# An integrated study of a Precambrian granite aquifer, Hvaler, Southeastern Norway.

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The Geological Survey of Norway (NGU) is performing an integrated study of the groundwater resources of the Precambrian Iddefjord Granite of Hvaler municipality in southeastern Norway. Linear fracture zones are identifiable from topographic maps, aerial photos, field inspection and geophysics. The two consistently most successful geophysical methods for identification of such zones have been total magnetic field and VLF measurements. However, investigations in a newly constructed subsea road tunnel and test-pumping of boreholes on land indicate that a topographic or geophysical anomaly is no guarantee of a substantially transmissive fracture zone. The permeability of the Iddefjord granite appears rather low; a background value of around $10^{-9}$ m/s has been calculated from test-pumping and from leakage into the Hvaler tunnel. The top 12 m or so of the granite appear to have an average permeability c.2-3 orders of magnitude higher.

The groundwaters can be divided into 4 hydrochemical types, based on the degree of rock-water interaction and saltwater mixing. Saline groundwaters appear to be derived from fossil or current seawaters. Bicarbonate buffering, anion exchange and calcite/fluorite saturation appear to be important processes controlling pH, bicarbonate, fluoride and calcium concentrations.

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## Introduction

The occurrence and flow of groundwater in igneous and high-grade metamorphic rocks is poorly understood. Most detailed studies have focussed on single localities, often in very low permeability terrain, and usually in connection with proposals for disposal of hazardous or radioactive waste or for nuclear/hydroelectric power. Relatively few studies have attempted to rationalise the occurrence and flow of groundwater in fractured aquifers on a coarser regional scale, from a practical, water-resources point of view. No really reliable guidelines for the location of boreholes in such aquifers exist.

The aim of the Geological Survey of Norway's (NGU) Hvaler project is to carry out an integrated study of the groundwater resources of a hard rock aquifer, encompassing the following:

(a) evaluation of methods for detecting transmissive fractures and fracture zones - aerial photography, topographical maps, field surveys and geophysical methods.
(b) evaluation of most important geological processes which determine the water-yielding capacity of fracture systems; e.g. earlier and current stress-fields, secondary mineralisation, neotectonic (post-glacial) fault movement and fracture development, overlying drift deposits.
(c) identification of hydrochemically distinct groundwater types, and their chemical evolution.
(d) evaluation of the use of hydraulic fracturing, explosives, acids or dispersing chemicals as methods for increasing the capacity of a borehole.
(e) development of standard methods for test pumping boreholes in fractured aquifers, and a standard programme for chemical analysis.

## Geology of the Hvaler Area

The Hvaler municipality consists of a group of islands (Hvalerøyene) in the mouth of Oslofjord in south-east Norway (Fig.1 and 2). The dominant lithology is the Precambrian Iddefjord Granite, described by Oxaal (1916). The granite consists of 13 separate plutons (Pedersen & Maaløe 1990), some of the youngest

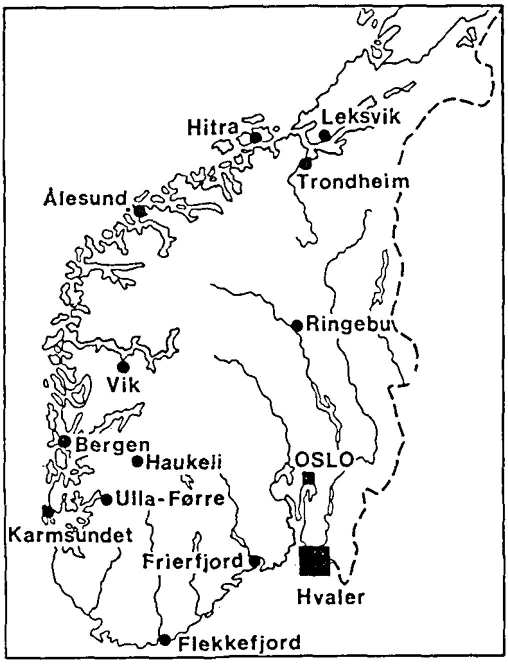



Fig. 1. Map of Norway showing location of Hvaler area, and other sites named in text (after Banks et al. 1992).

