

Nordic Hydrology 3, 1972, 140–164

Published by Munksgaard, Copenhagen, Denmark

No part may be reproduced by any process without written permission from the author(s)

CHEMISTRY OF GROUNDWATER IN IGNEOUS ROCK AT ANGERED, GOTHENBURG

GUNNAR JACKS

Royal Institute of Technology, Stockholm

Within an area of about 5 km², 25 wells were drilled to depths of about 90 m for groundwater observation. The wells were sampled and analyzed, and from the results it could be concluded that the ionic composition is controlled by equilibrium between water and solids. The dominating weathering product is Ca-montmorillonite.

Establishing equilibrium takes approximately ten years. On higher grounds, equilibrium is far from being reached, and these waters are only a couple of years old. In lower positions of the terrain, sea water from glacial times is retained. Turn-over time of water is thus in the order of hundreds to thousands of years. The degree of washing out of the salt water is correlated to the transmissibility of the tectonic zones by which the water is found.

At Angered north of Gothenburg, Sweden, a large number of wells have been drilled in Precambrian gneisses. The purpose of the wells is to check groundwater levels during the construction of a new suburb. The rock will be penetrated by tunnels for water, sewage, cables, etc. The area covers about 5 km², and 25 wells have been drilled. The wells were drilled to a depth of around 80–90 m. In connection with the drilling, recovery tests were made after emptying the wells of water. From these tests the capacity of the wells can be calculated.

GEOLOGY

The area is situated east of the river Göta älv (see Fig. 1). Variation in altitude is from 40–100 m above sea level. The heights consist of bare rocks, while the valleys are covered with clays to depths of more than 20 meters. The area dips southwards towards the river Lärjeån. The rocks are mainly gneisses with interstratified bodies of basic rocks. According to Lundegårdh (1958) they can be classified as follows.

Granite gneisses in the western part:

quartz, micro line, oligoclase (15–20 % An), hornblende, biotite and from it secondarily formed chlorite. Hornblende may dominate, giving the gneiss the appearance of a hornblendegneiss.

Granite gneiss, porphyric in the eastern part:

quartz, microcline, oligoclase (25–30 % An), biotite.

Basites, postgenetic to the gneisses, generally along S-planes:

andesin (30–45 % An), hornblende, biotite.

In larger bodies, the plagioclase is more calcic (40–60 % An).

The tectonic pattern is characterized by overthrust movements from the west towards the east (Larsson 1967). This manifests itself in valleys in the direction of NNW to SSE. The tectonic pattern, together with the southward dip, directs groundwater flow to a regional north-to-south direction.

SAMPLING

The wells were investigated with electric conductivity logging. The conductivity logging was decisive for the water sampling. In most wells only one distinct type of water was found, while in others there were two, or even three types of waters, probably emanating from different crack-systems.

From 25 wells a sum of 28 samples were collected. Three wells were sampled at two depths. In tunnels in the northwestern part of the area, 9 samples were collected. Thus altogether 37 samples were included in the study.

The sampling was made with a special water-sampler constructed for the purpose of sampling deep wells. The sample can be taken at a chosen level and collected without being aerated. Samples were taken for analysis of Na⁺, K⁺, Ca⁺⁺, Mg⁺⁺, alkalinity, Cl⁻, SO₄⁻, NO₃⁻ and H₄SiO₄. An oxygen sample was also taken as well as a sample for Fe. pH and E_h (redoxpotential) were measured

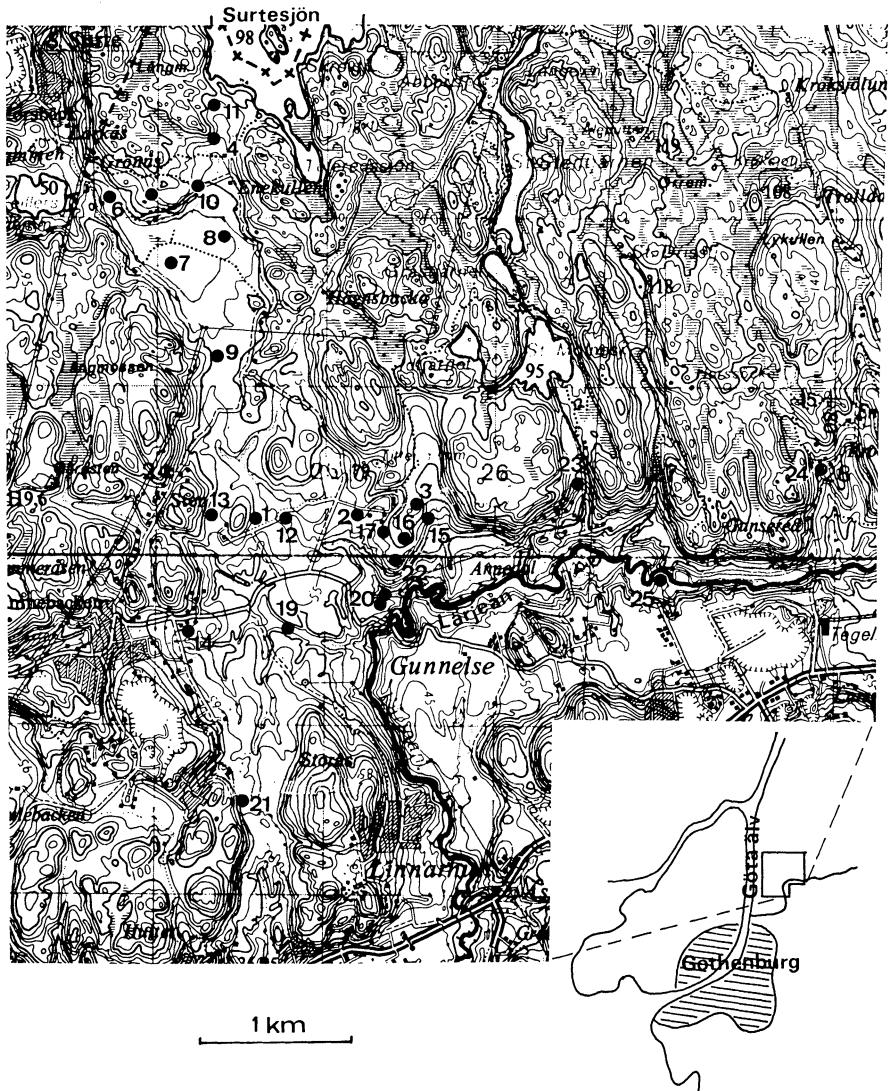


Fig. 1.
The Angered area, north of Gothenburg, Sweden.

in the field with a special electrode-pot which was connected with the sampler by means of a rubber tube. The connection permitted no entrance of air. The electrode-pot had a volume of 50 ml and was washed through with a volume of about 200 ml.

TREATMENT OF DATA

For the equilibrium calculations, activities instead of concentrations have been used. This means that when concentrations are growing each ion has not the possibility of expressing itself and the activity is decreased per unit concentration. This is expressed as:

$$a_{Ca^{++}} = \lambda [Ca^{++}]$$

λ is the coefficient of activity, in the actual waters ranging from 0.54–0.86 for Mg^{++} , Ca^{++} , and SO_4^{--} and from 0.83–0.96 for Na^+ , K^+ , Cl^- and HCO_3^- . The coefficient of activity is a function of ionic strength. The function used is that of Debye-Hückel:

$$-\log \lambda = A \cdot z^2 \sqrt{J}$$

A is a constant depending on temp. and solvent (here: water), z is the charge on the ion, and J is ionic strength.

$$J = \frac{1}{2} \sum_i c_i z_i^2$$

c is concentration in moles per liter.

It has been considered advisable to calculate with activities as the error from using concentrations would be in one direction as λ is less than 1 for all charged species.

The influence of complex formation on chemical equilibrium calculations has been considered negligible. For one of the more saline waters, that from well 25 out of a total of 34 mmoles dissolved ionic species only 0.35 mmoles consist of complexes. In this special case with a high concentration of Ca^{++} and SO_4^{--} , $CaSO_4(aq)$ will be the most abundant complex.

