-type which is found for $- 0.228 < y_{q=0} < \varepsilon$. The curve with $y_{q=0} = - 0.228$ is a particular one; the upper asymptotic type, as it goes asymptotically to $y = \varepsilon$. A lower asymptotoc type curve is also drawn in this figure. It goes from below to that same asymptote. The upper and lower asymptotic type curves have been found by trial and error; the starting conditions were iteratively changed until the asymptotic behaviour was reached. The curve with $y_{q=0} = - 0.3$ is one of the S-type which occurs for $y_{q=0} < - 0.228$.

Figure 4. $\varepsilon = 0$.
$q = 0$ at $u = 0$ for respectively $y_{q=0} = + 0.4$; 0 and - 0.4. Only two curve-types were found, namely the reverse U-type for $y_{q=0} \geq 0$ and the S-type for $y_{q=0} < 0$.
Figure 5. $\varepsilon = 0$.
$q = 0$ at $u = 1$ for respectively $y_{q=0} = + 0.4$; + 0.05 and - 0.4. Only two curve-types were found, namely the reverse U-type for $y_{q=0} > 0$ and the S-type for $y_{q=0} < 0$.
Figure 6. $\varepsilon < 0$ is chosen at $\varepsilon = - 0.4$.
$q = 0$ at $u = 0$ for respectively $y_{q=0} = + 0.4$; 0; - 0.2; - 0.6; - 0.8; - 1.0 and - 1.2. The curves with $y_{q=0} = 0$ and $y_{q=0} = - 0.8$, the parabola, are of the intersecting type, separating the three types for which:

$0 < y_{q=0}$ reverse U-type
$2\varepsilon < y_{q=0} < 0$ oscillating type (new term)
and $y_{q=0} < 2\varepsilon$ S-type

Figure 7. $\varepsilon < 0$ is chosen at $\varepsilon = - 0.4$.
$q = 0$ at $u = 1$ for respectively $y_{q=0} = + 0.4$; 0; - 0.2; - 0.4; - 0.708 and - 1.0. The curves with $y_{q=0} = 0$ and $y_{q=0} = - 0.708$ are particular ones separating the three types:

$0 < y_{q=0}$ reverse U-type
$- 0.708 < y_{q=0} < 0$ oscillating type
and $y_{q=0} < - 0.708$ S-type

The value $y_{q=0} = - 0.708$ was found by trial and error until that particular curve was found for which the intersection of the horizontal axis (at $u > 1$) was with a finite gradient ($q = 0$).

The reader may have observed that in all six figures the parts of the curves for $y < 0$ have been drawn in dotted lines and the parts for $y > 0$ in continuous lines. This is because only the parts for $y > 0$ make sense for the geohydrological problem considered in this paper.

The steeper parts of the continuous lines, though mathematically correct, must also be rejected from the geohydrological point of view for two reasons. First of all the Dupuit approach of only horizontal flow in the aquifer is not justified for the steeper parts. Moreover, there where the curves intersect the horizontal axis (top of the aquifer) vertically the water velocities would be infinite which is physically impossible. Among the curves that intersect the horizontal axis only those particular ones which intersect with a small angle are geohydrologically acceptable.

The reader will recognize parts of the continuous curves which can serve as solutions or parts thereof for cases as: abstraction wells or injection wells in the centre or in a circle around the centre of a polder, a circular polder bounded by one or several annular polders surrounded by another polder extending to infinity or by open water.

SOME REMARKS ON THE PROGRAMME AND ITS IMPLEMENTATION

The programme requires the following inputdata: u_o, $\Delta u (= u - u_o)$, y_o, q_o and ε and gives for each step Δu: u, y and q. These values enter automatically as u_o, y_o and q_o for the next step. It is of course also possible to change these values before starting the run of the next step. This can be useful for Δu, q_o and ε. A change in q_o corresponds to the input of an abstraction or injection at u_o (in reality a change in Q_o at x_o). A change in ε corresponds to the passage of a boundary (at r_o) between two polders with different polderlevels p. The stepsize Δu can always be adapted to the required degree of accuracy (related to the values

of y and y') and the desired speed of operation. In many cases a stepsize $\Delta u = \pm 0.2$ gave very accurate results. This statement is based on a test with the only curve which is exactly known, the parabola. Mostly a stepsize of ± 0.1 or $+ 0.05$ was applied. In the steeper parts of the curves even smaller values of $|\Delta u|$ have been used. For the curves with vertical passage of the horizontal axis too great values of $|\Delta u|$ gave "error" because the curve does not exist there as in expression (24) $B^2 - 4C < 0$. In such cases the stepsize $|\Delta u|$ was automatically reduced to half its original value until $B^2 - 4C > 0$. At the vertical passage of the horizontal axis it was always necessary to reverse the direction of progress i.e. to change the sign of Δu. This change is made automatically in the programme, not only for the sign of Δu, but also for the sign of y, once the value of $|y|$ is smaller than a prescribed small value (0.02 was chosen). The value of q_o remains unchanged in these reversals.

In some cases tricks had to be applied to run the programme. For example, at $u_o = 0$ the term $\ln u/u_o$, which occurs in expression (23), would become ∞. Therefore, instead of $u_o = 0$ a very small value, as $u = 10^{-99}$, was applied. It was already mentioned that for the greater values of u the results must be approximately equal to those obtained in the one-dimensional case. This was confirmed indeed by the numerical results obtained when starting with $u_o = 100$ (better than $u_o = 10$). For values of u_o in the order of 10^3 to 10^6, where the curvature of the cylinder is even smaller, the results were worse because the values of $\ln u/u_o$ in expression (23) became very small e.g. $\ln \frac{10^6 + 10^{-1}}{10^6} = \ln 1.0000001$, which led to loss of accuracy.

CONCLUSIONS

There exists a great variety of possible shapes for the interface between flowing fresh water above stagnant saline water in a semi-confined aquifer. For various practical problems sections of different curves can be used and interconnected so as to satisfy boundary conditions and connection conditions.

The one-dimensional case, presented earlier, can be seen as a special case of the axial-symmetric case.

A programme for an HP67 pocket calculator has been developed, and is available, for rapid straight forward calculation of interfaces. It starts from given or assumed values for the interface depth and the flow of fresh groundwater at one end and continues even over discontinuities, such as differences in polderlevel, abstractions and injections. In case certain conditions must be met at the other end the calculation must be repeated with other assumed values at the beginning (shooting) until the required conditions at both ends are met.

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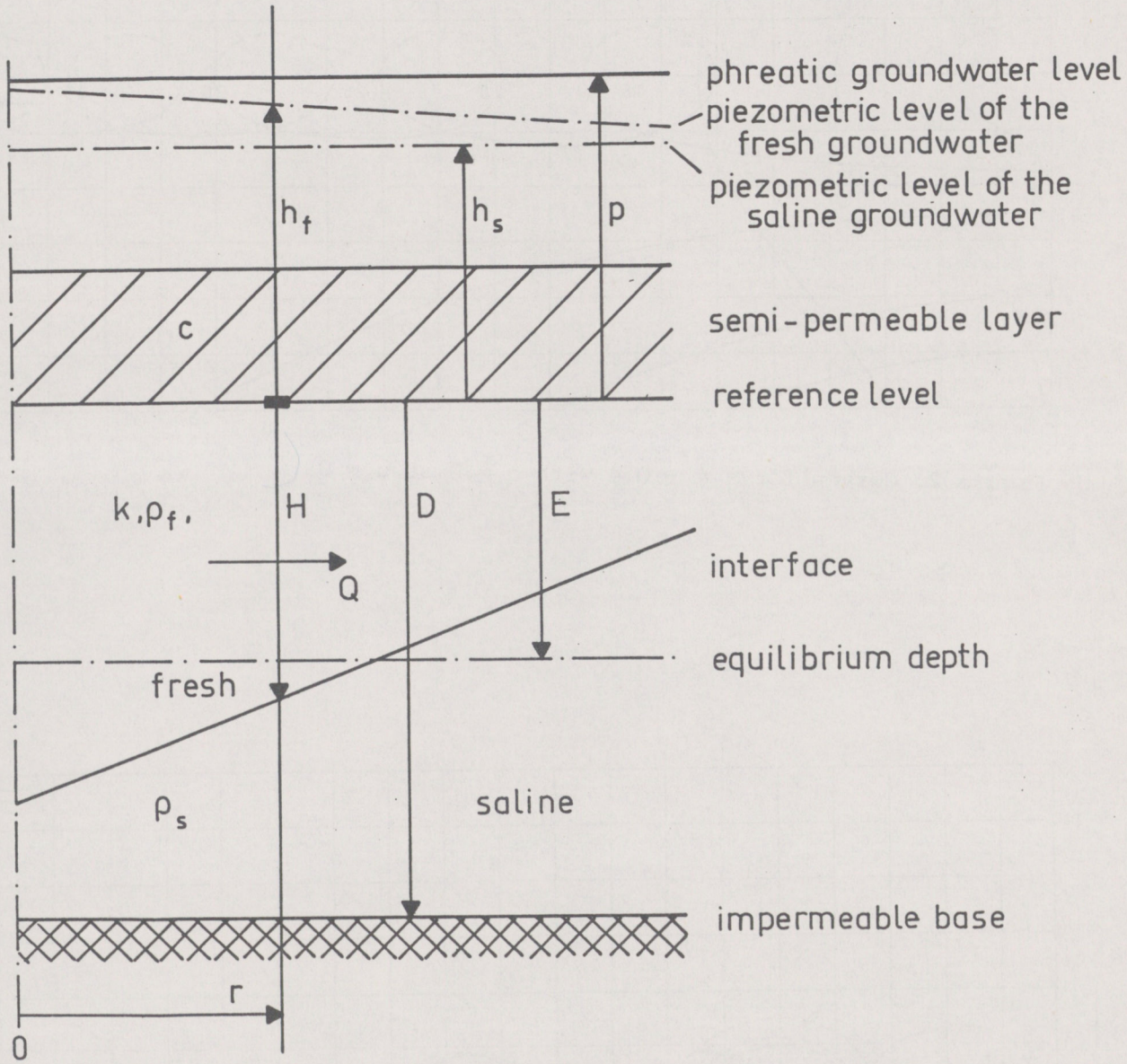