of which yield a Rb/Sr age of 918 ± 7 million years, corresponding to the end of the Sveconorwegian orogeny. Quartz, microcline and plagioclase are the dominant minerals in the granite. Accessory minerals include biotite, hornblende, muscovite, iron-oxides, chlorite, apatite, titanite, zircon (Pedersen & Maaløe 1990) and occasionally garnet. The granite commonly includes basic clots, pegmatites and xenoliths of gneissic host-rock. In some areas the xenolith content may be extremely high; in the new Hvaler tunnel the gneiss content reached some 55 % (Larsen 1990, Banks et al. 1992). Ramberg & Smithson (1971) describe the Iddefjord granite as a tabular intrusion on the basis of geophysical evidence.

In common with most high latitude areas, the Hvaler area has no regional development of a heavily degraded layer of weathered granite. Relatively fresh bedrock outcrops over large areas of the islands, often showing signs of glacial scouring, or other sub-glacial features such as potholes. The Iddefjord granite area is dissected by a pattern of linear valleys resulting, at least in part, from preferential glacial erosion along zones of fractured and crushed rock. These valleys are usually partially infilled by Quaternary deposits, rendering the surface outcrops of the fracture zones unexaminable. The linear channels between the islands of the Hvaler group, such as the two straits between Vesterøy and Asmaløy and the channel between Asmaløy and Kirkeøy (Fig.2), are also believed to have arisen by such a process. The origin of the fracture zones themselves is uncertain. It is likely, however, that they date from an early period of the granite's history, as a result of regional tectonic stresses or stresses related to emplacement and cooling of the granite. The fracture pattern is likely to have been reactivated or modified several times during its history; for example, during the Permian opening of the Oslo rift, post-rifting strike-slip movements along the Oslo graben boundary fault (Størmer 1935 - see Fig.2), and possibly even by glacial and post-glacial stresses.

The islands have undergone substantial post-glacial isostatic uplift in the past 10,000 years or so. The highest marine limit is c. 170 m above current sea-level (Selmer-Olsen 1964). The islands have therefore only emerged from the sea within the last several thousand years. The hydrogeological environment of the rocks encountered onshore is thus only likely to have differed significantly from those in the subsea tunnel during that period.

The Hvaler islands' Quaternary deposits are to a large extent limited to the lineament-controlled valleys, and consist mainly of shallow marine (or littoral) sands and silts (Olsen & Sørensen 1990). Limited deposits of peat, wind-blown sand, and coarser gravelly/pebbly beach deposits can be found on the southern part of Kirkeøy. The massive areas between the lineament valleys consist of bare bedrock or bedrock with a thin covering of humus.

## Permeability of fractured aquifers

It is a common assumption that the most pronounced fracture zones identifiable in a crystalline rock terrain are those that will yield the most water. Such fracture zones are typically located by their topographic expression, by use of remote sensing (Ronge 1988, Ericsson 1988) and by various geophysical techniques, such as electromagnetic induction, VLF profiling, seismic refraction, magnetic anomaly detection, resistivity profiling and georadar (Mullern 1980, Henkel & Eriksson 1980, Davis