Well 25

pH	K ⁺	Na ⁺	Ca ⁺⁺	Mg ⁺⁺	Fe ⁺⁺	HCO ₃ ⁻	Cl ⁻	SO ₄ ⁻⁻
8.17	0.10	12.6	3.46	0.03	0.18	1.56	14.7	1.53 mmol/l

Complexes: CaSO ₄	0.25 mmol/l
CaHCO ₃ ⁺	0.05 mmol/l
NaSO ₄ ⁻	0.05 mmol/l

Calculations are made with data given by Garrels & Thompson (1962).

ORIGIN OF CHEMICAL COMPOSITION OF WATERS

The chemical compositions of the deep groundwaters in Angered are mainly dependent on three processes; atmospheric deposition, weathering reactions and sea water retention. In addition to these there is uptake by vegetation, and the breakdown of organic matter. The latter process gives CO_2 which dissolves in the percolating water thereby giving it a high weathering potency.

Salts from atmospheric origin

Salts are deposited with precipitation and as a dry deposition of small crystals. The salts resulting from precipitation have been measured for a number of years by the Meteorological Institute at the University of Stockholm. The sampling site is in the harbour of Gothenburg, thus some kilometers from the investigated area. Ten years' mean (1957-66) is:

K ⁺	Na ⁺	Ca ⁺⁺	Mg ⁺⁺	Cl ⁻	SO ₄ ⁻⁻	
0.32	3.01	1.62	0.56	4.86	7.50	ppm

These salts are found in groundwaters and in addition to them, the dry deposition. They are concentrated by evapotranspiration. Evapotranspiration can be estimated. The dry deposition is, however, very difficult to estimate. There is one way to calculate the salts of atmospheric origin in groundwaters if one assumes that Cl⁻ has only atmospheric origin. This is reasonable in igneous rocks where chlorine is found only as minor fluid inclusions in the minerals. One can then make the approximation that the other ions have had the same increments from dry deposition and evapotranspiration as Cl⁻ had. In Angered it seems reasonable to take the mean concentration of Cl⁻ in waters having an electric conductivity of less than 35.0 umhos/cm as a measure of Cl⁻ in ground water. Above 350 umhos/cm there is probably an influence of old sea water as seen from Fig. 2. The mean Cl⁻ concentration is 18.8 ppm. This means a concentration of 3.9 times higher than in precipitation. If the evapotranspiration is calculated with the formula of Tamm (Ångström 1958) it will be 452 mm. With a mean precipitation of 738 mm (Ångström, 1958) this will concentrate the 4.86 ppm Cl⁻ in precipitation to 12.5 ppm. This gives a dry deposition of 50 % of the deposition with precipitation. With the above assumption the concentrations of the different ions in groundwater having an atmospheric origin are calculated:

K ⁺	Na ⁺	Ca ⁺⁺	Mg ⁺⁺	Cl ⁻	SO ₄ ⁻⁻	
1.25	11.7	6.32	2.18	18.8	29.2	ppm
0.03	0.51	0.31	0.18	0.53	0.61	meq/l

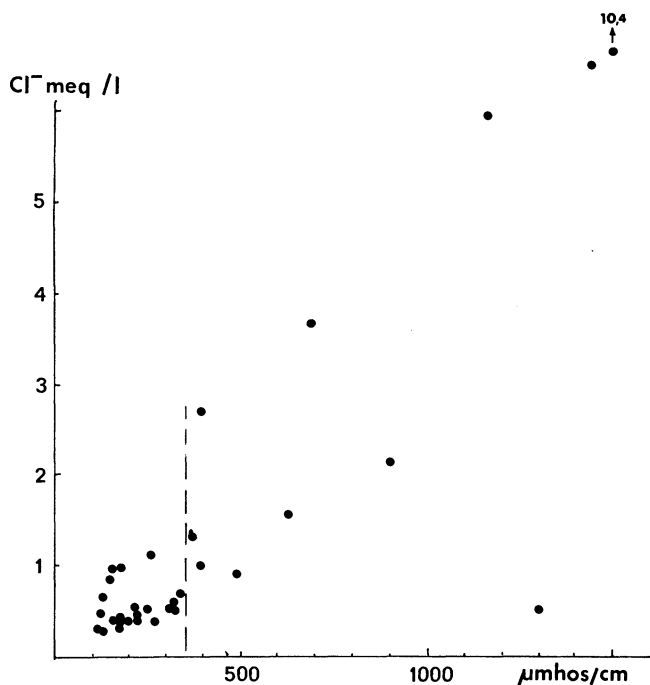


Fig. 2.
Plot of Cl^- against electric conductivity.

Weathering reactions

When the precipitation enters the ground its content of dissolved carbon dioxide is equilibrated to the 0.03 % in the atmosphere. When it passes the decaying organic matter in the upper parts of the soil, it is exposed to a considerably higher partial pressure of CO_2 . In the present study some of the most acid waters have a pPCO_2 of 1.70, that is to say, a partial pressure of 0.02 atm. or 2 % CO_2 . This gives the water a high capability of dissolving minerals in the ground.

Potassium is a major component in all igneous rocks and in Angered as well, in the form of microcline and biotite. However, microcline has a very low solubility and the groundwaters are soon saturated with K^+ . This is clear from Fig. 3 where K^+ is plotted against electric conductivity. K^+ is increased from 0.03 to about 0.09 meq/l. The stability diagram for K-minerals shows that at $\text{pH} - \text{pK}^+$ around 4 an equilibrium should be established between kaolinite and