Figure 1. Geohydrologic profile

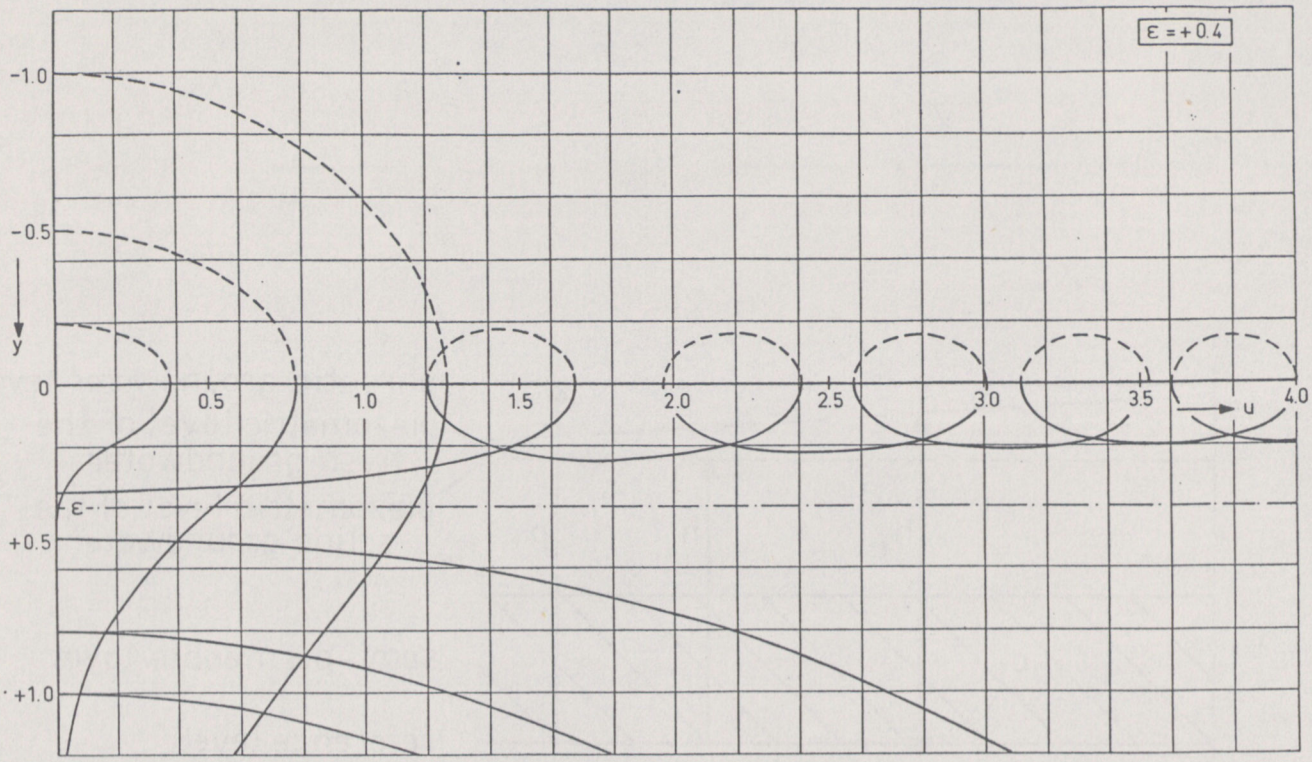


Figure 2. Curves for ε = + 0.4 with q = 0 at u = 0

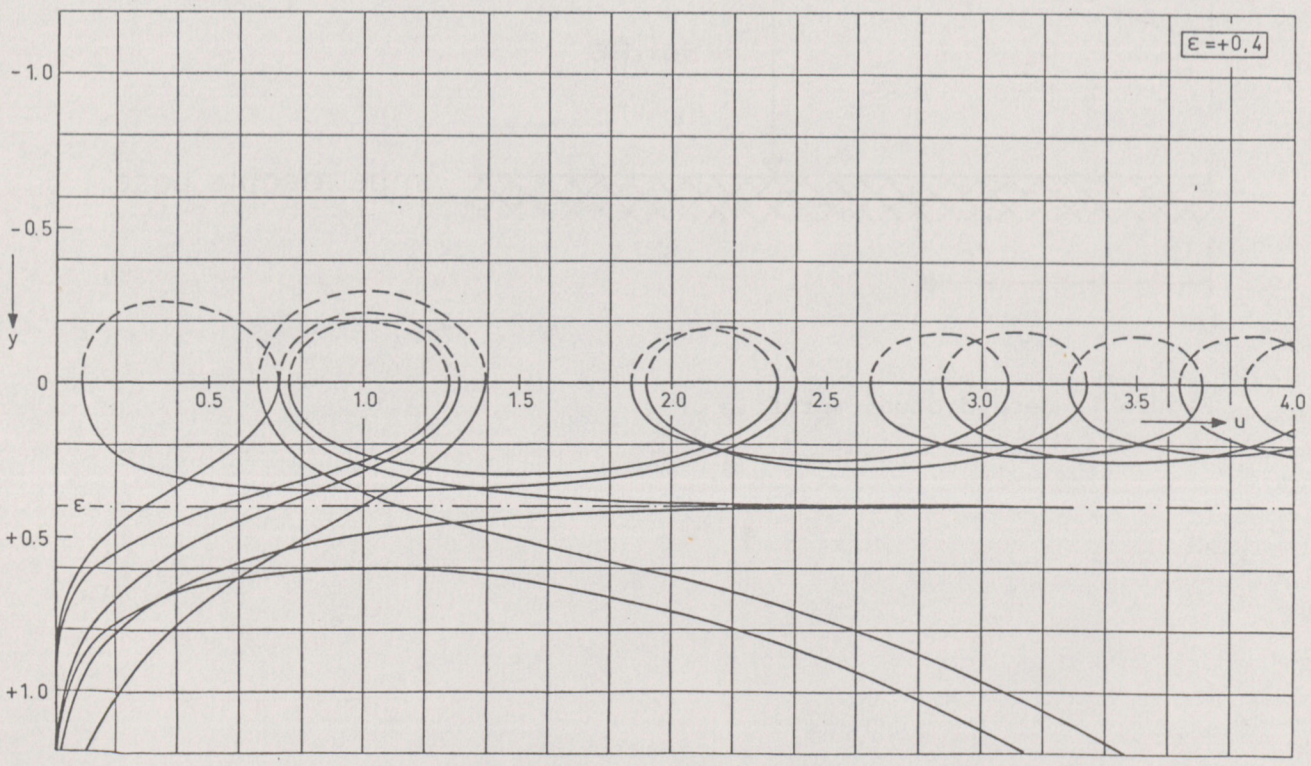


Figure 3. Curves for ε = + 0.4 with q = 0 at u = 1
and with q = 0 at u = ∞, y = ε