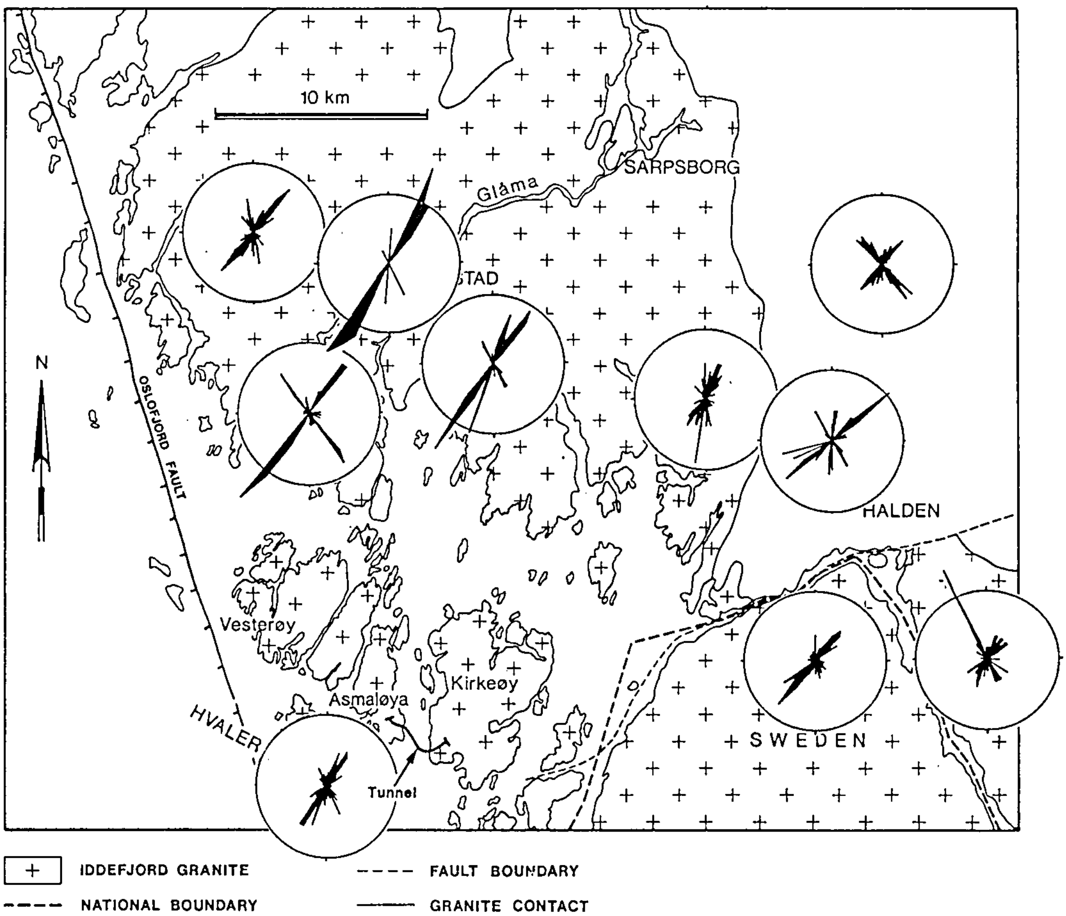



Fig. 2. The Iddefjord Granite area, with rose diagrams showing lineaments (granite, number of lineaments (N) = 341, 8 diagrams; gneiss, N = 92, 2 diagrams) identified on 1:50,000 topographic maps. Right hand side shows total lineaments, left hand side shows total length.

& Annan 1989). Some hydrogeologists, however, have gone beyond the «biggest is best» hypothesis for fracture zone transmissivity, and have examined the influence of tectonic stress. Some workers (Larsson 1972, Huntoon 1986, Rohr-Torp 1987) have identified a regional correlation between the past tectonic stresses which created or reactivated a fracture pattern and the permeability of the constituent fractures/fracture zones, while others (Olsson 1979, Selmer-Olsen 1981, Carlsson & Christiansson 1987) have found a correlation between permeability and the current stress field within the rock.

While the common assumption that major fracture zones are significantly transmissive has been shown to be true in some cases (e.g. Carlsson & Olsson 1977, Skjeseth 1981), many recent studies have cast doubt upon the general applicability of such a rule. Recently drilled boreholes in topographically prominent fracture zones in a variety of Precambrian and Palaeozoic bedrock lithologies on the island of Hitra, and in geophysically prominent zones in Palaeozoic schists in Leksvik municipality (Rueslåtten et al. 1984a & b) have yielded spectacularly little water. Furthermore, during the course of tunnel excavation in Norway, particularly of subsea tunnels, it has been noted that the largest fracture-zones crossed by tunnels often give rise to very few water leakage problems. The majority of large water leakages tend to arise from smaller fracture zones, or individual fractures/groups of fractures in relatively massive bedrock. Examples (see Fig.1) are described from Ålesund (Olsen

& Blindheim 1987), Ulla-Førre (Bertelsen 1981), Lysaker-Slemmestad, near Oslo (Løset 1981), Flekkerøy (Gulbrandsen 1989) and Karmsund (Kluver 1983) and are summarised by Nilsen (1988, 1990) and Banks et al. (1992). This phenomenon is ascribed, in many cases, to the largest fracture zones being 'tightened' by secondary clays resulting from weathering or hydrothermal activity. It is also noted that the most 'leaky' subsea tunnels have included the Godøy (Storås 1988) and Frierfjord (Kluver 1983) tunnels, where, significantly, there was very little clay mineralisation and, in the case of the Godøy tunnel, no major fault zone was crossed.