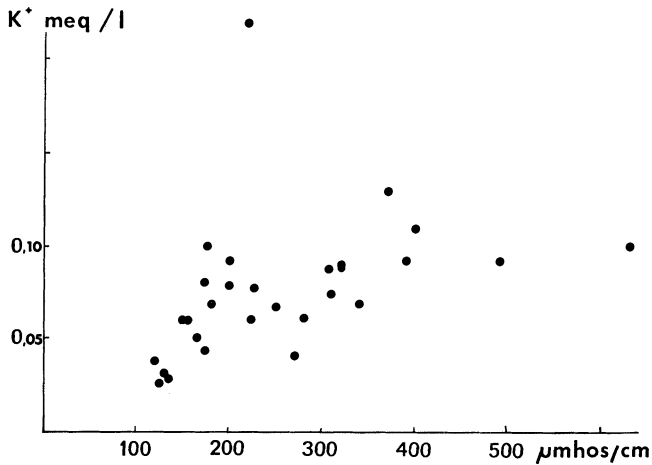
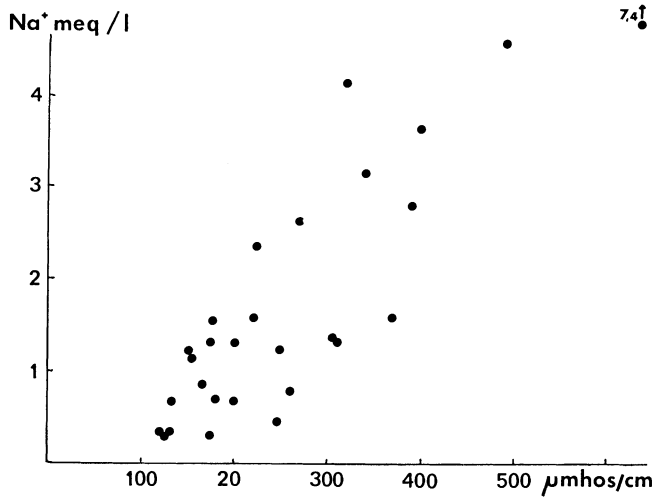
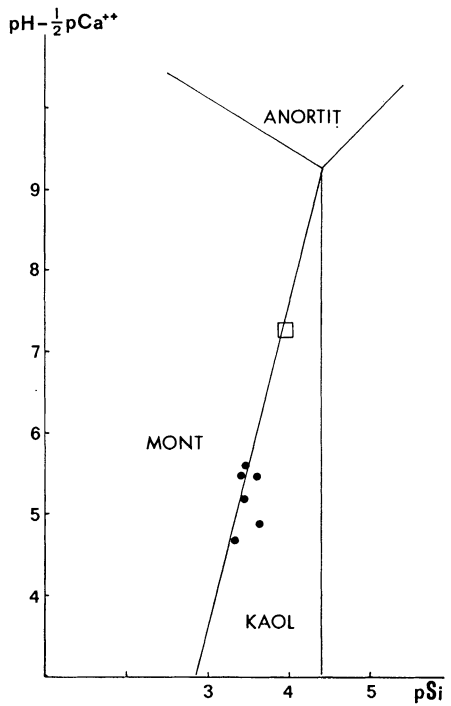
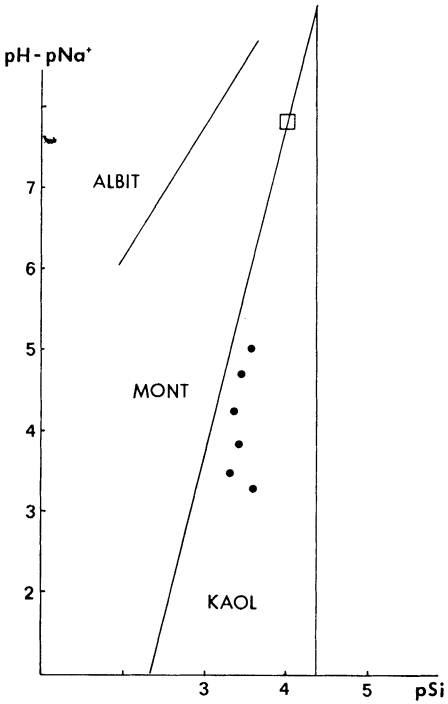
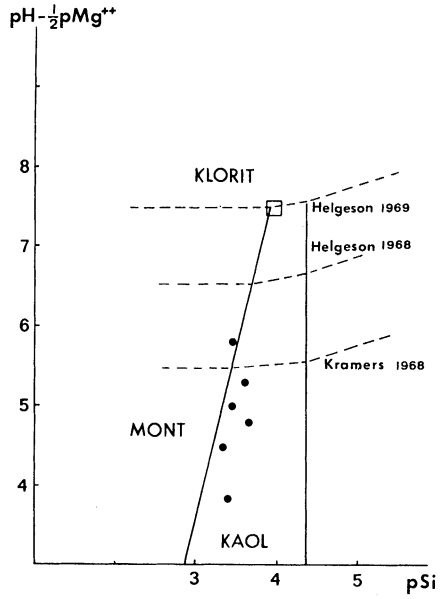
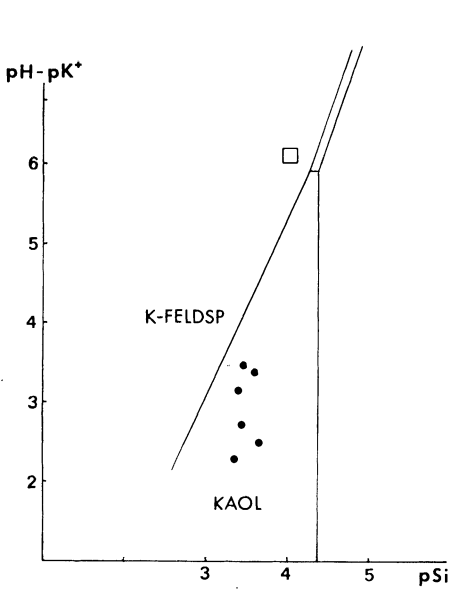


Fig. 3.
Plot of Na^+ and K^+ against electric conductivity.

Fig. 4.
Stability diagrams for potassium, magnesium, calcium and sodium silicates. Data from Helgeson (1969) extrapolated to 7°C , 1 atm. Six analyses are plotted in the diagrams. Squares means sea water conditions.

Chemistry of Groundwater in Igneous Rock



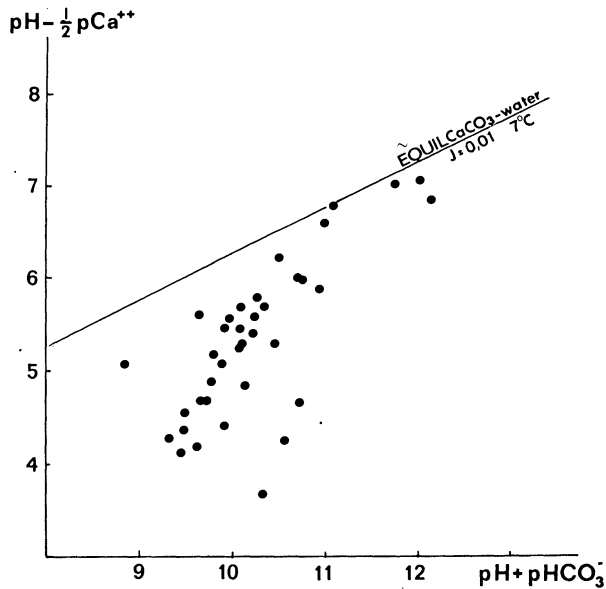


Fig. 5.

Stability of calcite at 7°C, 1 atm and ionic strength of solution = 0.01.

microcline (Fig. 4). Illite is also most probably a member of this equilibrium. That this equilibrium is really established is evident from the plot of $\text{pH} - \text{pK}^+$ against electric conductivity in Fig. 6. The relevance of this ionic ratio is based on the fact that in weathering, H^+ replaces cations in minerals. The replacement goes to a certain ratio, an equilibrium ratio. Equilibrium constants for reactions have been taken from Helgeson (1969) and extrapolated to 7°C.

Magnesium is found in biotite, chlorite and in hornblende in Angered. Of these minerals biotite is the most common. Biotite easily weathers to chlorite which is found in the actual rocks (Lundegårdh 1958). In Angered it seems as though the groundwaters are fairly rapidly saturated with respect to Mg^{++} (Fig. 8). This could be due to the formation of chlorite and Mg-montmorillonite as could be imagined from the stability diagram for Mg-minerals in Fig. 4. The stability of chlorite is somewhat uncertain. There are many thermodynamic data available; Helgeson (1967, 1969) has published some, Kramer (1968) and Karpov (1968) additional ones. The value of Helgeson (1969) emanates from Garrels who calculated the solubility of chlorite in sea water. Sillén (1967) has shown that chlorite probably controls Mg^{++} -activity in the oceans. Kramers

data fits a sweet-water environment. The author has collected a set of water samples from Håksberg mine in Central Sweden where chlorite is a most common mineral and found $\text{pH} - \frac{1}{2}\text{pMg}^{++}$ being just below 6 within a narrow range. Mg^{++} is added groundwaters in minor quantities in Angered. The amount carried in from the atmosphere lies near saturation for equilibrium between the secondary minerals kaolinite, chlorite and Mg-montmorillonite. This is obvious from Fig. 8 where 22 samples of low Cl^- -content are arranged in a staple-diagram with alkalinity as parameter. These waters should represent pure weathering conditions without the influence of old sea salts.

Helgeson (1969) also gives data for a K-Mg-illite. According to his data this illite should be stable only at high values of $\text{pH} - \text{pK}^+$ and $\text{pH} - \frac{1}{2}\text{pMg}^{++}$. On the sea floor illite is a common mineral and Sillén suggests that it has a regulating effect on the sea water composition. In Angered K-Mg-illite as mentioned above should generally not be stable as it requires higher contents of K^+ and Mg^{++} than are present.