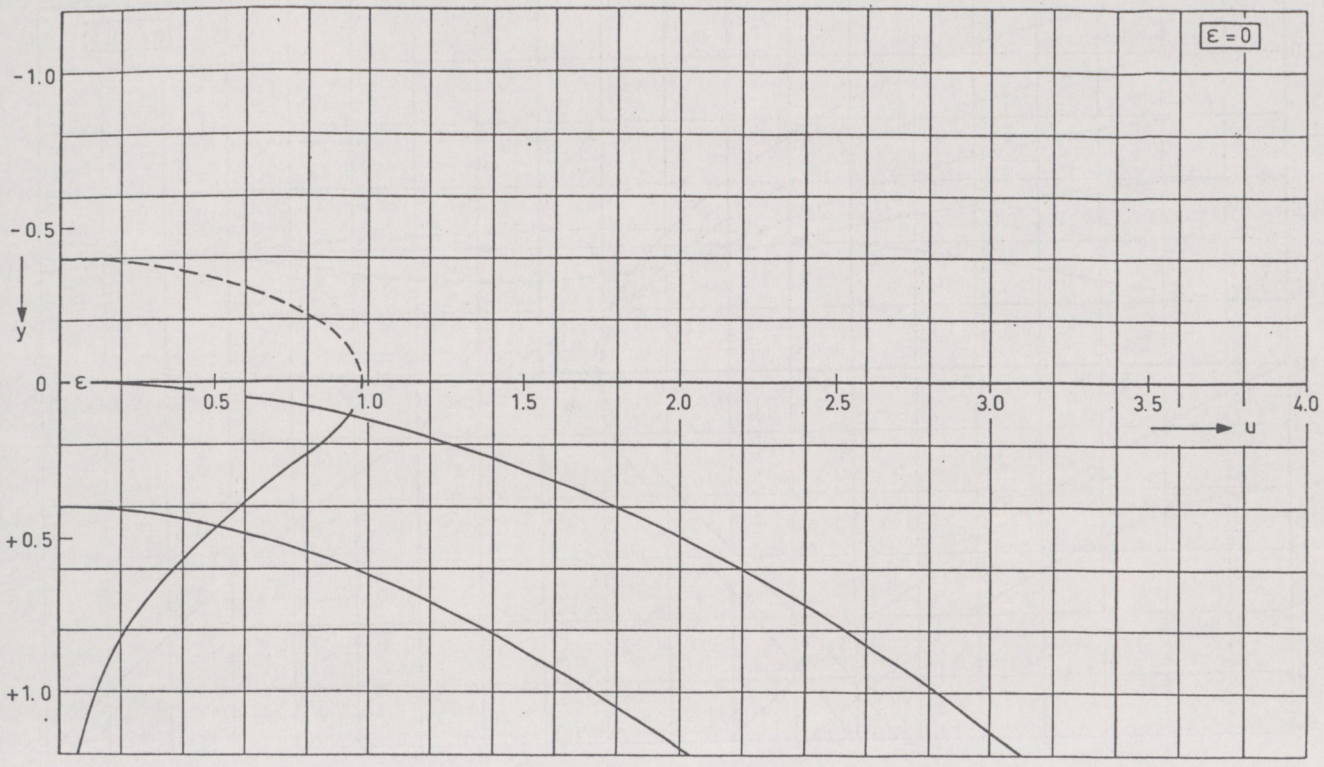


Figure 4. Curves for $\varepsilon = 0$ with $q = 0$ at $u = 0$

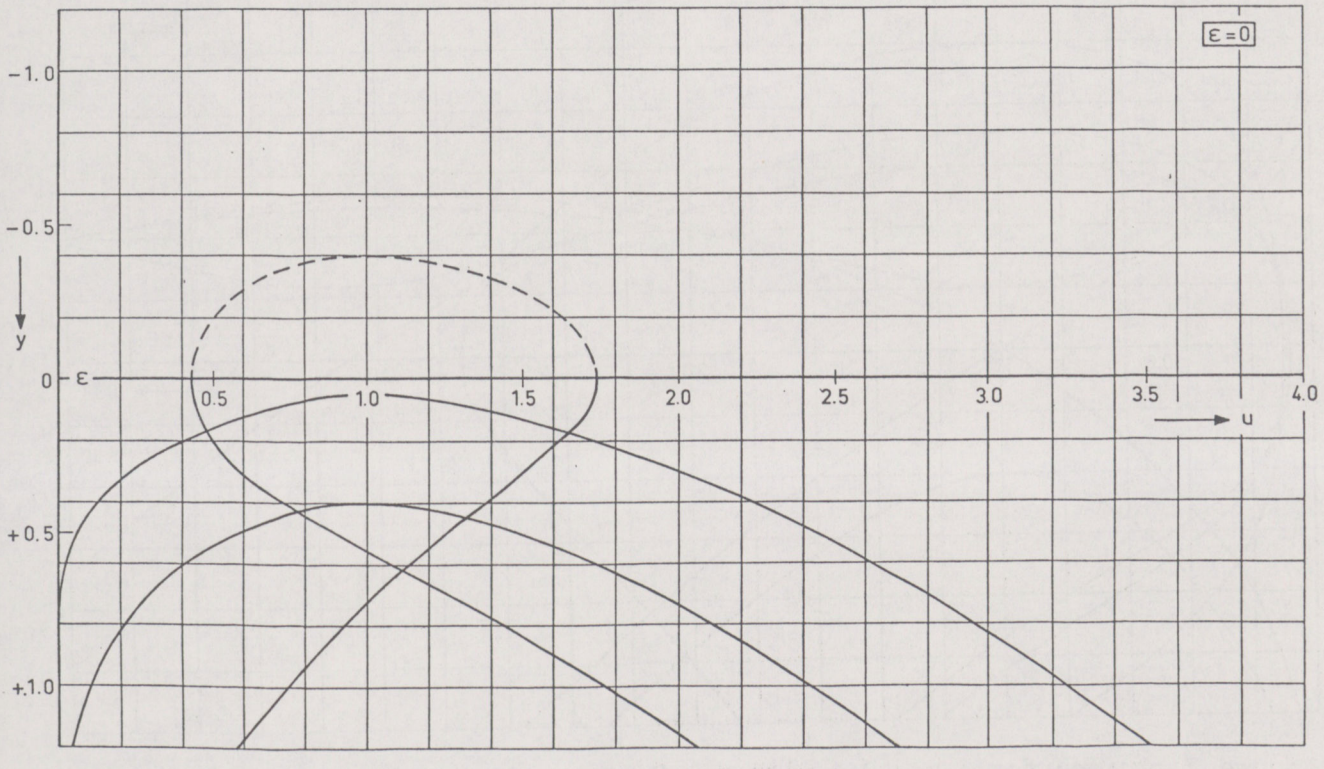


Figure 5. Curves for $\varepsilon = 0$ with $q = 0$ at $u = 1$

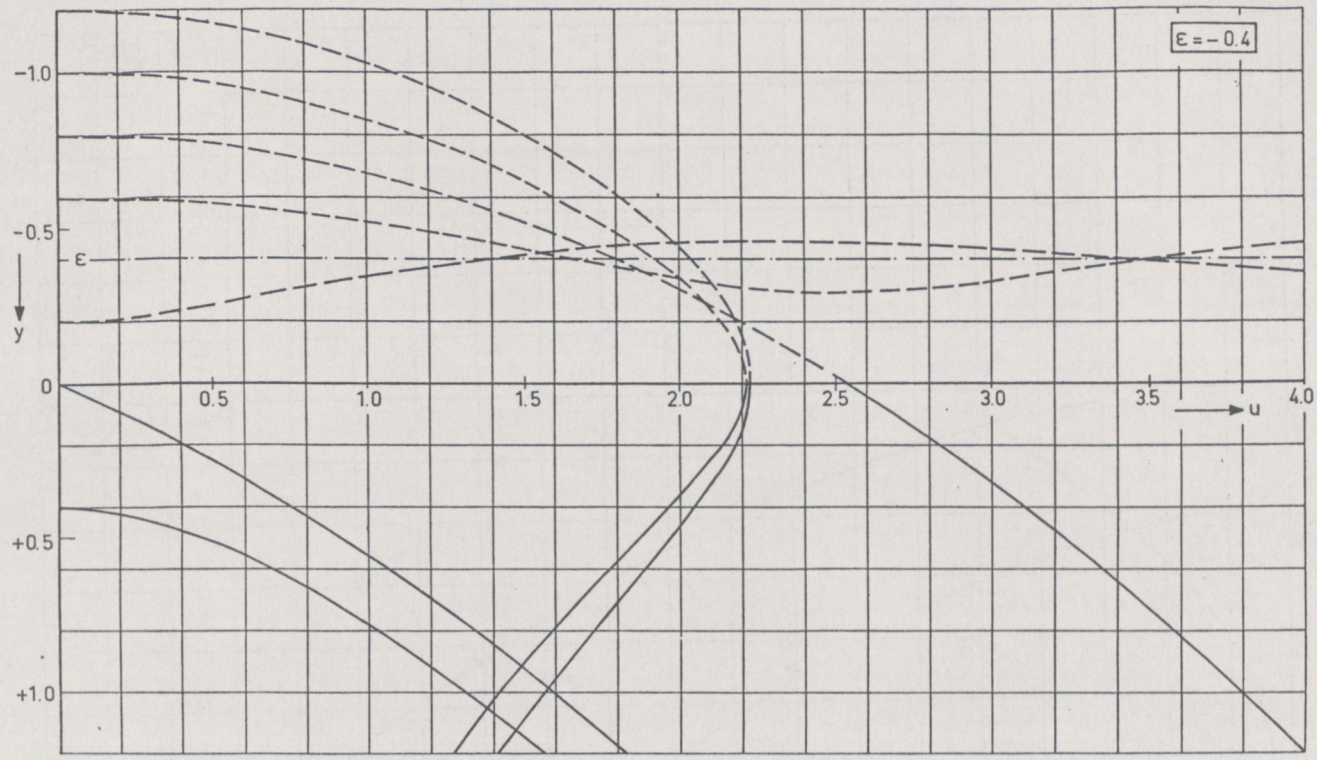


Figure 6. Curves for ε = - 0.4 with q = 0 at u = 0

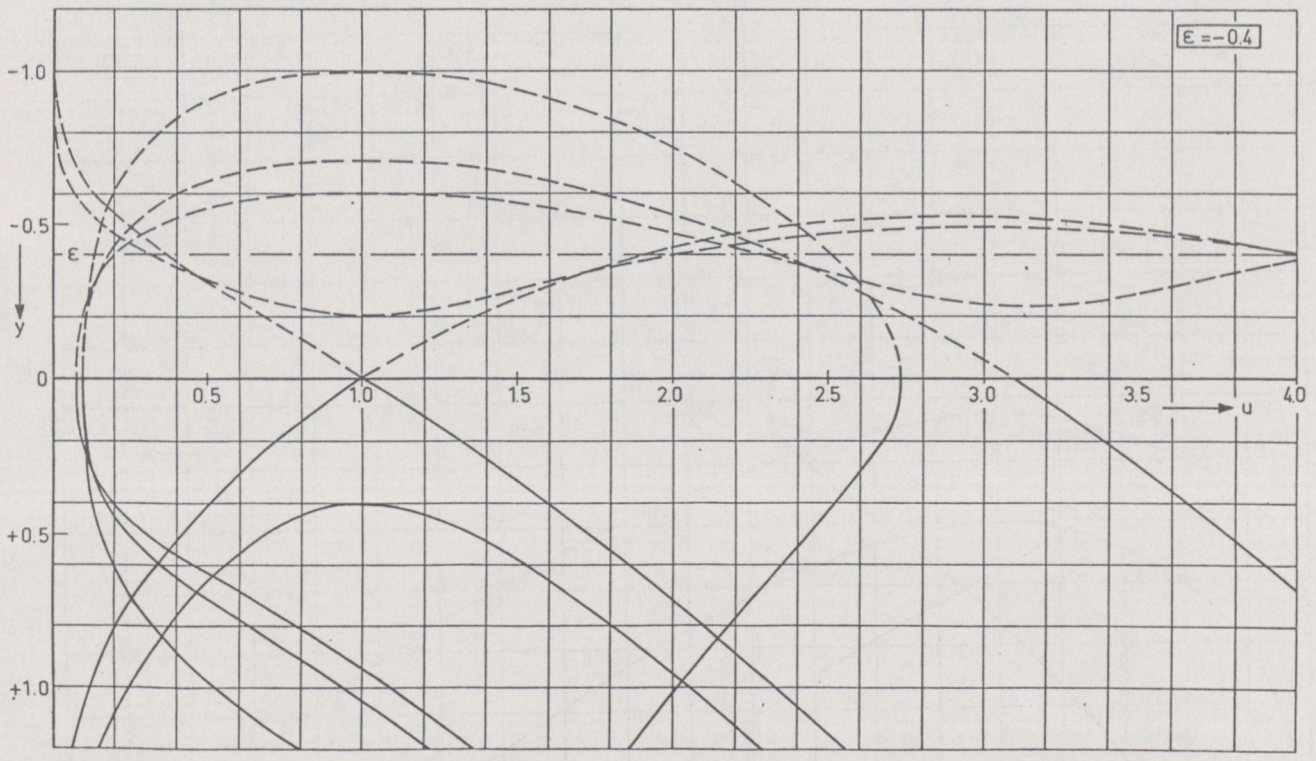
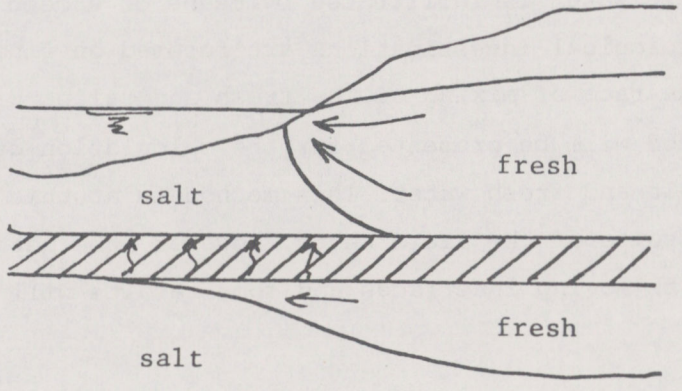


Figure 7. Curves for ε = - 0.4 with q = 0 at u = 1

THE SHAPE OF THE FRESH WATER-SALT WATER INTERFACE WHEN CROSSING AQUITARDS AND ON EMERGING IN OPEN SALT WATER

by W.H.C. ten Hoorn[1]

In the case of fresh groundwater flowing seaward in a geohydrological system consisting of aquifers and aquitards it is often observed that fresh water flowing under an aquitard and seeping upward through this aquitard arrives in salt groundwater as the interface between salt and fresh water in the upper aquifer is lying more to the landside than in the lower aquifer.
To calculate such an interface in a proper way would require to take into account three-dimensional groundwaterflow, diffusion and density currents, the latter being introduced by the diffusion of saline and fresh water and responsable for the removal of the upward seeping fresh water.
To get a first insight in the shape of the interface diffusion and secondary currents are disregarded as if fresh and saline water were separated by an impenetrable film.



Considering a cross section through an infinite strip, the remaining problem should require the calculation of two-dimensional flow under the boundary condition of equilibrium in pressure on the interface between fresh and salt groundwater.

Up till now no analytical solution for this problem is available and neither is a programme for numerical calculation. However by considering the way in which the streamlines intersect the boundaries between the porous media, the observed phenomena can be better understood and the distance over which fresh water seeps upward through the sea bottom can be calculated.

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THE NUMERICAL CALCULATION OF SALT WATER/FRESH WATER INTERFACES USING THE STREAM FUNCTION

by C. van den Akker[1]

INTRODUCTION

In the Netherlands the problem of salt water intrusion has always been very important. For that reason many investigations has been done in finding calculation techniques for salt/fresh water problems in groundwater.
However, today there is also under investigation the possible use of deep saline aquifers for storage of fresh water. This means that there has to be found design rules in order to install infiltration systems. Before that a lot of work has to be done in the field of groundwater flow with fluids of different density. To support the investigations field tests are performed in the dune area of Holland. In one of this field tests fresh water is infiltrated by means of a deep well in a saline aquifer. The geohydrological investigations are focused on finding the shape of the fresh water body, the rate of mixing of the fresh and saline water etc.
In this paper a method will be presented for the calculation of the instationary interface between salt and fresh water. This method is at this moment not yet compared with the results of the field tests. However, the method is used in doing some calculations with moving interfaces and some results will be presented here.

USE OF THE STREAMFUNCTION

Dealing with interface problems usually the groundwaterhead or pressure is taken as one of the variables in the equations. However, most of the time the position of the interface has to be calculated and this is merely a problem of moving of fronts closer related to the streamfunction then to the heads or pressure.
Of course by using heads or pressures a lot of calculation methods has been developed but sometimes it has advantages to use more direct the streamfunction. Also there are some disadvantages related to the streamfunction but for two dimensional flow it turns out to be rather helpful in solving a number of flow problems. However, before the streamfunction can be used in plane groundwater flow problems with instationary interfaces, some basic derivations has to be done. One of the problems that arises is the many-valuedness of the streamfunction around sinks and sources. According to C. van den Akker (ref.2) this problem can be solved.

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THE BASIC EQUATIONS

In plane groundwater flow Darcy's law states

$$\left.\begin{aligned} q_x &= -k_{xx}\,\frac{\partial \varphi}{\partial x} - k_{xy}\,\frac{\partial \varphi}{\partial y} \\ q_y &= -k_{xy}\,\frac{\partial \varphi}{\partial x} - k_{yy}\,\frac{\partial \varphi}{\partial y} \end{aligned}\right\} \qquad 1$$

where q_x and q_y are the components of the specific discharge in x and y direction of a cartesian coordinate system.

φ is the groundwatershead and k is the permeability. With the continuity equation

$$\frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} = 0 \qquad 2$$

this results in

$$\frac{\partial}{\partial x}\left(-k_{xx}\,\frac{\partial \varphi}{\partial x} - k_{xy}\,\frac{\partial \varphi}{\partial y}\right) + \frac{\partial}{\partial y}\left(-k_{xy}\,\frac{\partial \varphi}{\partial x} - k_{yy}\,\frac{\partial \varphi}{\partial y}\right) = 0 \qquad 3$$

This equation (3) can be solved numerically with help of e.g. the finite element method.

On the basis of the continuity equation can be stated that there exists a stream function ψ such that

$$\left.\begin{aligned} q_x &= -\frac{\partial \psi}{\partial y} \\ q_y &= +\frac{\partial \psi}{\partial x} \end{aligned}\right\} \qquad 4$$

From equation 1 and 4 it follows that

$$\frac{\partial \varphi}{\partial y} = -\frac{k_{xy}}{k_{xx}k_{yy} - k_{xy}^2}\,\frac{\partial \psi}{\partial y} - \frac{k_{xx}}{k_{xx}k_{yy} - k_{xy}^2}\,\frac{\partial \psi}{\partial x} \qquad 5$$

In a similar way it can be derived that

$$\frac{\partial \varphi}{\partial x} = \frac{k_{xy}}{k_{xx}k_{yy} - k_{xy}^2}\,\frac{\partial \psi}{\partial x} + \frac{k_{xx}}{k_{xx}k_{yy} - k_{xy}^2}\,\frac{\partial \psi}{\partial y} \qquad 6$$

If the groundwater head φ is single valued, one can state that

$$\frac{\partial}{\partial y}\left(-\frac{\partial \varphi}{\partial x}\right) + \frac{\partial}{\partial x}\left(\frac{\partial \varphi}{\partial y}\right) = 0 \qquad 7$$

From equation 5, 6 and 7 it follows that

$$\frac{\partial}{\partial y}\left(-\frac{k_{xy}}{k'}\cdot\frac{\partial\psi}{\partial x}\right)+\frac{\partial}{\partial y}\left(-\frac{k_{yy}}{k'}\frac{\partial\psi}{\partial y}\right)+\frac{\partial}{\partial x}\left(-\frac{k_{xy}}{k'}\cdot\frac{\partial\psi}{\partial y}\right)+\frac{\partial}{\partial x}\left(-\frac{k_{xx}}{k'}\cdot\frac{\partial\psi}{\partial x}\right)=0 \qquad 8$$

where $K' = k_{xx}k_{yy} - k^2_{xy}$

If the permeabilities in the principal directions of anisotropy are k_1 and k_2, equation 8 can be written as

$$\frac{\partial}{\partial y}\left(-\frac{k_{xy}}{k_1 k_2}\cdot\frac{\partial\psi}{\partial x}\right)+\frac{\partial}{\partial y}\left(-\frac{k_{yy}}{k_1 k_2}\cdot\frac{\partial\psi}{\partial y}\right)+$$

$$+\frac{\partial}{\partial x}\left(-\frac{k_{xy}}{k_1 k_2}\cdot\frac{\partial\psi}{\partial y}\right)+\frac{\partial}{\partial x}\left(-\frac{k_{xy}}{k_1 k_2}\cdot\frac{\partial\psi}{\partial x}\right)=0 \qquad 9$$

If the principle directions of the anisotropy coincide with the coordinates x and y of the cartesian coordinate system the equation 9 reduces to

$$\frac{\partial}{\partial x}\left(-\frac{1}{k_2}\frac{\partial\psi}{\partial x}\right)+\frac{\partial}{\partial y}\left(-\frac{1}{k_1}\frac{\partial\psi}{\partial y}\right)=0 \qquad 10$$

In this form the equation was already given by Bear (1972)

CALCULATION OF INSTATIONARY INTERFACES BETWEEN SALT/FRESH WATER

The basic equations

Using equation (10) derived in the former chapters the streamlines in two dimensional groundwaterflow can be calculated. A numerical method which can be used very well is the finite element technique. However, if there are inmiscible fluids with different densities the basic equations can be derived as indicated below:

For plane groundwater flow in an inhomogeneous anisotropic porous medium the law of Darcy can be written as

$$\left.\begin{aligned} q_x &= -\frac{K_1}{\mu}\left(\frac{\partial p}{\partial x}\right) \\ q_y &= -\frac{k_2}{\mu}\left(\frac{\partial p}{\partial y}+\rho g\right) \end{aligned}\right\} \qquad 11$$

where p = pressure , K = intrinsic permeability

μ = dynamic viscosity , ρ = density

g = acceleration of gravity

The principle of the conservation of mass yields:

$$\frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} = 0 \qquad 12$$

Combining equations 11 and 12 results in

$$\frac{\partial}{\partial x}\left(-\frac{K_1}{\mu}\frac{\partial p}{\partial x}\right) + \frac{\partial}{\partial y}\left(-\frac{K_2}{\mu}\left(\frac{\partial p}{\partial y} + \rho g\right)\right) = 0 \qquad 13$$

However, Darcy's law can also be written as:

$$\left.\begin{aligned} q_x &= -\frac{\partial \psi}{\partial y} \\ q_y &= +\frac{\partial \psi}{\partial x} \end{aligned}\right\} \qquad 14$$

The equations 11, 13 and 14 result in:

$$\frac{\partial}{\partial x}\left(\frac{\mu}{K_2}\cdot\frac{\partial \psi}{\partial x}\right) + \frac{\partial}{\partial y}\left(\frac{\mu}{K_1}\cdot\frac{\partial \psi}{\partial y}\right) = -\frac{\partial}{\partial x}(\rho g) \qquad 15$$

If $\gamma = \rho g$ than equation 15 can be written as:

$$\frac{\partial}{\partial x}\left(\frac{\mu}{K_2}\cdot\frac{\partial \psi}{\partial x}\right) + \frac{\partial}{\partial y}\left(\frac{\mu}{K_1}\cdot\frac{\partial \psi}{\partial y}\right) = -\frac{\partial \gamma}{\partial x} \qquad 16$$

According to the theory developed by de Josselin de Jong (1969) the part $\frac{\partial \gamma}{\partial x}$ is a term that can be explained as follows. In a liquid with a constant density the term $\frac{\partial \gamma}{\partial x}$ is equal to zero. Only at places with no uniform density e.g. at an interface between two inmiscible liquids this vortex is present.

The adjoint groundwater flow problem.

It is well-known that every plane groundwater flow problem has an adjoint problem in such a way that the equipotential lines of one of the problems are the streamlines of the other problem. Also this means that a sink or source in one of the problems is equal to a vortex in the other problem.
This means that the vortices on the interface between two liquids with different densities in the streamfunction represention can be described as sinks or source terms.

The strength of this sinks or source terms can be derived:
Assume there is a part of an interface with a length Δl and assume that the thickness of this interfaces, where the change of the density ρ occurs, is equal to Δn (see figure 1).