A similar phenomenon is observed in tropical areas. Studies of the weathered (saprolite) layer overlying fresh bedrock (e.g. Acworth 1987) indicate that while a low degree of weathering can be effective at destroying the bonding between mineral grains to give a gravelly texture with enhanced permeability, a higher grade of weathering results in extensive alteration to clay minerals and a substantial decrease in permeability. Although the alteration processes involved at Hvaler (presumably low-temperature hydrothermal alteration or deposition - Storey & Lintern 1981) are somewhat different to those in tropical weathering, observations from the recently completed Hvaler tunnel suggest that clay-alteration may also have a substantial effect on the permeability of fracture-zones at some depth within a bedrock aquifer (Banks et al. 1992).

### *The Hvaler tunnel*

In Hvaler, a 4 km subsea road tunnel was constructed in 1988-89 to link the islands of Asmaløy and Kirkeøy (Figs.2,14). Prior to excavation, major fracture zones were located by the use of aerial photos, acoustic profiling and seismic refraction (Taugbøl & Øverland 1987, Larsen 1990). On encountering these zones during tunnel construction, the majority were found to be of low transmissivity and filled with clay minerals. These clay minerals contained 50 - 100 % smectite with extreme swelling capacities on contact with water (up to 400 % free swelling). Smectite fracture fillings have been found in many Norwegian hard-rock lithologies and areas (Selmer-Olsen 1964). They are probably the result of low-temperature hydrothermal alteration by Mg- and Ca-rich fluids (Storey & Lintern 1981), and might be expected to be rather efficient at tightening fractures. Alteration processes would be particularly intense along major fracture zones due to (a) their presumably high pre-alteration permeability and (b) the high specific surface area of the gouges and breccias within the zones. Water inflows to the Hvaler tunnel tended to occur through lesser fractures or fracture groups, in most cases not detected by preliminary investigations (Fig.3). Calculations from Lugeon testing (method, e.g. Moye 1969) and total inflow to the tunnel indicated a 'background' permeability of $10^{-9}$ - $10^{-8}$ m/s. The permeability in the vicinity of the major leakages is estimated to be 100 to 1000 times higher, around $10^{-7}$ -$10^{-5}$ m/s The case study is detailed in Larsen (1990) and Banks et al. (1992).

It appears, therefore, that the identification of major fracture zones by geophysical and remote sensing techniques may not be a satisfactory method of locating groundwater resources in hard rock aquifers. No current geophysical method can adequately distinguish between water-transmissive and clay-filled fracture zones.

## Hydrogeology and borehole yield

### *Borehole statistics*

NGU maintains an archive of data on groundwater boreholes in bedrock in Norway. Statistical data for the Hvaler municipality, and the entire area of the Norwegian Iddefjord Granite within four map sheets 1913 I-IV (roughly the area of Fig.2), is presented in Fig.4 and Table 1. It is worth noting that the distribution of borehole depths is roughly symmetrical, the mean coinciding with the median. The distribution of borehole yields is, however, highly skewed towards low yield. The mode is considerably lower than the median, in turn considerably lower than the mean. The mean value is forced up by the existence of a few boreholes with very high yields. Such a phenomenon has also been recorded from the Bergen area by Ellingsen (1978), from the Østfold area (Bryn 1961) and from the Drøbak area (Rohr-Torp 1987). The *median* yield thus has greatest significance for planning groundwater abstractions.