Calcium and sodium are found in plagioclases. Plagioclases are easily weathered. Ca-rich ones are more apt to weather than Na-rich ones. Usually high temperature minerals are more sensitive to weathering than low temperature minerals. Ca-plagioclase for instance crystallizes before Na-plagio-

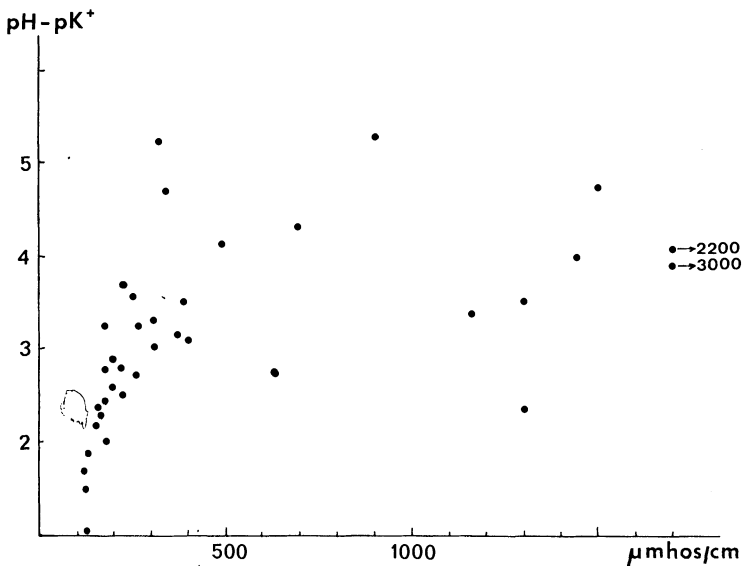


Fig. 6.
 $\text{pH} - \text{pK}^+$ against electric conductivity.

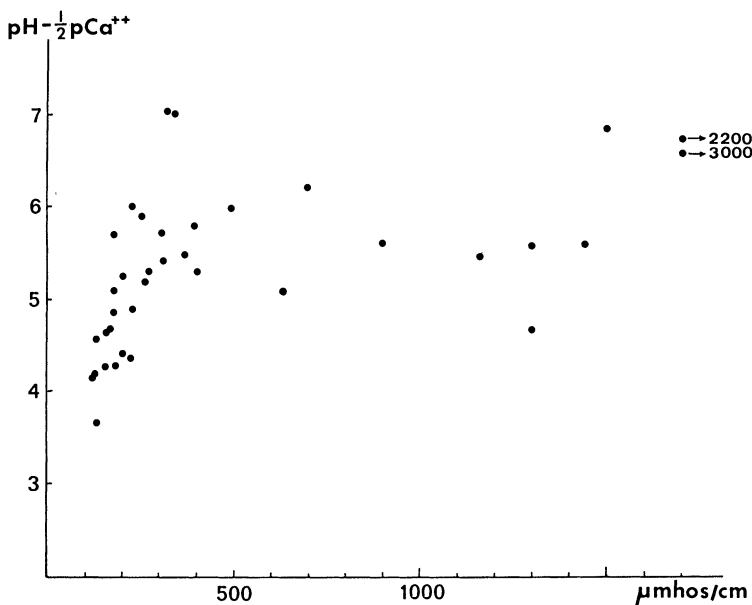
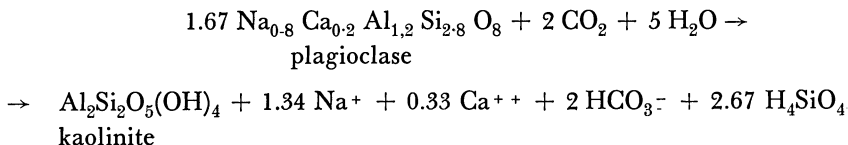
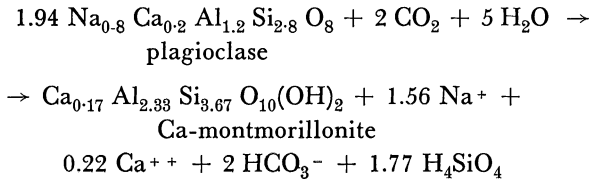


Fig. 7.
 $\text{pH} - \frac{1}{2} \text{pCa}^{++}$ against electric conductivity.

clase as stated by Bowen's scheme. In Angered, plagioclases have Ca-percentage (An-percentage) of 15-30 % in gneisses and up to 60 % in basites. The dissolution of plagioclase proceeds until some equilibrium is reached. For Ca^{++} this could be the saturation with respect to calcite or Ca-montmorillonite. For Na^+ an equilibrium between kaolinite and montmorillonite could be reached (Fig. 4). In Fig. 5 it is shown that very few waters have reached saturation with respect to calcite. The value of $\text{pH} - \frac{1}{2}\text{pCa}^{++}$ should be about 7 when solubility of calcite is reached. In Fig. 7 $\text{pH} - \frac{1}{2}\text{pCa}^{++}$ is plotted against electric conductivity. It can be imagined that weathering stops at $\text{pH} - \frac{1}{2}\text{pCa}^{++} \equiv 6$. This should fit the presence of montmorillonite (Fig. 4). Like Mg^{++} , Ca^{++} is not largely added to the water (see Fig. 8), which should mean that little kaolinite is formed and mainly montmorillonite. Reactions are for a 20 % An-plagioclase:



Chemistry of Groundwater in Igneous Rock



The formation of Ca-montmorillonite in favor of Na-montmorillonite is obvious from Fig. 9 where the ratios $p\text{Na}^+ - 1/2p\text{Ca}^{++}$ are plotted against electric conductivity. As salinity increases the ratios approach the theoretical exchange

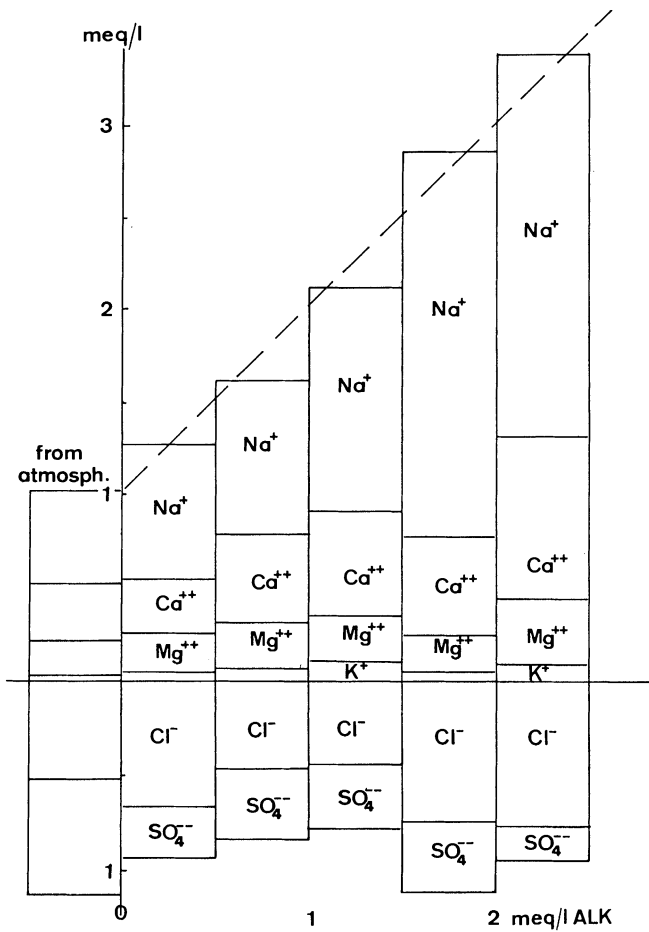


Fig. 8.

Staple diagram of cations and anions in waters with low content of Cl⁻.

ratio. In pure kaolinite weathering, the waters should dissolve Na and Ca in a molar ratio of 1.34/0.33 according to the reaction above. From Fig. 8 it is evident that this is not the case in Angered. The montmorillonite formation gives a molar ratio of 1.55/0.22 for the increment from weathering. This fits the actual data better.

Moreover in kaolinite weathering the ratio $\text{Na}^+/\text{H}_4\text{SiO}_4$ would be 1/2. There are only six Si-analyses available but in none of them is this ratio under the 1/1 ratio in montmorillonite weathering.