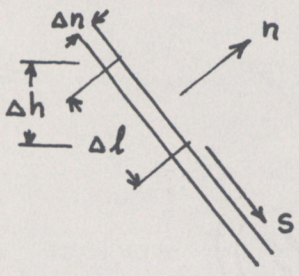
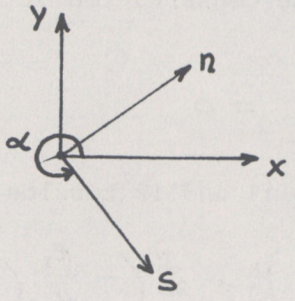


fig. 1



If now x is the angle between the positive x-axis and the direction of the interface (counter clock wise) then:

$$\sin\alpha = -\frac{\Delta h}{\Delta l} \qquad 17$$

If instead of the coordinates perpendicular to the interface (n-direction) and along the interface (s-direction) the cartesian coordinate system x and y is taken this results in:

$$\left.\begin{aligned} s &= x\cos\alpha + y\sin\alpha \\ n &= -x\sin\alpha + y\cos\alpha \end{aligned}\right\} \qquad 18$$

From equation 18 it follows

$$\frac{\partial n}{\partial x} = -\sin\alpha \qquad 19$$

The change of the γ into the s-direction along the interface is equal to zero so

$$\frac{\partial \gamma}{\partial s} = 0 \qquad 20$$

Perpendicular to the interface the change of γ over the thickness of the interface can be expressed as:

$$\frac{\partial \gamma}{\partial n} = \frac{\rho_1 - \rho_2}{\Delta n}\, g \qquad 21$$

where ρ_1 and ρ_2 are the densities of the two inmiscible fluids.

Also there can be written that

$$\frac{\partial \gamma}{\partial x} = \frac{\partial \gamma}{\partial n}\cdot\frac{\partial n}{\partial x} + \frac{\partial \gamma}{\partial s}\cdot\frac{\partial s}{\partial x} \qquad 22$$

By substitution of equation 20 and 21 into equation 22 one obtains:

$$\frac{\partial \gamma}{\partial x} = -\frac{(\rho_1 - \rho_2)}{\Delta n}\cdot g.\sin\alpha \qquad 23$$

In the application of the finite element using the variational calculus this term $\frac{\partial \gamma}{\partial x}$ appears as:

$$Q = \iint_A \frac{\partial \gamma}{\partial x} \cdot dx\,dy = \iint_A - \frac{(\rho_1 - \rho_2)}{\Delta n} g \cdot \sin\alpha \cdot dx\,dy \qquad 24$$

or

$$Q = (\rho_1 - \rho_2)\, g \cdot \Delta h \qquad 25$$

This means that the sink or source Q in the application of the streamfunction can be represented as stated in equation 25. This term is considered to act in the centre of gravity of the surface $\Delta l * \Delta n$.

So now there can be stated that within the application of the streamfunction and dealing with sharp interfaces this interfaces can be represented by a number of sink or source terms.

Example

For reasons of simplicity a trivial example of a moving interface will be treated. Suppose there is a box filled with homogeneous isotropic sand. The dimensions of the box are 4 m. height and 4 meter width (see figure 2).
At time t=0 a vertical interface between 2 liquids with different densities is present at x=2. By application of the vortex theory this interface is now represented as a number of sinks in the streamfunction. The boundary condition of the problem is very simple. All along the boundary the streamfunction value is $\psi = 0$
During a time step Δt the interfaces is moved towards an other position. Again this interface is represented by sinks and again the streamfunction is calculated all over the flow domain. The interface moves along the streamlines during the time step Δt and again the new position is obtained. In the final situation the interface has become horizontal and there is no flow at all left in the system (see fig. 3).

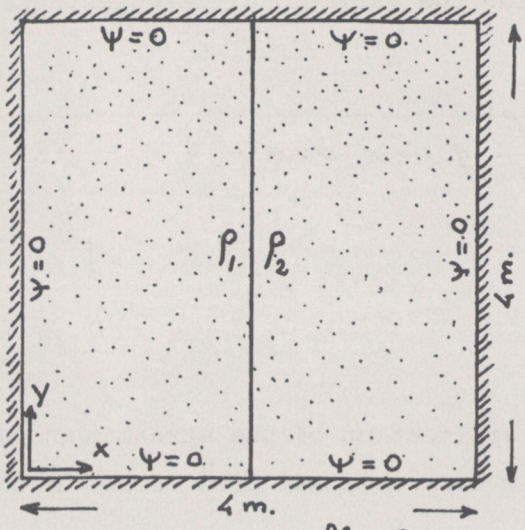


fig. 2

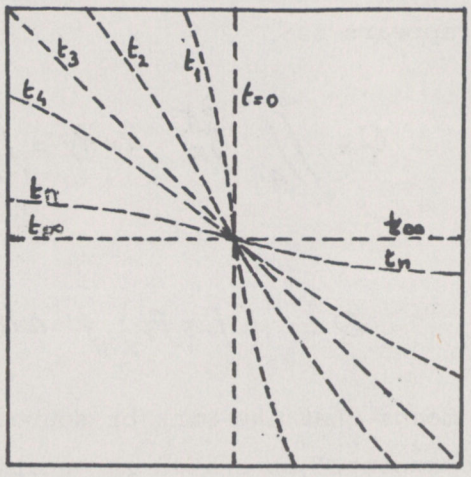


fig. 3

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ARTIFICIAL REMOVAL of INTRUDED SALINE WATER in a DEEP AQUIFER

by R.A. Schuurmans ')
and C. van den Akker '')

ABSTRACT

For the watersupply of Amsterdam groundwater is extracted from the dune area near the North Sea, since 1903 also from a deep aquifer. A heavy salt water intrusion occurred.

After 1957 the situation stabilized because the infiltration of riverwater at the surface of the area strongly diminished the deep extraction. The remaining fresh water pocket is considered as a storage, that can be used to improve or to replace the riverwater during pollution periods.

Because of the increasing production of drinkingwater and the worsening of the situation around the river Rhine (quality and accidents) the storage in the dunewater catchment area should be enlarged. As no works are allowed at the surface, the deep aquifers with salt water are investigated in order to create a storage of fresh water by means of deep well-infiltration.

INTRODUCTION

Amsterdam Waterworks is in charge with the drinkingwater supply for more than 1.000.000 people. The production amounts now about 90.10^6 m3/a. The company has three sources:

a) East of the town:
   1. Lake Waterworks: Capacity 30.10^6 m3/a.
   2. Hilversum Pumping-Station: " 2.10^6 m3/a.
b) West of the town:
   3. Dune Waterworks: " 83.10^6 m3/a.

The total installed capacity (115.10^6 m3/a.) seems to be rather good in comparison with the demand. Even in case one adds a certain percentage to the demand for safetyreasons (in Holland 10 - 20%) there is still about 10.10^6 m3/a extra capacity available.

But there are some reasons that make the situation worse than it seems to be. The dune waterworks get the raw material for the production of drinkingwater from:

1. Effective rainfall, average 13.10^6 m3/a.
2. Riverwater from the Rhine.

The riverwater is infiltrated in the dune area by shallow canals and collected by drains and by deep canals around the infiltration area. The system was started in 1957 and is still very effective. One of the properties of the system is that an increasing appeal to the capacity of the infiltration works asks for lower levels in the deep canals around the area. This measure affects the storage in the first aquifer with phreatic water and the proportion of storage and real production decreased during the years 1957 - 1980. In fact the time of full production, when the intake of riverwater has to be stopped, was reduced from about 60 days to 30 days.

') Amsterdam Waterworks, The Netherlands
'') National Institute for Watersupply, The Netherlands

In the mean time the situation of the river Rhine grew worse. In general the quality became bad and the number of shipping-disasters, in which dangerous and poisonous load is involved, increased. Therefore the chance that the intake of riverwater should be stopped, increased as well. However there is little insight how long such a calamity could continue, the period of 30 days is considered very well possible.

So the storage in the dunewater catchment area now limits the capacity to 70.10^6 m3/a, while the production plant has the capacity of 83.10^6 m3/a.

Measures have to be taken to change this situation. The problems of the river Rhine could not be soluted until now and another source is not available. That makes it necessary to enlarge the storage in the dune area. The freatic first aquifer is not available for this anymore. Nowadays the area has the big interest of nature conservation and it is impossible to make new infiltration and storage works. So the attention is turned to the deep aquifers.

GEO-HYDROLOGICAL SITUATION

The deep aquifers are situated between 20 and 160 m below O.D. (=Mean Sea Level) and are separated from the first aquifer by a clay layer, which has an average hydraulic resistance of 12 years or 4400 days. At the depth of 160 m layers of fine sand with silt and clay layers start.

The deep aquifer itself can be divided in three important layers: two aquifers separated by a semi-pervious layer from about 60 until 90 m below O.D. This layer has a resistance varying from 500 days in the north-east to 5 days in the south of the area.

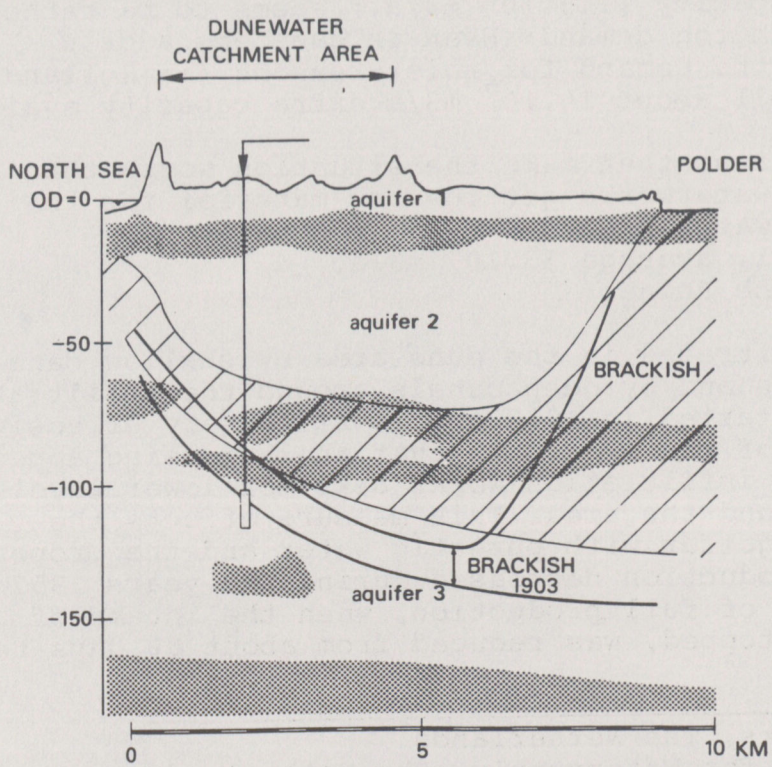

fig. 1
The system of the aquifers, the salt water intrusion and the situation of the infiltration well

The second aquifer extends from 20 to 60 m below O.D. It consists of rather coarse sand with a total transmissivity of 1000 m^2/d. The third aquifer, from 90 until 160 m, has layers of very coarse sand. The transmissivity is about 3000 m^2/d. Nearly the total aquifer always contained salt water, the fresh water pocket originally had a maximum depth of about 110 m at the eastside of the area. At the westside the depth was 70 m. The thickness of the brackish zone varied from 20 m (eastside) to 10 m (westside).

The second aquifer originally was filled with fresh water. In this layer pumping by deep wells started in 1903. The pumping increased until 1957 to an amount of 20.10^6 m3/year. The leakage from the first aquifer is about 7.10^6 m3/year. So an alarming situation arose during and after the second world war by intrusion and upconing of brackish water. This course of events was stopped by the infiltration of riverwater in 1957.

Since this year the fresh water in the second aquifer is used for quality reasons and during periods when the riverwater-intake has to be stopped. It is considered as a deep safe storage. To use this storage now a number of 290 deepwells are available with a total capacity of 3000 m3/h. Reconstruction of the well-system is going on in order to restore the original and allowed capacity of 4000 m3/h. This is about 40% of the wanted capacity of the catchment area.

The use of this capacity in time of calamities on the river Rhine is limited. In the first place from the point of view of the waterbalance. In about 60 days the total input per year of the second aquifer by leakage would be abstracted. Secondly there will be an immense salt water intrusion, specially by upconing from the third aquifer.

Therefore there has to be done something about this upconing of salt water. The solution can be to create a new fresh water pocket below the semi-pervious layer between the second and third aquifer. The artificial storage of fresh water in the salty third aquifer requires the technical science of deepwell infiltration. If the knowledge and experiences prove the validity of the operational system, maybe the extraction of the existing 4000 m3/h can be enlarged. The safety of the watersupply of Amsterdam is strongly served in that case.

DEEPWELL INFILTRATION

The technic of deepwell infiltration is in use for many years in the Netherlands. Operational systems exist for discharge of waste water, for cooling and for temporary drainage by deepwells. To keep the wells in operation for a long time special precautionary measures are required. If the wells get clogged they are cleaned in a mechanical or chemical way. After some time often the well has to be abandoned because the clogging can not be removed from the surrounding material.

Drinkingwater supply makes special demands on deepwell infiltration. The amount of water generally is rather high and the system must be in operation during a long time . The infiltration water has to be conditioned in a rather simple way and by preference without chemical measures. If there is some clogging of the wells there must be a good method for cleaning. This method also should be rather not chemical but mechanical or hydraulic. Besides the system has to be feasible from the economical point of view. In the Netherlands some research has been done upon the economical feasibility of the injectionwells.

The mentioned technical construction of the well and the conditioning of the infiltration water previously have been tested by Amsterdam Waterworks in a special well that was situated in the second aquifer to the east of the area. The available water for infiltration is the riverwater. This water is pretreated at the intakestation by coagulation, sedimentation and rapid sandfiltration. The experimental well was clogged by the riverwater because sometimes the water had not the right properties. An important property is represented by the MFI (Modified Fouling Index), a test of the water by membrane-passage. A low index, e.g. drinkingwater with number 1, does work excellent; the pretreated riverwater with sometimes 6 to 12 did clog the well. The important conclusion from the testwell was that the riverwater needs a better treatment.

During this period of tests the spreading of the infiltrated water was observed intensively. The difference in quality parameters of the riverwater and the groundwater showed the shape of the recharged waterbody with rather sharp interfaces. A fairly good agreement was obtained between a numerical calculation with a finite element technique and the observations from the test. In this case both liquids had the same density.

SALT WATER REMOVAL

After the prelimanary experiments a new deepwell for infiltration was designed. The objective in this case was to infiltrate fresh water with low density in the third aquifer, containing salt water with a high density. This well must confirm that the system also works in the new circumstances, being the deep aquifer near the sea. An extra treatment of the riverwater should improve the infiltration condition of the riverwater. Most of all, however, the proof was wanted that the fresh water would stay in a lens with rather sharp interfaces and without to much mixing with the salt water. Watersupply is not interested in brackish water.

The pretreatment of the riverwater was extended with a type of slow sandfiltration. This was realised by drains on a depth of 0,5 m below the bottom of existing infiltration canals. The value of MFI up to now was decreased below 3.

The construction of the well did not change much. The waterhead of the third aquifer was about 5 m below groundlevel. So the level of the infiltration canal was sufficient to realise a flow of about 20 m3/h. The diameter of the screen is 230 mm, the depth of the screen is from 101 m until 116 m below O.D.

In the surroundings of the injectionwell three observationwells were founded on distances of 5 m, 15 m and 40 m. In the wells many small screens were placed on different levels; geohmcables were installed for measuring the electric resistance on many points. A processor registrates several parameters. These can be transported by telephonecable to the office on a distance of 4 km. Here a second processor takes over the data for handling, printing and plotting.

The test started on January 26, 1981. The measured data are presented here until about August 14. So during about 200 days a volume of nearly 100.000 m3 riverwater has been recharged in the deep aquifer.

The objectives of the test with the injectionwell can be summarized as follows:

- to investigate the possibility of infiltrating pretreated riverwater in the deep aquifer of the dunewater catchment area.

- to investigate the salt-fresh water interaction in the deep aquifers, where problems as hydrodynamic dispersion, instationary recharge by the injection well, inhomogeneous and anisotropic soils and the extraction of the infiltrated water play an important role.

- to investigate the possibility of measuring and controlling the storage into the deep aquifer by means of observationwells with filters and geohm-cables. Here also the automatic processing of data with help of a computer is essential.

- to find out the basic design rules for a recharge/recovery system with respect to hydrology where also on a long term the system is useful.

THE RESULTS OF THE TEST

At this moment (September 1981) the injectionwell is nearly 8 months into operation. Of course this period is to short to deal with the questions and problems as stated in the previous chapter. However, with the results of the measurements and with the experience obtained, some interesting results can be presented. Furthermore it is important for the investigations to hear the opinion of specialists on salt/ fresh water problems with respect to tests as described here. One should keep in mind that tests like these take several years in order to obtain the maximum information of the groundwater movement and salt/fresh water interaction.
Therefore a discussion with colleagues and suggestions would be a great help in obtaining the maximum results from tests like these.

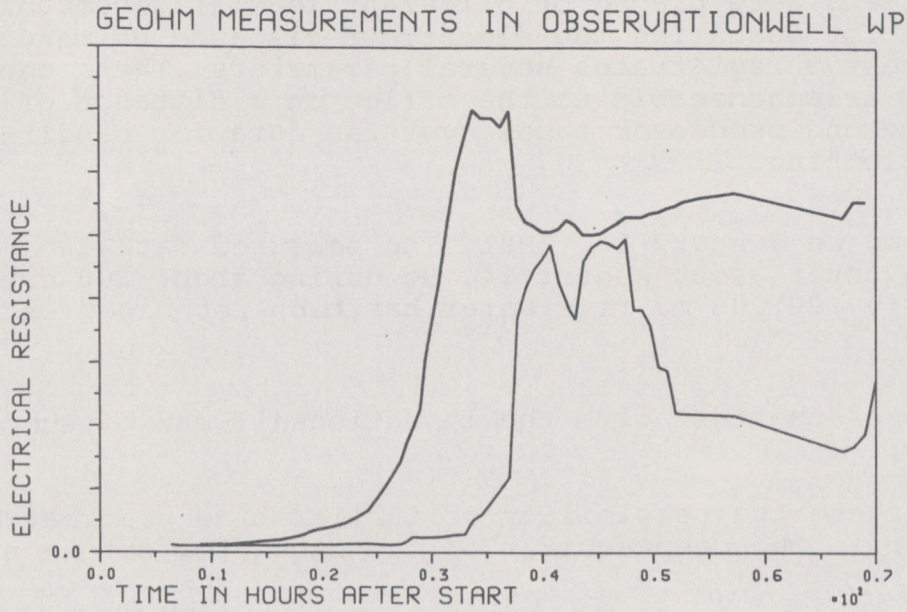

fig. 2 Examples of measuring the electrical resistance in two points of the geohmcable. In this case the cable is situated 5 meters from the infiltrationwell. In about 10 hours the front passes the point and the surrounding water turns from salt into fresh.

However the tests with the well shall continue a long time, already now can be established that the main objective of the pilot plant has been achieved, namely the proof that in the local aquifer, filled with salt water, a storage of fresh water can be created.
From the different measurements it is easy to conclude how the shape of the fresh waterbody is on a certain point of time.
Moreover the pictures of the measurements make clear that the extension of the fresh waterbody has some characteristics:

a. there is a brackish zone at the bottom of the fresh waterbody with a thickness in the order of magnitude of 5 m.
b. the quick extension of the fresh waterbody in the most coarse layer of the aquifer does not cause a very high amount of brackish water, in spite of the high velocity.
c. on the top of the fresh waterbody the present salt and brackish water is removed slowly.

THE BRACKISH ZONE

Until now it is calculated that about 50% of the infiltrated fresh water is present in the fresh waterbody, if this body has a regular circular shape. This will be controlled by new observation-wells in several directions with a distance of about 100 m from the infiltrationwell.

This rather high loose of fresh water has an important cause in the fact that the screen of the infiltrationwell is about 15 m below the semi-pervious layer. This depth is choosen because between the mentioned layer and the screen there is brackish water. Moreover the very coarse layer in the aquifer on a depth of 95 m until 125 m should be used to observate the phenomenon under condition of the high velocity. So the present well is not representative for a pilot plant, investigating a system with operational characteristics.

Within the period of 200 days the thickness of the brackish zone is relatively small in comparison with the natural situation in the aquifer. The brackish zone below the dunewater catchment area originally was 10 to 20 m, now a thickness of about 5 m is found. Very near to the injectionwell the thickness is even less.

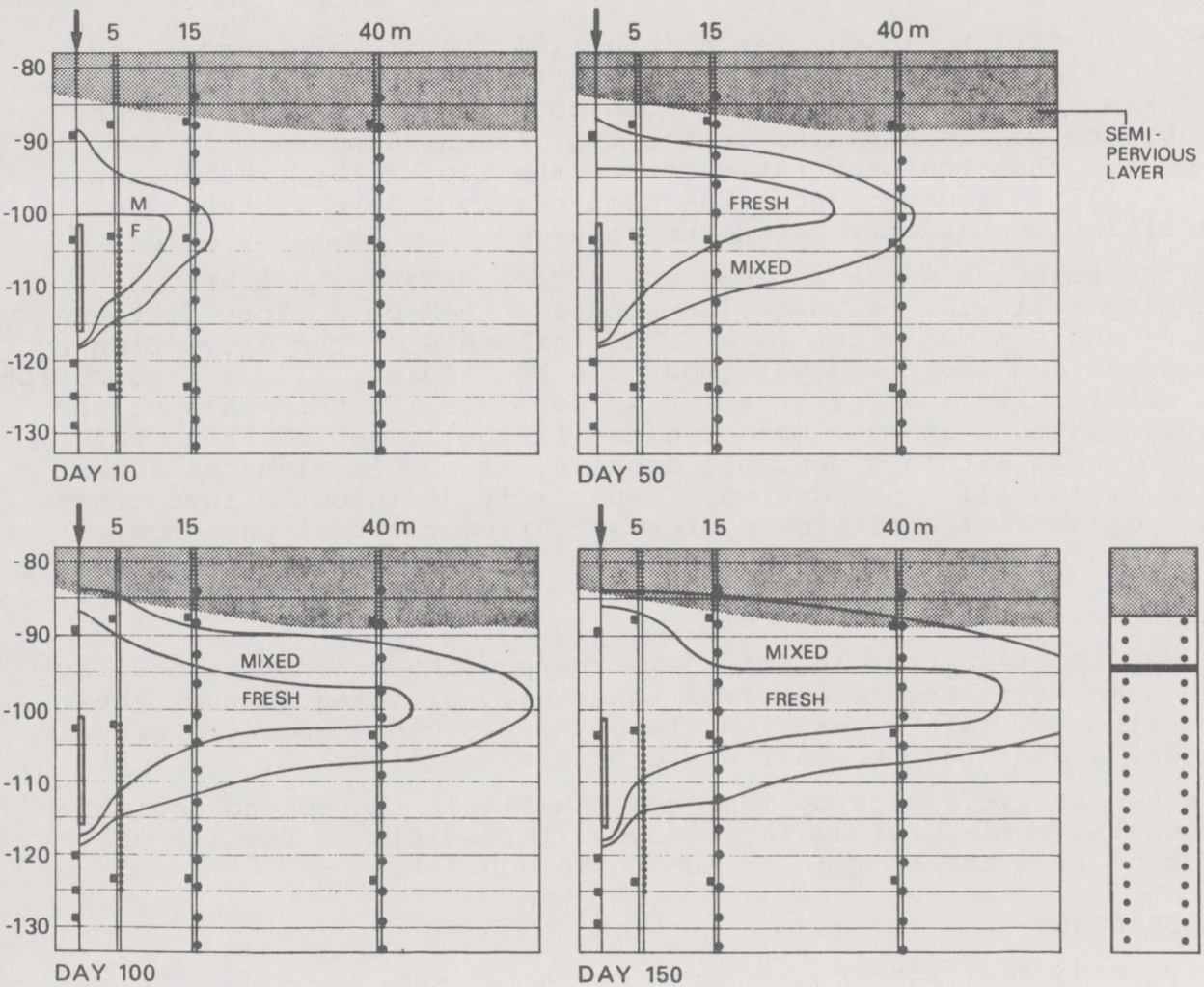

fig. 3 The spreading of the infiltrated fresh water on several points of time. Fresh, brackish and salt water zones are measured by geohmcables, water-samples from filters and by temperature-logging in the wells.

It is unknown if the several small silt or clay layers in the aquifer have a big influence on the amount of brackish water. These layers will cause dispersion. The inhomogeneous properties of the aquifer, as a result of the sedimentology, should be studied better in a fundamental way. Tests on laboratory-scale should be done to evaluate the influence of sedimentary properties on dispersion.

CONCLUSIONS

During the first 200 days of the test with the injectionwell it is made sure that

a. the available riverwater can be conditioned by a simple method of drainage, that is useful during a long time;
b. the contruction of the well is allright;
c. a storage of fresh water can be made in a total salt aquifer. It seems that dispersion stays within predictable limits.

FUTURE AIMS

The well will be in use for a long time. During at least 1 year there will be observations of the increasing body of fresh and brackish water. These observations concern the outskirts of the body, the bottom and also on the top. In the long term it is expected that the brackish water on the top of the fresh water body will be removed until the semi-pervious layer. Maybe the infiltration capacity has to be enlarged from 20 to 30 m3/h.

If the brackish water is removed, a second fase can start. A pumping well will be installed with a screen on a short distance above the semi-pervious layer. The influence of the extraction by this well on the fresh waterbody can be studied. Also a combination of abstraction and infiltration is possible. In this fase of the investigations it must become clear if the losses of fresh water can be kept within acceptable amounts. If so, it might be a next step to install a new deepwell, specially designed to investigate the possibilities of a deep storage for operational purposes, among them the capacity-increasing effect and influence on quality owing to the choice of a good distribution of traveltimes and residencetimes in the system.

It may be clear that in this case the pumping can not be done by the same well. Extraction from this well, situated so near the brackish and salt water, immediately would cause upconing of salt water. A dual-purpose well would be a misapplication.

Instead an extraction by many small wells in the second aquifer above the semi-pervious layer makes it possible to utilize the leakage from the storage in the third aquifer.

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CONCLUDING SESSIONS

by Malin Falkenmark[1]

TERMINOLOGY DISCUSSION

A discussion took place on terminological issues, especially the best terminology to be used for intruded and fossil water of marine origin.

It was concluded that connate ("born together") is the term to be used for salt water, that was trapped when the sediment was originally deposited, in other words the entrapped, original water. A number of processes during the following time period makes this water move gradually (recharge, tidal processes, land uplift, sea transgression etc). For the salt water after such processes the preferred term would be relict water. This word is preferable to "fossil", which is linguistically deceptive (Latin: "dug out").

Intruded salt water refers to salt water which has entered aquifers in the recent past, due to either natural causes or human interference. The term encroached might be used for water, intruded by human interference.

In conclusion, the differentiation of saline water in aquifers could be structured as follows:

connate	relict	intruded
. originally trapped when deposited	. after displacement through water exchange	. sea water entered in recent past due to
. still present in the original formation	. still old (>1000 yrs)	- natural causes - human interferences (also encroached)

GENERAL DISCUSSION

The general discussion led to the following conclusions:

. it was recommended to give wider distribution of results from SWIM-meetings, due to the general interest of the problems

. achievements of the SWIM-meetings should be analyzed and a summary presented at SWIM-83 (responsible Messrs. de Breuck, Pomper, de Vries and Haman)

. next meeting will be held in Italy on the invitation from Mr. Cotecchia through the Italian participants

. instrumentation issues would be a new focus of interest to next generation, esp practical guidelines for drilling, sampling, chemical analysis etc (1-2 pages for each technique)

. it was decided to keep saline water as the main aspect of SWIM, i.e. not bring in pollution problems, which might lead too far away from saline water aspect.