A third evidence for the dominating montmorillonite formation is the comparison of the composition of the waters with the An-percentage of the plagioclases in the area. An-percentage is $\text{Ca}/\text{Ca} + \text{Na}$ on a molar basis. Fig. 10 shows a deviation to a smaller ratio in waters than in plagioclases. This can be explained by the Ca-retention in Ca-montmorillonite.

Among the anions, HCO_3^- can originate from CO_2 dissolved in the water during the passage through the soil. HCO_3^- could also be added from the dissolution of calcite. In Fig. 11 the carbonate species are plotted against pH. If CO_2 is the only source, total carbonate species would remain constant through the pH-range. There are too few observations to give a clear picture but it seems to be a tendency for the total carbonate content to rise with pH at least for some waters. This comes into conflict with the fact that only a couple of

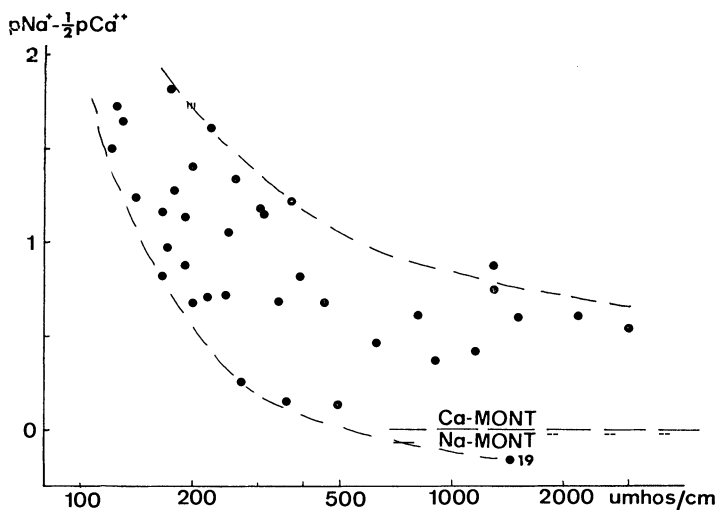


Fig. 9.

$\text{pNa} + \frac{1}{2}\text{pCa}^{++}$ against electric conductivity. Equilibrium between Ca-mont. and Na-mont. after Helgeson (1969).

Chemistry of Groundwater in Igneous Rock

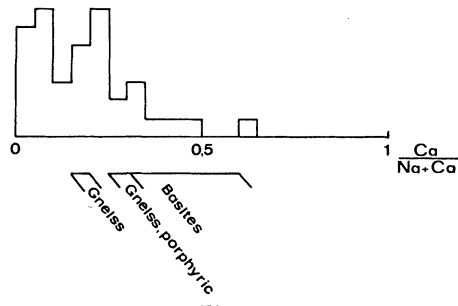


Fig. 10.

Ca/Ca+Na in waters and in plagioclases in the area.

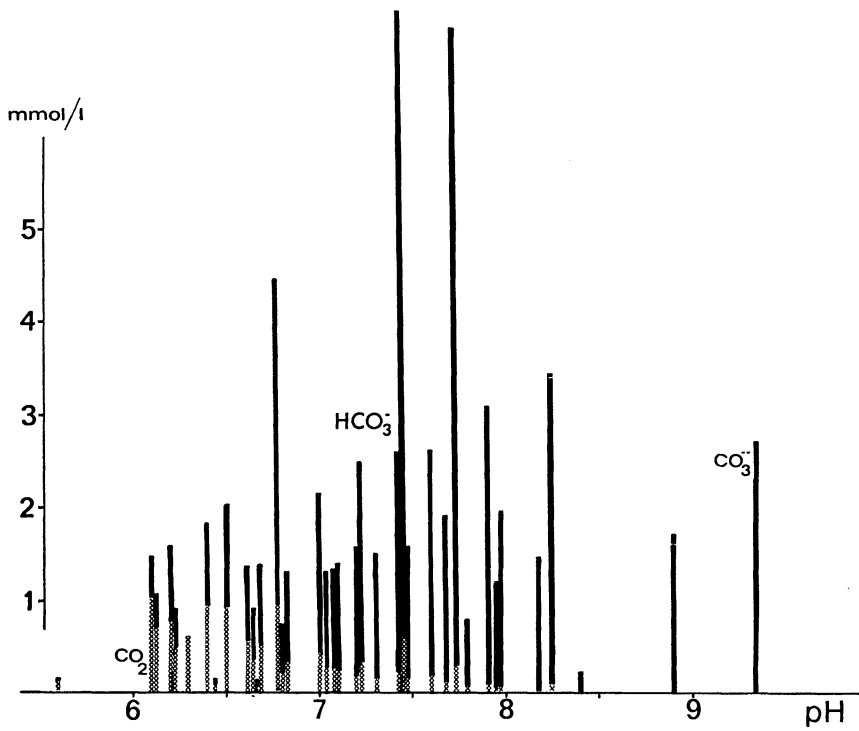


Fig. 11.

Carbonate species as function of pH.

the analyses show equilibrium with calcite. There is, however, the possibility that montmorillonite (or kaolinite) takes up Ca^{++} and thus makes the water unsaturated with respect to calcite after contact with the latter mineral.

Sulfate is moved in from the atmosphere in an amount which is double that found in groundwaters not affected by old sea salts. Sulfate is taken up by vegetation and also occupies exchange sites on minerals and organic matter.

Sea water retention

Some waters are quite saline. There are 5 samples with a Cl^- -content of over 200 ppm, clearly pointing out a sea water retention from glacial time. The highest marine shoreline is just above 90 m above the present sea level (Sandegren & Johansson, 1931).

The elution of the salt water progresses with different rapidity for each ion. Cl^- has almost no interactions with the minerals and it is thus a kind of tracer substance. Most of the other ions will be delayed in comparison with Cl^- . In Fig. 12 Na^+ and SO_4^{--} are plotted against Cl^- . Na^+ is seen to be washed out as rapidly as Cl^- . Below about 200 ppm Na^+ , there is obviously interac-

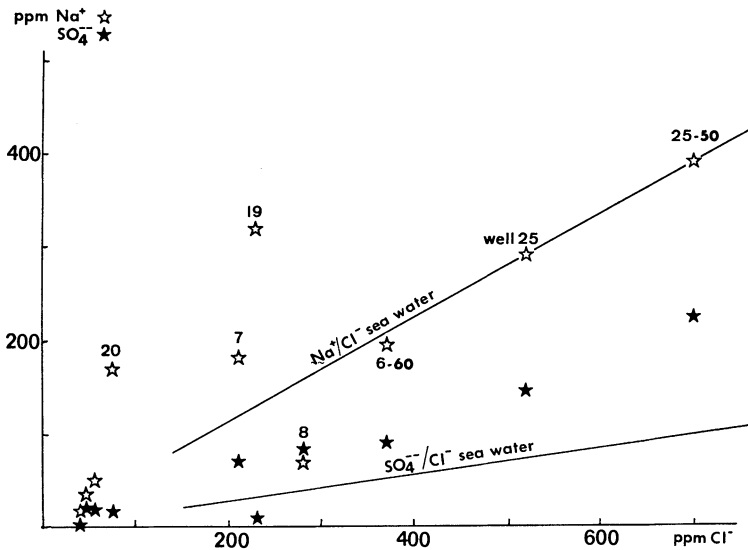


Fig. 12.

Na^+ and SO_4^{--} as function of Cl^- for the waters with a moderate to high electric conductivity.

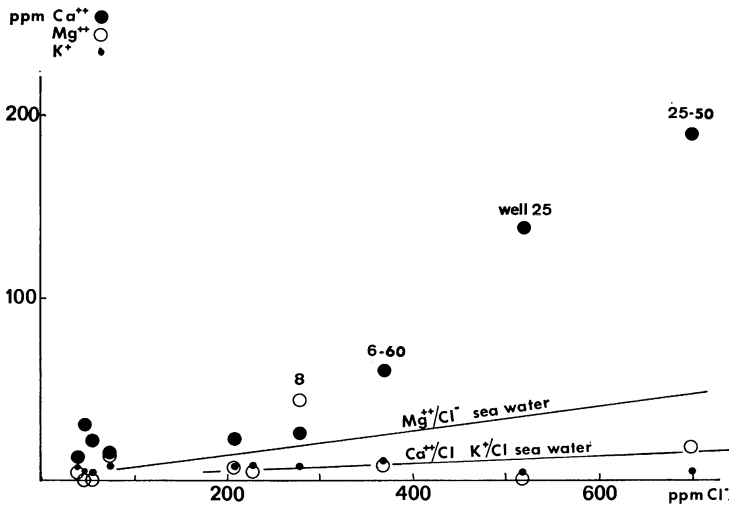


Fig. 13.