SYNTHESIS

Messrs. Geirnaert, de Breuck and van Dam contributed to the following synthesis of papers and discussions during SWIM-81:

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1 NFR's Committee for Hydrology, Stockholm, Sweden

1. Contributions represented a great diversity with some differences as to national foci:

- problems	Poland, Spain
- surveys	Sweden, Spain, Italy
- hypotheses	Netherlands
- calculations	"
- experiments	"
- instruments	Sweden

2. Field work and techniques used: it was considered desirable, in future field studies, to make better use of combinations of different tools available. Variables and methods could be combined on a broader scale in the practical field work: hydrochemistry, isotopes, temperature, surface hydrology, geophysical prospecting, pumping tests etc. Furthermore tools are continuously changing. This makes it essential to devote more time in the future to the interaction between tools and field work. Special attention should be paid to age dating and isotope tracing methods, as tools for studying salt water mechanisms. Taking into account that there were no reports on these techniques to SWIM-81, such reports would be welcomed for the discussions in 1983.

3. Hydrochemistry was considered a perfect tool in fresh/salt water studies, especially as boreholes are so expensive. The present approach in regional surveys of comparing old and new data has however the serious drawback of incomparable precisions. In other words, the present technique has reached the limit of its capacity. A new interesting technique now opening up is factor analysis with large potential as much more variables can be studied. The Piper-diagram is particularly useful for analyzing mechanisms, even if it is difficult to find "clean" water types (i.a. a question of sampling level precision). Mechanisms, as sea water intrusion, cation exchange and mixing of seawater with exchanged water, all occupy different zones and give different trends in the diagram. A field for expanding efforts is research to try to explain the hydrochemical processes in the soil. More specifically, salt water in fissured rocks, not replaced by infiltrated water, should be studied in the Piper-diagram, in order to clarify what processes have affected its composition.

4. Modelling constitutes a new tool in the field of saline water research. Although the present models greatly oversimplify real conditions (simple configurations, neglection of dispersion and diffusion etc), their mathematics is already quite sophisticated. The question therefore emerges: how far can sophistication develop? Where does it reach its limit? The geology will never be completely known; there will always remain data gaps to bridge. Another problem is the problem of field checks: how far should data collection go? All this brings the sensitivity of the models into focus. In future work it is finally essential that teams, representing different foci around the mathematicians, can cooperate.

5. Future SWIMs: it was considered advisable to allocate, in the future, more time to discussions on field work, especially methods used and problems encountered. Also discussions in general should be given more time to allow a larger amount of criticism, as this would contribute to the benefit of the participants. The IHP-umbrella makes sense: SWIM offers an exchange of information on problems, and a way of finding solutions in one area of hydrology, of considerable importance also when seen in a world-wide perspective. Even if it is advisable to maintain the informal character of the SWIM-meetings, it was found desirable that the reports should be given world-wide distribution.

THE SHAPE OF A FRESHWATERLENS IN THE CASE OF TWO AQUIFERS SEPARATED BY A SEMI-PERVIOUS LAYER

by W.H.C. ten Hoorn[1]

ABSTRACT

In the case of fresh water flowing seaward in several aquifers separated by semi-pervious layers, it is often observed that fresh water in a lower aquifer is covered by salt water in the upper aquifer. Starting from potential stream in the fresh domain the form of the freshwaterbody is discussed. The evacuation of the fresh water seeping upward into the salt domain can be explained by differences in pressure introducing secondary currents.

INTRODUCTION

In the Dutch coastal area as in many other coastal regions outflow from fresh water towards sea takes place in several aquifers on top of each other, separated by semi-pervious layers. In these cases fresh water in a lower aquifer may flow seaward under salt water being present in the upper aquifer, the fresh water seeping upward in a saline environment (fig. 1). Calculation of such a case may be performed by considering the flow to be horizontal in the aquifers and vertical through the semi-pervious layer (Mualem and Bear). In this way however, apart from neglecting the vertical flowcomponent in the aquifers, no true account is given for the fresh water seeping upward towards a saline environment. In the following it is tried to get more insight in the total process of fresh water outflow for the considered case.

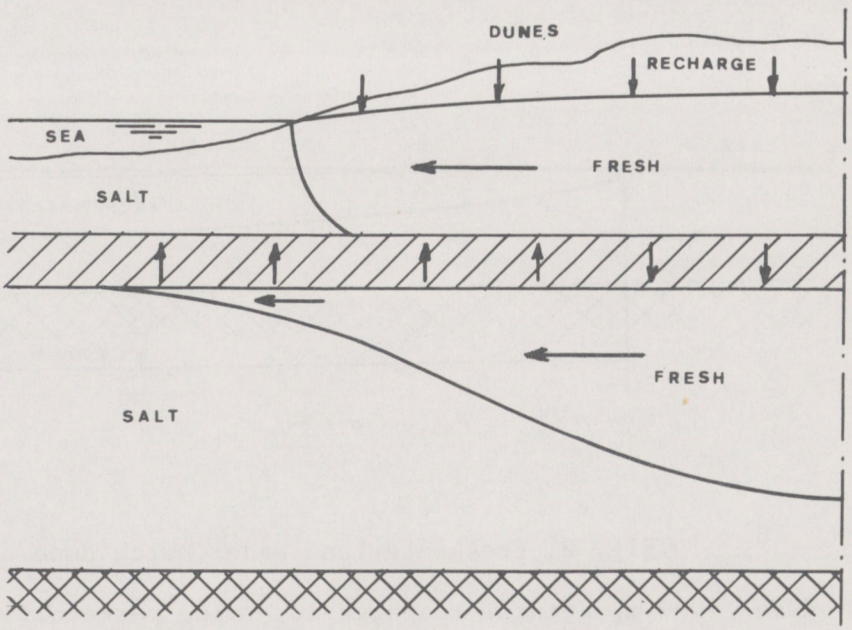

Fig. 1. Outflow of fresh water towards sea in two aquifers

MECHANISMS INVOLVED

In the following the case of two-dimensional steady flow is considerd, the relevant soil layers, aquifers as well as semi-pervious layers being homogeneous and isotropic. The phenomenon is studied for a freshwaterlens as is present below the dunes along the Dutch coast, for reasons of simplicity this freshwaterlens is considered to be symmetrical and only the seaward side of it is considered. It will be clear that in the absence of a semi-pervious layer and assuming a sharp interface there will be potential stream in the freshwaterlens, the pressure of the salt water on the interface giving a boundary condition, viz., the pressure of salt water on a point of the interface has to balance the pressure following from the freshwaterhead at that point. If we consider however fig. 1 it is obvious that there will be no potential stream

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1 Delft University of Technology

in the area where fresh water flowing through the semi-pervious layer reaches a saline environment. If we could hoewever prevent the fresh water from mixing with the saline water, e.g. by introducing an impermeable fleeze between fresh and saline water then the fresh water would be forced to flow according to the Laplace equation, i.e. as a potential stream, a boundary condition being again equal pressure on both sides of the fleeze.

It can be expected that if we would calculate the fresh waterflow in the above mentioned way (i.e. with fleeze) we would more or less accurate find the form of the interface, only the part just above the semi-pervious layer being non-realistic. In reality there is no fleeze and the fresh water can mix with the saline water. Moreover, owing to differences in density secondary currents will be introduced. So, in order to describe the phenomenon properly we have to take into account: potential flow, mixing processes and secondary streams.

POTENTIAL STREAM

In the case of one homogeneous isotropic aquifer there is no objection against calculation according to the Dupuit approximation i.e. with neglect of the vertical flow component. This will be clear when considering fig. 2, where the parameters have values that are normal under Dutch conditions e.i. width of the

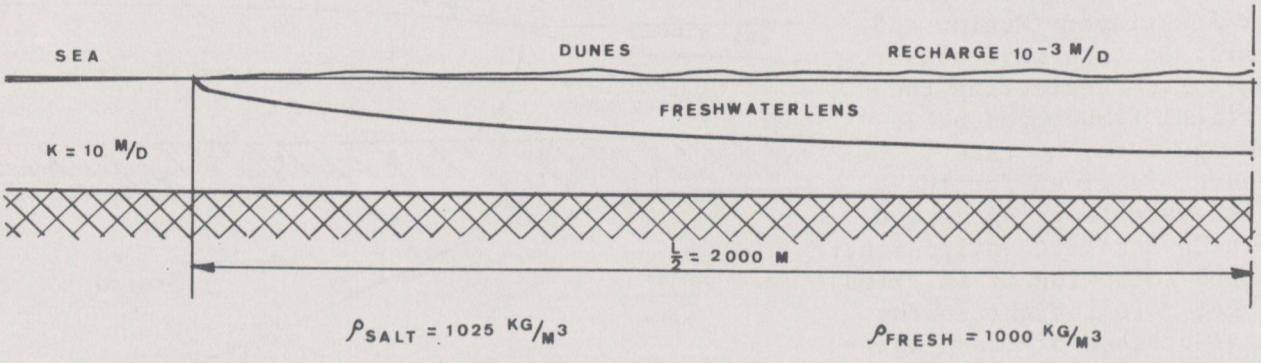

Fig. 2. Freshwaterlens under Dutch dune area

dunes 4000 m, natural recharge 10^{-3} m/d, permeability 10 m/d, density of fresh water 1000 kg/m^3 and density of salt water 1025 kg/m^3, the aquifer is considered to be homogeneous and not subdidvided by semi-pervious layers. Under these conditions the rise of the water table is about 3 m above sealevel and the freshwaterlens goes down to about 125 m below sealevel. It is obvious that it will be difficult to draw a flownet in fig. 2. In order to be able to look at the consequences following from a flownet another freshwaterlens is drawn in fig. 3. For the same width (4000 m) the same densities (1000 and 1025 kg/m^3) and the same recharge (1 mm/d) as in fig. 2, the consequences are seen from a much smaller permeability (0,1 m/d instead of 10 m/d). From the flownet it can be seen that the Badon Ghijben-Herzberg (B.G.-H.) relation does not hold between A and A', respectively B and B' but between A and A", respectively B and B".

There will be a greater difference between the depth of the interface in reality and that given by the B.G.-H. relation when a semi-pervious layer is present. In that case the flow through the semi-pervious layer will be practically vertical, involving nearly horizontal equipotentiallines in this layer. As a result the depth of the interface in the middle of the lens will be less than given by the B.G.-H. relation and on the seaside it will be more.

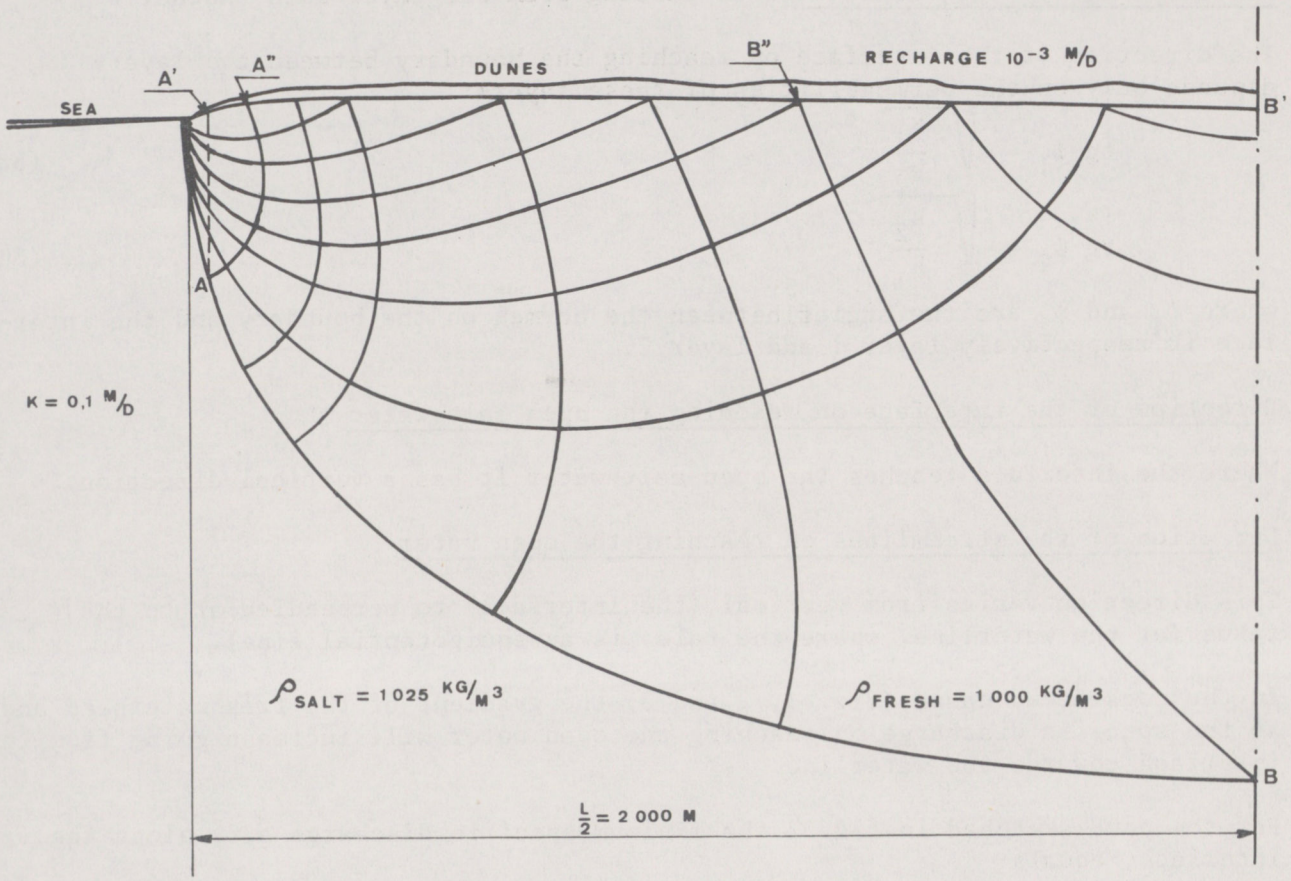

Fig. 3. Flownet in a freshwaterlens

We can try to get more insight in the form of the interface by considering some particular features of streamlines and equipotential lines (for the derivation of the equations see the Appendix).

Slope of the interface

$$\sin \phi = \frac{v}{\alpha k} , \qquad (1)$$

where ϕ - angle between the direction of the interface and the horizontal
v - specific discharge of fresh water along the interface
α - relative density difference, e.i.

$$\alpha = \frac{\rho_s - \rho_f}{\rho_f} ,$$ with ρ_s - density of salt water and ρ_f - density of fresh water

k - permeability

Interfraction of steamlines when passing the boundary between layers with different permeabilities.

Here the general law of refraction holds:

$$\frac{\operatorname{tg} \psi_1}{\operatorname{tg} \psi_2} = \frac{k_1}{k_2} \qquad (2)$$

where ψ_1 - angle between the normal on the boundary and the streamline in layer 1
ψ_2 - angle between the normal on the boundary and the streamline in layer 2
k_1 - permeability of layer 1
k_2 - permeability of layer 2

Interfraction of the interface when passing from one layer into another.

The direction of the interface on reaching the boundary between two layers depends only on the permeabilities of these layers:

$$\text{tg } \psi_1 = \sqrt{\frac{k_1}{k_2}} \qquad (3a)$$

$$\text{tg } \psi_2 = \sqrt{\frac{k_2}{k_1}} \qquad (3b)$$

where ψ_1 and ψ_2 are the angles between the normal on the boundary and the interface in respectively layer 1 and layer 2.

Direction of the interface on reaching the open salt water

Where the interface reaches the open salt water it has a vertical direction.

Direction of the streamlines on reaching the open water

This direction varies from vertical (the interface) to perpendicular to the talus (at the waterline, where the talus is an equipotential line).

In the considered cases (fig. 1, 2 and 3) the gradient of the freshwaterhead and so the specific discharge on reaching the open water will increase going from the interface towards the waterline.

For the case sketched in fig. 2 the minimum specific discharge e.i. along the interface, equals

$$\alpha k = 0{,}025 \times 10 = 0{,}25 \text{ m/d}$$

The recharge flowing towards sea amounts to

$$\tfrac{1}{2} \times 4000 \times 10^{-3} = 2 \text{ m}^3/\text{md}$$

So the length over which outflow will take place is less than 8 m.

In reality owing to the tide, resulting in a shore line shifting twice a day over a distance of perhaps a hundred meters, a considerable mixing zone will be formed.

PASSING THE SEMI-PERVIOUS LAYER

Let us consider (fig. 4, left) a freshwaterlens in a homogeneous isotropic aquifer above a semi-pervious layer, under which there is again an aquifer with the same permeability as the upper aquifer. The undermost part of the lens is sketched again in detail A (fig. 4, right), where line a represents the position as given in the left hand figure. It is clear that under other conditions, e.g. a larger recharge the lens will be larger and will eventually touch the semi-pervious layer (line b). (Mind, that every line is considered to represent a new steady-state). For a still larger recharge the interface will be forced into the semi-pervious layer (line c). Now the direction under which the interface intersects the boundary between the different layers corresponds with the angles given by (3a) and (3b). Also for a larger lens these angles apply and essentially the same form will be found when the interface reaches the bottom of the semi-pervious layer (line d).

A lens which is penetrated into the lower aquifer will again have essentially horizontal stream lines in this lower aquifer and the passing of the interface from the lower aquifer into the semi-pervious layer will also take place under

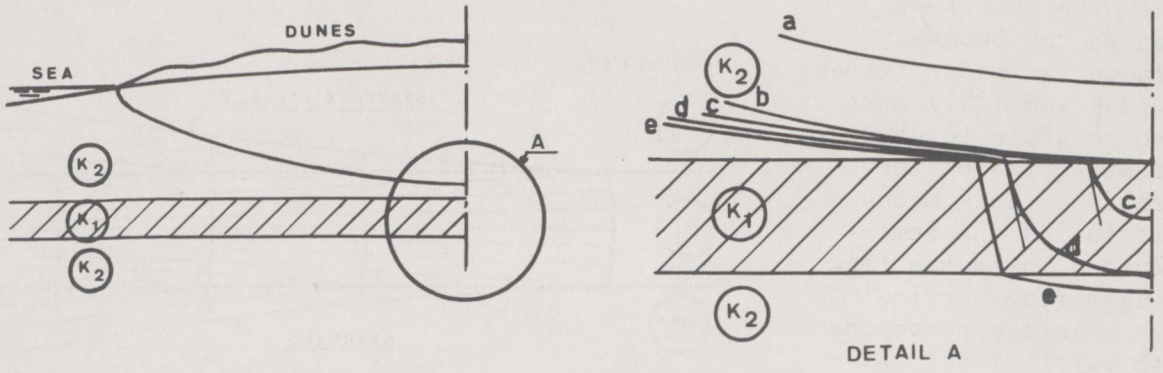

Fig. 4. Crossing of a semi-pervious layer by the fresh-salt interface

angles as given by (3a) and (3b)(line e). Now the question arises whether a further increase in recharge will result in essentially the same form of the freshwaterlens.

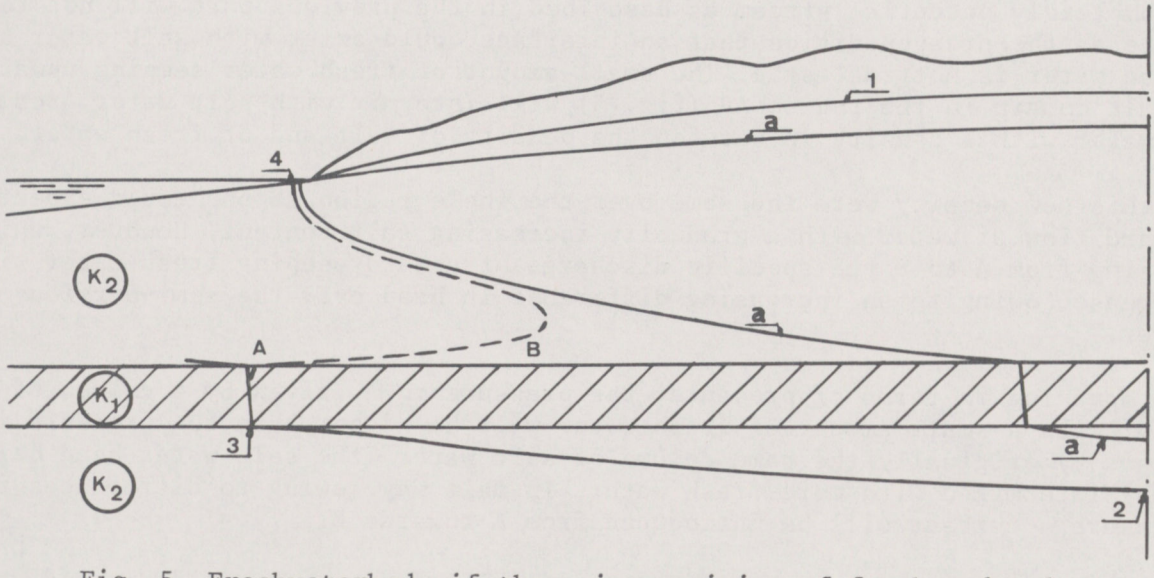

Fig. 5. Freshwaterbody if there is no mixing of fresh and salt water

In fig. 5 by line a the form of the freshwaterlens is given for a small penetration into the lower aquifer. The results of a considerable increase in recharge are also given in fig. 5:

1. a higher piezometric level,
2. a greater depth of the interface in the center of the lens,
3. a considerable shift to the left of the intersection between the interface and the bottom of the semi-pervious layer, owing to the low specific discharge in this part of the lens and the consequently slow rise of the interface,
4. a small shift towards the left of the intersection between interface and open water.

It will be clear that a continuation of the interface from point 3 in figure 5 upward and leftward makes it impossible ever to reach the point of outflow of the interface given by pont 4.

However, the angles under which the interface crosses the boundary between two media (given by (3a) and (3b)) make it also possible to change the direction of the interface. This means that inside the semi-pervious layer the flow changes from vertical with a leftward component into vertical with a rightward component (fig. 6).

If we introduce this possibility in fig. 5 we find an 'interface' as given by the dotted line.

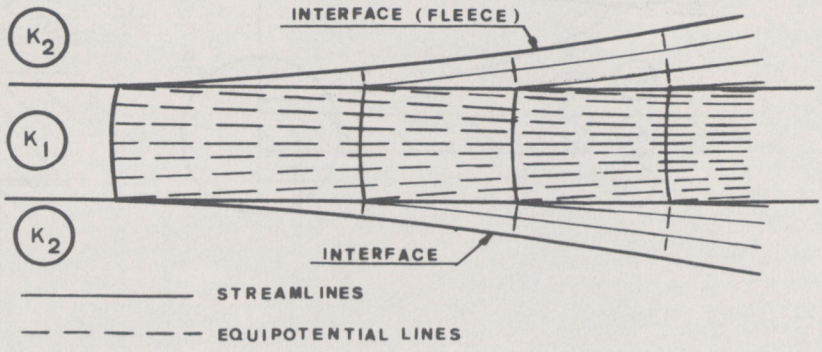

Fig. 6. Part of flownet in the region where upseeping fresh water enters a salt domain

SECONDARY STREAMS

A completely potential stream as described in the previous part will not take place as the presupposition that an interface could exist with salt water above fresh water is not realistic. The small amount of fresh water seeping upward in a salt domain in the region AB (fig. 5) will intermix with salt water, resulting in water with a density in between the density of salt and of fresh water.

If this new density were the same over the whole region AB one could expect an upward flow of water with a gradually increasing salt content. However, while passing from A to B the specific discharge of upward seeping fresh water will increase, owing to an increasing difference in head over the semi-pervious layer.

So, speaking in terms of pressure, the pressure in A, given by a column of salt water with a small amount of fresh water will be higher than the pressure in B, caused by originally the same column of salt water (the salt water head being the same) intermixed with more fresh water. In this way. owing to differences in pressure a current will be introduced from A towards B.

Flowing from the left to the right the transition zone between fresh and salt water will grow, owing to dispersion and the tendency of fresh water to flow upward. Still, as the specific discharge of fresh water flowing upward through the semi-pervious layer increases, there will be formed a layer of fresh water at the bottom of the aquifer, growing while passing towards the right (fig. 7).

This will continue until the specific discharge of fresh water in the upper layer is so large that it equals about αk as at that moment, according to (1) the fresh-salt interface has a vertical position and from that point on the normal situation of fresh water flowing on top of salt water is present, the freshwaterlens being completed according to the dotted line in fig. 5.

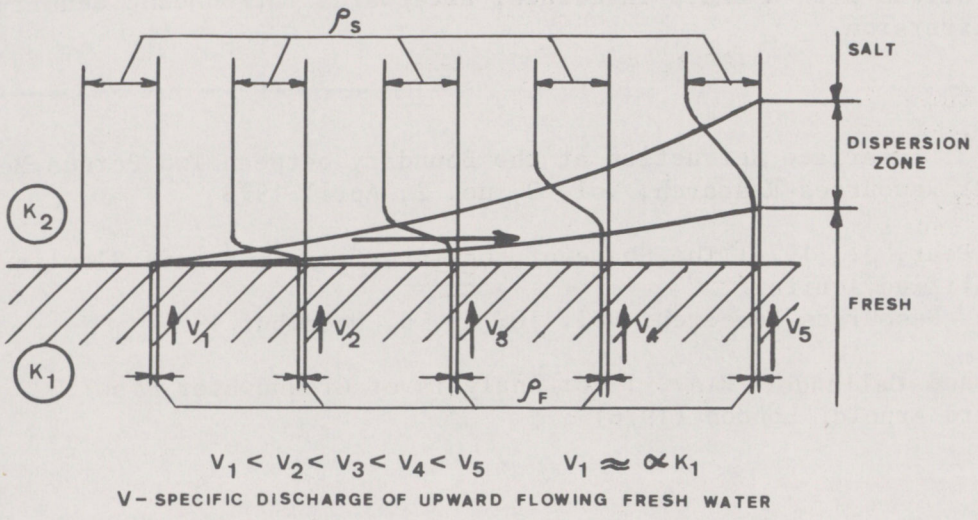

Fig. 7. Upseeping fresh water with dispersion zone

CONCLUSION

In the foregoing it is tried to give an explanation of the phenomenon of fresh water flowing seaward under a semi-pervious layer while on the top of that layer salt water is present. Considering the stream of fresh water to be potential, a form of the freshwaterbody is found in which part of this body consists of fresh water with salt water on top in the same aquifer. Owing to differences in pressure secondary currents will be introduced by which the fresh water is evacuated, while a dispersion zone between salt and fresh water comes into existance. This results in a final freshwaterlens as given in fig. 8.