Ca⁺⁺, Mg⁺⁺ and K⁺ as function of Cl⁻ for the waters with a moderate to high electric conductivity.

tions between Na⁺ and solids in the rock. This point represents water from well 19, and from Fig. 9 it is seen that this water has a Na⁺/Ca⁺⁺ ratio just like the theoretical exchange ratio on montmorillonite. Probably this fact is the explanation for the high Na⁺-content.

Elution of SO₄^{- -} is retarded if compared with sea water ratio. It is well known that SO₄^{- -} is more prone to ionic exchange than Cl₋. In the salt waters, Ca⁺⁺ is present in higher concentrations than would be expected from sea water ratio (Fig. 13). This could be due to release of Ca⁺⁺ from Ca-saturated montmorillonite or from calcite (Fig. 9).

Mg⁺⁺ is less abundant than in purely diluted sea water. This could also be an effect of the action of montmorillonite. Montmorillonite could be under-saturated with respect to Mg⁺⁺ and thus tend to take it up from water.

K⁺ is not far from sea water ratio.

The interaction between minerals and water seems to be explained by the presence of a more or less cation saturated montmorillonite. The cations seem to be mainly Ca⁺⁺ and to a lesser degree Mg⁺⁺.

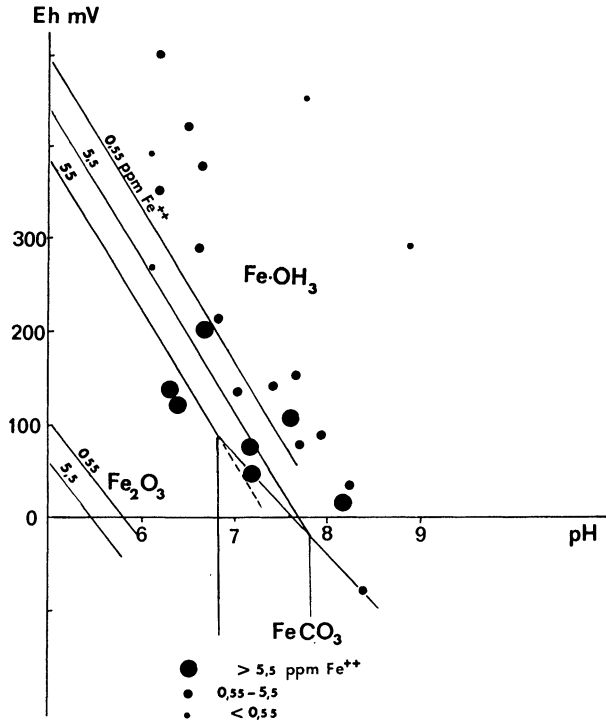


Fig. 14.

Redox diagram for iron species. Dots represent water analyses.

Redox-status of the waters

Redox-potential was measured on most of the samples. Oxygen-samples were taken as well. Almost all waters were oxygen-free. Analysis of iron in the water was made. Redox-potentials were plotted against pH (Fig. 14). In this diagram it is useful to draw stability lines for different iron solids and check up which solid determines the level of Fe^{++} in the solution. Data from Langmuir, Garrels & Christ and Stumm have been used. The analysis values for Fe^{++} are represented in the diagram by bigger or smaller dots. The pattern found fits fairly well with the presence of Fe(OH)_3 as the solid controlling Fe^{++} concentration. Some waters are probably in equilibrium with FeCO_3 .

CHEMISTRY AND TURN-OVER RATE OF WATERS

The processes within the aquifer determining the water composition, are weathering and salt water eluation.

Weathering per se is a slow proces. Eriksson (1968) has given the value of around ten years for reaching equilibrium in igneous rocks. This is for the Stockholm region. As precipitation contains more salts in Gothenburg it is reasonable to expect a somewhat faster equilibration in Angered.

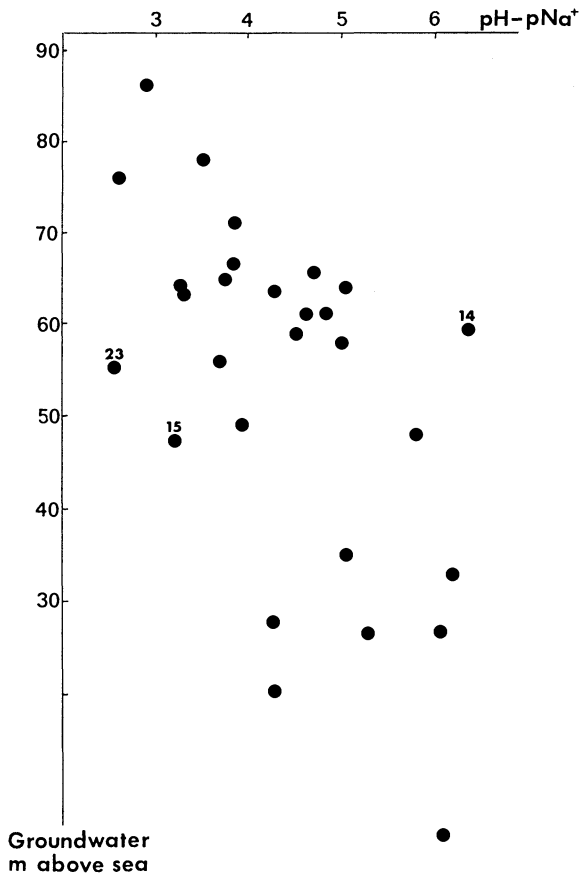


Fig. 15.

pH - pNa+ as function of groundwater level in the wells. Three points represent deeper groundwater bodies found in three of the wells. The upper interface found in conductivity logs is used as groundwater level for this water.

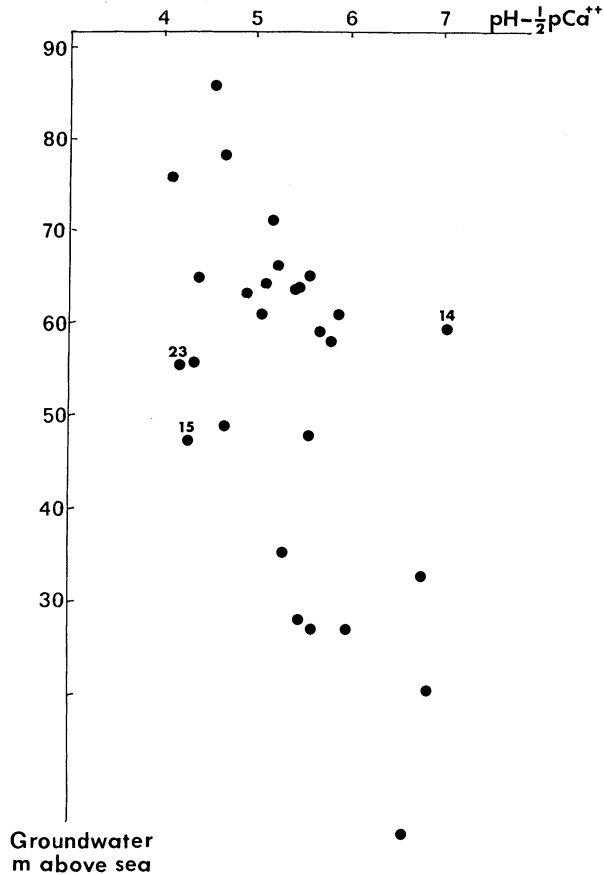


Fig. 16.

$\text{pH} - \frac{1}{2} \text{pCa}^{++}$ as function of groundwater level in the wells. Three points represent deeper groundwater bodies (see Fig. 15).

If $\text{pH} - \text{pK}^+$, $\text{pH} - \text{pNa}^+$, $\text{pH} - \frac{1}{2} \text{pCa}^{++}$ and $\text{pH} - \frac{1}{2} \text{pMg}^{++}$ are plotted against the groundwater level in the wells, this will give an illustration of the turn-over in the upper parts of the aquifer (Fig. 15, 16). The result shows a considerable variation as would be expected as groundwater level is only one of many parameters affecting the flow velocity of groundwater. Actually it is the gradients of the groundwater table that are driving the flow.

If $\text{pH} - \text{pNa}^+$ is studied, equilibrium is reached at first on a groundwater

level of 60–65 m above sea level. On the other hand there are wells below 50 m that are quite far from saturation. Wells 15 and 23 are extremes in this respect. Both are situated in narrow valleys with steep slopes and are probably subjected to locally large gradients of groundwater flow (see Fig. 1). Well 14 is an extreme on the other side. This well has a very thick cover (45 m) of clays which could cause the groundwater level to be less representative.

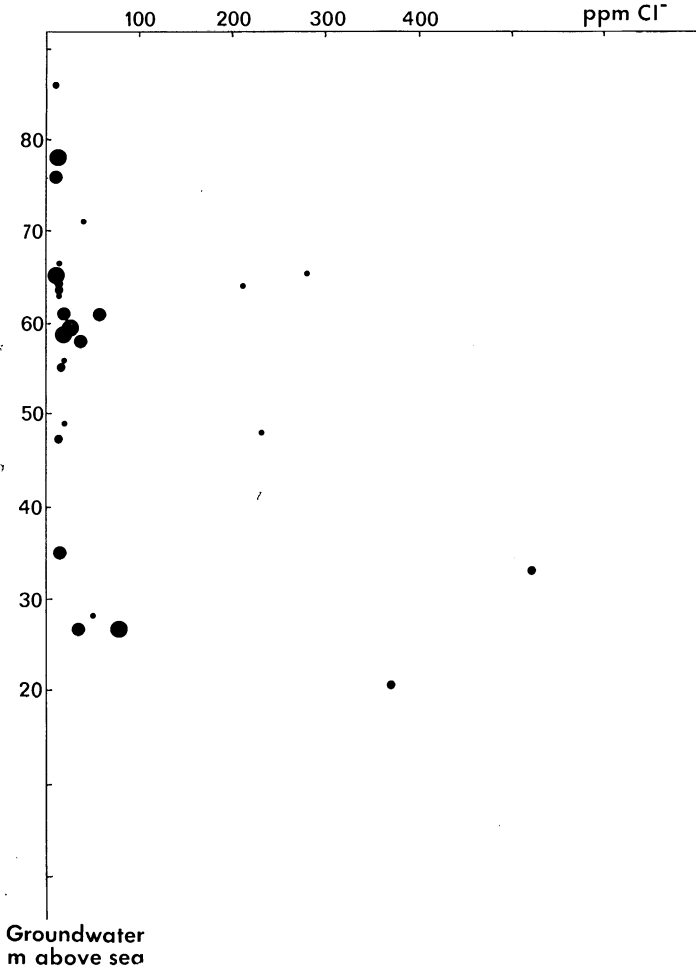


Fig. 17.

Cl⁻ plotted against groundwater level. Three points represent deeper groundwater bodies (see Fig. 15). Size of dot represents yield of well (see Fig. 18).

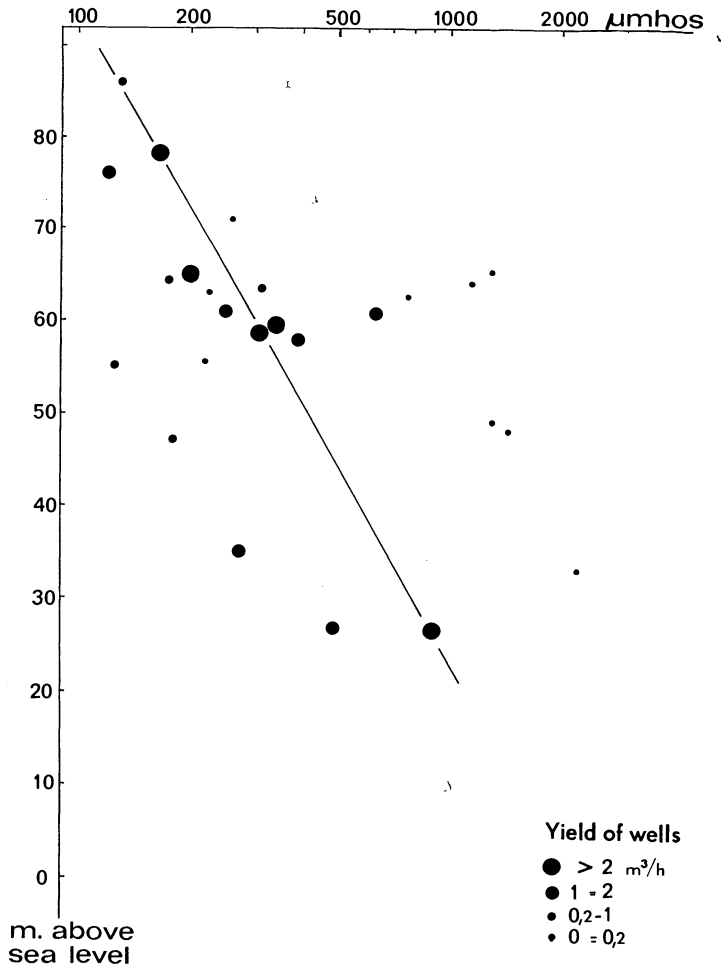


Fig. 18.
Electric conductivity plotted against groundwater level.

When comparing the establishment of equilibria with depth it seems that $pH = pK^+$ is first formed. Below the 45 m level equilibria are more or less reached.

The other process that illustrates the turn-over is the sea water retention. If the content of Cl^- is plotted against groundwater level (Fig. 17), it can be concluded that all wells with groundwater levels above 65 m are free from

retained Cl^- . In the plot, the yields of the wells have been represented as bigger or smaller dots. The wells with much retained Cl^- are all small wells situated on small tectonic zones with a low transmissibility for water. Larger wells on large tectonic zones contain less than 100 ppm Cl^- irrespective of groundwater level.

If electric conductivity is plotted against groundwater level in a diagram (Fig. 18), both the effect of weathering and the effect of sea water retention affect the picture as both affect the conductivity. The electric conductivity is plotted on a logarithmic scale, and it is interesting to note that the large wells fall almost on a straight line. They are probably not so affected as the small ones by local factors. The straight line seems to be an illustration of the exponential decrease of groundwater velocity with depth. Also in this diagram the phenomenon with the small wells having higher salinity is noticeable.

The influence of topography on groundwater flow and chemistry has been discussed in a generalized way. If a single tectonic zone is studied, the chemistry varies along the zone and towards depth in it. Wells 20 and 20 A are drilled in the same tectonic zone and have penetrated it at different depths. Fig. 19 shows how the content of Cl^- is doubled at a vertical distance of 25 m.

The wells 4, 8, 10 and 11 are drilled through the same overthrust zone or branches of it. The zone goes from the west shore of Lake Surtesjön southwards

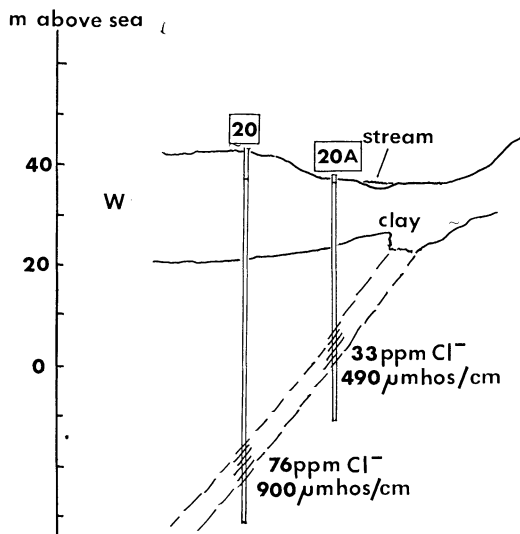


Fig. 19.
 Cl^- towards depth in a tectonic zone.