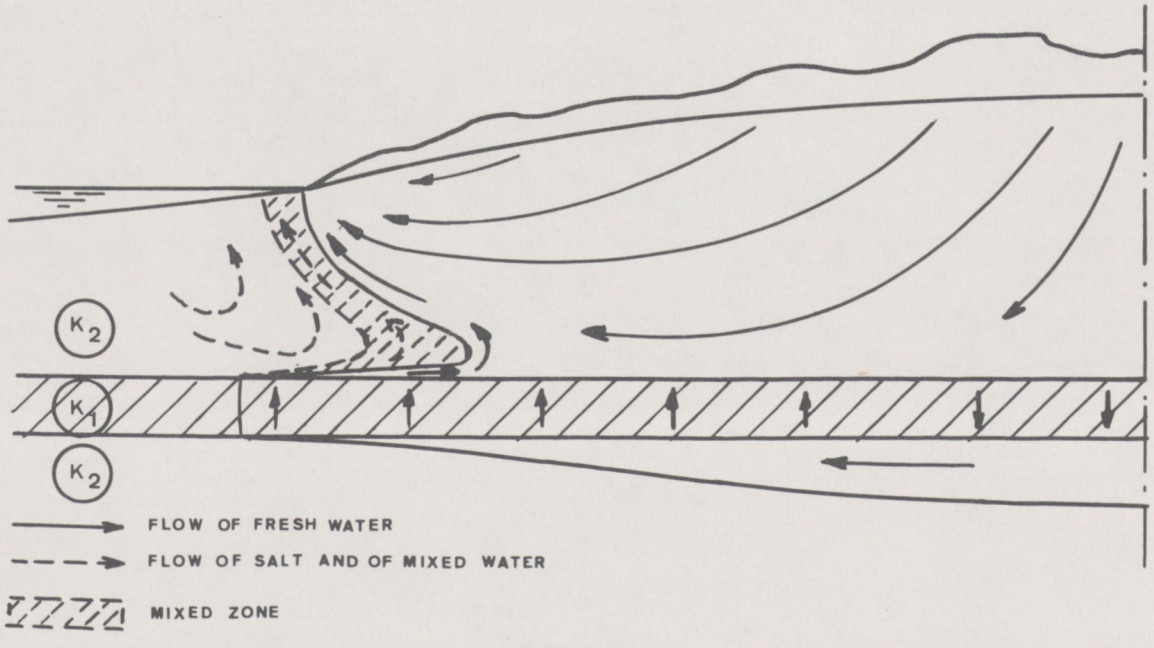

Fig. 8. Outflowing fresh water with secondary currents and mixed zone

In order to be able to compute this case it will be necessary, first to consider only potential stream with a sharp interface, afterwards introducing density currents and dispersion.

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APPENDIX

Notations

a, b	- height above reference level
g	- gravity acceleration
h_A, h_B	- water head with respect to reference level
h_{f_A}, h_{f_B}	- fresh water head in A, B, ... with respect to reference level
h_s	- salt water head with respect to reference level
k (k_1, k_2 ...)	- permeability (for medium 1, 2 ...)
Δn	- distance between streamlines
v (v_1, v_2 ...)	- specific discharge (in medium 1, 2 ...)
α	- relative density difference ($= \frac{\rho_s - \rho_f}{\rho_f}$)
β	- slope of underwater talus
ρ_f	- density of fresh water
ρ_s	- density of salt water
ϕ (ϕ_1, ϕ_2)	- angle between interface and horizontal
ψ (ψ_1, ψ_2)	- angle between interface and normal on boundary between different media

Slope of the interface

Let AB (fig. A1) be a small part of the interface.
Owing to balance of pressure in A:

$$(h_s - a)\rho_s g = (h_{f_A} - a)\rho_f g$$

$$h_{f_A} = \frac{\rho_s}{\rho_f} h_s - \frac{\rho_s}{\rho_f} a + a$$

$$h_{f_A} = \frac{\rho_s}{\rho_f} h_s - a \frac{\rho_s - \rho_f}{\rho_f} = \frac{\rho_s}{\rho_f} h_s - \alpha a$$

The same for B:

$$h_{f_B} = \frac{\rho_s}{\rho_f} h_s - \alpha b$$

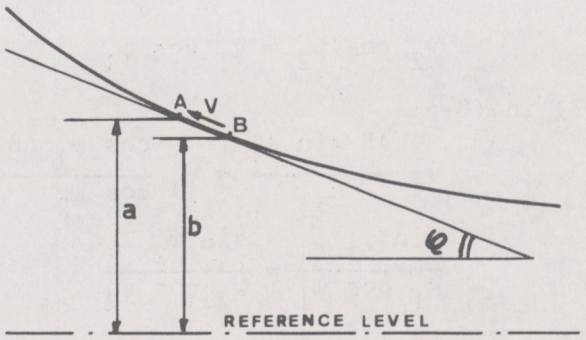

Fig. A1. Specific discharge along interface

$$h_{f_B} - h_{f_A} = \alpha(a-b) \qquad (1)$$

As $a - b = AB \sin\phi$, $h_{f_B} - h_{f_A} = \alpha . AB \sin\phi$ (2)

Equation of motion:

$$v = k \frac{h_{f_B} - h_{f_A}}{AB} = \frac{k.\alpha AB \sin\phi}{AB} = k\alpha \sin\phi$$

From which:

$$\sin\phi = \frac{v}{\alpha k} \qquad (3)$$

Interfraction of streamlines

Equation of motion:
For upper medium:

$$v_1 = k_1 \frac{h_B - h_D}{BD} \qquad (4)$$

For lower medium:

$$v_2 = k_2 \frac{h_C - h_A}{CA} \tag{5}$$

From (4) it follows:

$$h_B = v_1 \frac{BD}{k_1} + h_D = v_1 \frac{AB \sin\psi_1}{k_1} + h_D \tag{6}$$

From (5) it follows:

$$h_C = v_2 \frac{CA}{k_2} + h_A = v_2 \frac{AB \sin\psi_2}{k_2} + h_A \tag{7}$$

As AD and CB are parts of equipotential lines, it follows:

$$h_D = h_A \text{ and } h_B = h_C$$

This, in combination with (6) and (7) gives:

$$v_1 \frac{AB \sin \psi_1}{k_1} = v_2 \frac{AB \sin \psi_2}{k_2} \tag{8}$$

Continuity on the boundary between the media gives:

$$v_2 \cos \psi_2 = v_1 \cos \psi_1 \tag{9}$$

(9) in (8):

$$v_1 \frac{AB \sin \psi_1}{k_1} = v_1 \frac{\cos \psi_1}{\cos \psi_2} \frac{AB \sin \psi_2}{k_2}$$

$$\frac{\sin \psi_1}{k_1 \cos \psi_1} = \frac{\sin \psi_2}{k_2 \cos \psi_2}$$

$$\frac{\operatorname{tg} \psi_1}{\operatorname{tg} \psi_2} = \frac{k_1}{k_2} \tag{10}$$

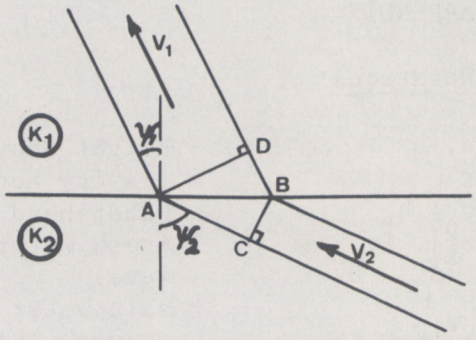

Fig. A2. Interfraction of streamlines on boundary between layers

Interfraction of interface

As the interface is a streamline eq (3) holds, so

$$\sin \phi_1 = \cos \psi_1 = \frac{v_1}{\alpha k_1} \tag{11}$$

$$\sin \phi_2 = \cos \psi_2 = \frac{v_2}{\alpha k_2} \tag{12}$$

From continuity it follows:

$$v_1 \cos \psi_1 = v_2 \cos \psi_2$$

With (11) and (12):

$$\alpha k_1 \cos^2 \psi_1 = \alpha k_2 \cos^2 \psi_2$$

$$k_1 \cos^2 \psi_1 = k_2 \cos^2 \psi_2$$

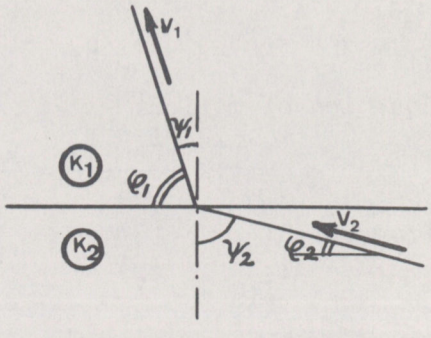

Fig. A3. Interfraction of interface on boundary between layers

As $\cos^2 \psi = \frac{1}{tg^2 \psi + 1} : \frac{k_1}{k_2} = \frac{\cos^2 \psi_2}{\cos^2 \psi_1} = \frac{tg^2 \psi_1 + 1}{tg^2 \psi_2 + 1}$

With (10) $\frac{k_1}{k_2} = \frac{\frac{k_1^2}{k_2^2} tg^2 \psi_2 + 1}{tg^2 \psi_2 + 1}$

$$\frac{k_1}{k_2} tg^2 \psi_2 + \frac{k_1}{k_2} = \frac{k_1^2}{k_2^2} tg^2 \psi_2 + 1$$

$$k_1 k_2 tg^2 \psi_2 + k_1 k_2 = k_1^2 tg^2 \psi_2 + k_2^2$$

$$k_1 (k_2 - k_1) tg^2 \psi_2 = k_2 (k_2 - k_1)$$

$$tg^2 \psi_2 = \frac{k_2}{k_1}$$

$$tg \psi_2 = \sqrt{\frac{k_2}{k_1}}, \text{ and (with (10)): } tg \psi_1 = \sqrt{\frac{k_1}{k_2}} \qquad (13)$$

Direction of interface on reaching the open (salt) water

Let the interface reach the open water in A. Another streamline at a small distance Δn reaches the open water in B.

$$BC = \Delta n \, tg \{\beta - (\tfrac{\pi}{2} - \phi)\} \qquad (14)$$

$$CD = \Delta n \, tg (\tfrac{\pi}{2} - \phi) \qquad (15)$$

$$b - a = BE = BD \sin \phi =$$

$$(BC+CD) \sin \phi =$$

$$\Delta n \left[tg\{\beta - (\tfrac{\pi}{2} - \phi)\} + tg(\tfrac{\pi}{2} - \phi) \right] \sin \phi \qquad (16)$$

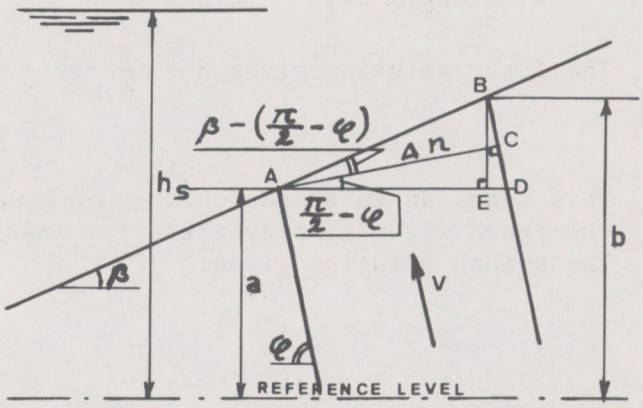

Fig. A4. Direction of interface on outflow into open salt water

Equation of motion:

$$v = k \frac{h_{f_C} - h_{f_B}}{CB} \qquad (17)$$

$h_{f_C} = h_{f_A}$ and follows from $(h_{f_A} - a)\rho_f = (h_s - a)\rho_s$

$$h_{f_C} = h_{f_A} = \frac{\rho_s}{\rho_f} h_s - a \frac{\rho_s - \rho_f}{\rho_f} = \frac{\rho_s}{\rho_f} h_s - \alpha a$$

Also $$h_{f_B} = \frac{\rho_s}{\rho_f} h_s - \alpha b$$

$$h_{f_C} - h_{f_B} = \alpha(b-a) \tag{18}$$

(18), (16) and (14) in (17) gives:

$$v = \frac{k.\alpha\ \Delta n\left[tg\{\beta-(\frac{\pi}{2}-\phi)\}+ tg(\frac{\pi}{2}-\phi)\right]\sin\phi}{\Delta n.tg\{\beta-(\frac{\pi}{2}-\phi)\}} \tag{19}$$

For the interface follows from (3): $v = \alpha k \sin\phi$ (20)

(20) in (19) results in

$$\frac{tg\{\beta-(\frac{\pi}{2}-\phi)\}+tg(\frac{\pi}{2}-\phi)}{tg\{\beta-(\frac{\pi}{2}-\phi)\}} = 1$$

$$tg\{\beta-(\tfrac{\pi}{2}-\phi)\}+ tg(\tfrac{\pi}{2}-\phi) = tg\{\beta-(\tfrac{\pi}{2}-\phi)\}$$

Now there are three possibilities

$$tg\{\beta-(\tfrac{\pi}{2}-\phi) = \infty$$

$$tg(\tfrac{\pi}{2}-\phi) = \infty$$

$$tg(\tfrac{\pi}{2}-\phi) = 0$$

The first solution gives $\beta - (\frac{\pi}{2}-\phi) = \frac{\pi}{2} + k.\pi$

$$\phi = k.\pi - \beta$$

This means an interface which coincides with the talus and which can only exist if the fresh water is prevented from emerging into the open water.
The second solution gives:

$$\frac{\pi}{2} - \phi = \frac{\pi}{2} + k.\pi$$

$$\phi = k.\pi$$

This means a horizontal interface and stagnant fresh water. This can only be stable if the fresh water is on top of the salt water and outflow from fresh water into the open salt water is prevented (as there is balance in pressure at the point where the interface reaches the open water, there will be a higher pressure in the fresh water than in the salt water in higher points of the talus).
The third solution yields:

$$\phi = \frac{\pi}{2} + k.\pi$$

This gives a vertical interface on reaching the open water. As this is possible without restrictions, this will be the normal case.

FIGURE CAPTIONS

fig. 1 Outflow of fresh water towards sea in two aquifers

fig. 2 Freshwaterlens under Dutch dune area

fig. 3 Flownet in a freshwaterlens

fig. 4 Crossing of a semi-pervious layer by the fresh-salt interface

fig. 5 Freshwaterbody if there is no mixing of fresh and salt water

fig. 6 Part of flownet in the region where upseeping fresh water enters a salt domain

fig. 7 Upseeping fresh water with dispersion zone

fig. 8 Outflowing fresh water with secondary currents and mixed zone

fig. A1 Specific discharge along interface

fig. A2 Interfraction of streamlines on boundary between layers

fig. A3 Interfraction of interface on boundary between layers

fig. A4 Direction of interface on outflow into open salt water

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