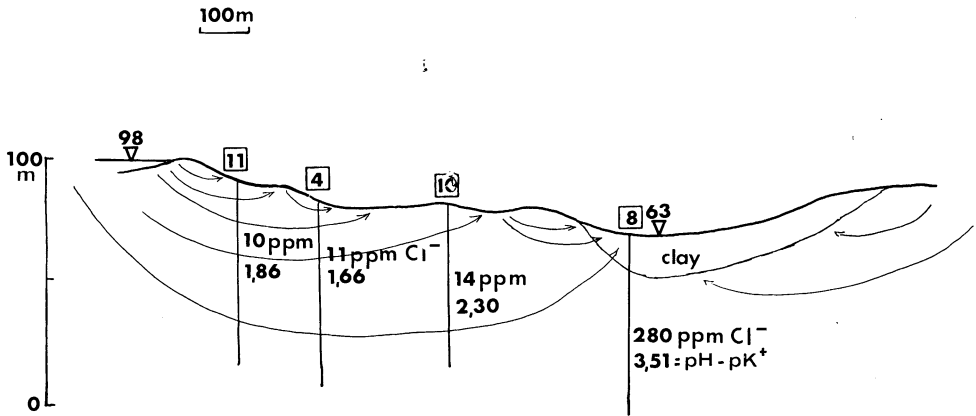


Fig. 20.

Cl^- and mineral-water equilibria expressed as $\text{pH} - \text{pK}^+$ along an overthrust zone.

towards Lövgärdet which is the plain around well 8. The upper part of the zone serves as inflow area for groundwater as sketched in Fig. 20. The lower parts of the zone are outflow areas according to Gustafsson (1968). The gradient is quite large, 35 m on 900 m or 4%. Cl_- is washed out in the wells 11, 4 and 10. In the centre of the outflow area, a body of more stagnant water is created around well 8 and there the chlorinity rises to 280 ppm. $\text{pH} - \text{pK}^+$ varies along the zone being very low in wells 11 and 4 indicating that the water in these positions is only a couple of years old. In well 10 the water is somewhat older and in well 8 it is near equilibrium and the chlorinity says that it has not been frequently changed since glacial times. Tritium content of this water was 2 units when the well was drilled.

CONCLUSIONS

The composition of the groundwaters depends mainly on two processes in the aquifer: weathering and retention of salt from glacial times.

Weathering adds Na^+ as main cation to the groundwater. K^+ and Mg^{++} go in solution in minor amounts. The weathering thus mainly affects the plagioclases. As the plagioclases are fairly rich in calcium in the area, Ca^{++} would be expected to go into solution in considerable amount if the weathering is a kaolinitization. However Ca^{++} is not added to waters in considerable degree.

Low silica content of the waters and stability considerations make it most probable that montmorillonite is the main formed secondary mineral. It has probably a varying saturation of Ca^{++} . Only in more saline parts of the aquifer can Na-montmorillonite be expected to be stable.

Salt water retention has affected different ions to a varying degree. It is known that Cl^- is readily washed out. Na^+ is seen to follow Cl^- closely. SO_4^{--} is delayed and is present in an amount double that expected. Ca^{++} is more common in the saline waters than in diluted sea water and Mg^{++} is less common. This is probably the effect of exchange on montmorillonite.

From the progress of weathering and the fact that salts were left from glacial time, some statements could be made on the turn-over of the aquifer. Some waters in high positions in the terrain are far from equilibrium with minerals and they are very young, a couple of years or so. In low positions of the terrain several hundreds ppm Cl^- (700 is the highest value) were found. As only about 20 ppm could come from atmospheric deposition and igneous rocks contain only minor amounts of chlorine, this must be a left-over from glacial times when the whole area was below the ocean surface. In these positions the water has thus been exchanged only a few times during the last 5,000 to 10,000 years. The larger tectonic zones are cleared to larger degree from saline water than the smaller ones. The high values of Cl^- are found in wells with yields less than 0.2 m³/hour.

In general the water flow is slow. This is an explanation for the proposed presence of montmorillonite. Only in the more intensively leached parts of the aquifer would kaolinite be found as the only or dominating weathering product. Loughnan (1969) has shown this for soil profiles.

ACKNOWLEDGMENTS

The author is indebted to Professor Yngve Gustavsson, Royal Institute of Technology, Stockholm and Professor Erik Eriksson, University of Uppsala for reading the manuscript. Fil. kand. Tom Lundgren, Swedish State Geotechnical Institute, has given very valuable information concerning the geology of the area. Jan Tobison, AB Hydroconsult, Gothenburg, assisted in sampling the wells.

REFERENCES

- Back W. & Barnes, I. (1965) Relation of electrochemical potentials and iron content to groundwater flow patterns. USGS Prof. paper 498-C, Washington.

- Eriksson, E. & Khunakasem, V. (1968) The chemistry of groundwaters. Ground Water Problems. Pergamon Press, Oxford.
- Garrels, R. M. & Christ, C. L. (1965) Solutions, minerals and equilibria. Harper & Row, New York.
- Garrels, R. M. & Mackenzie, F. T. (1967) Origin of the chemical composition of some springs and lakes. Equilibrium concepts in natural water systems. Amer. Chem. Soc., Washington.
- Garrels, R. M. & Thompson, M. E. (1962) A chemical model for sea water at 25°C and one atmosphere. Amer. Jour. of Science Vol. 260 Jan.
- Gustafsson, Y. (1968) The influence of topography on ground water formation. Ground Water Problems. Pergamon Press, Oxford.
- Helgeson, H. C. (1967) Solution chemistry and metamorphism. Researches in geochemistry. Wiley & Sons, New York.
- Helgeson, H. C. (1969). Thermodynamich of hydrothermal systems at elevated temperatures and pressures. Amer. Jour. of Science, Vol. 267.
- Karpov, J. K., Kasjik, C. A. & Pampura, V. D. (1968) Constants for thermodynamic calculations in geochemistry. Nayka, Moskva.
- Kramers, J. R. (1968) Mineral-Water equilibria in silicate weathering. XXIII International Geol. Congress, Prag.
- Langmuir, D. (1969) Geochemistry of iron in a costal-plain groundwater, New Jersey. USGS Prof. paper 650-C, Washington.
- Larsson, I. (1967) Anisotropy in precambrian rocks and post-crystalline deformation models. Lund studies in geography A 38. Gleerup, Lund.
- Loughnan, F. C. (1969) Chemical weathering of the silicate minerals. Elsevier, New York.
- Lundegårdh, P. H. (1958) Göteborgstraktens berggrund. SGU Ser. C 1957, Stockholm.
- Posokhov, Y. V. (1968) Factors forming the chemical composition of precipitation. Soviet Hydrology, Selected papers. Amer. Geophys. Union, Washington.
- Sandegren, R. & Johansson, H. E. (1931) Beskrivning till kartbladet Göteborg. SGU Ser. Aa 173, Stockholm.
- Sillén, L. G. (1967) Gibbs phase rule and marine sediments. Equilibrium concepts in natural water systems. Amer. Chem. Soc. Washington.
- Stumm, W. & Morgan, J. (1970) Aquatic chemistry. Wiley, New York.
- Wolff, R. G. (1967) Weathering of Woodstock granite. Amer. Jour. of Science, Vol. 265.
- Ångström, A. (1958) Sveriges klimat. Generalstabens Litografiska Anstalt, Stockholm.

Address:

Department of Land Improvement and Drainage,
Royal Institute of Technology,
KTH,
S-10044, Stockholm 70,
Sweden.

Received 13 March 1972.