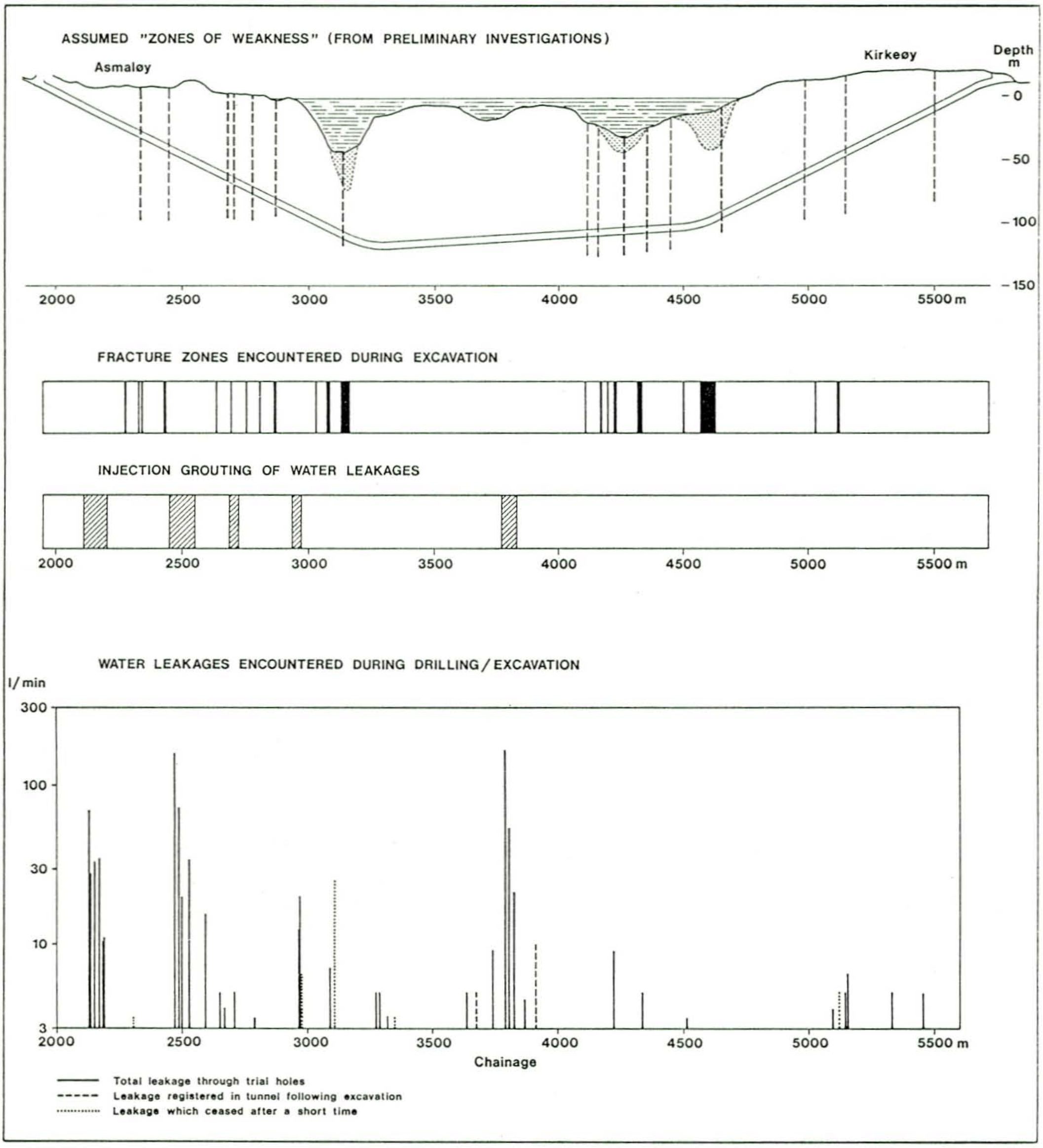



Fig. 3. Correspondance between presumed (from geophysics and aerial photos) and actual fracture zones, and water leakages into the Hvaler tunnel (after Larsen 1990, and Banks et al. 1992). Leakages are shown as the total inflow from all probe holes at a specific chainage (note logarithmic scale).

If one drills a borehole on the expectation of a 50 % chance of achieving the *mean* yield, one will be severely disappointed. The mean only begins to have significance when planning a combined abstraction consisting of several boreholes.

The mean yield for the Iddefjord Granite appears to be in the region of 1100 l/hr, and the median yield around 500 l/hr. This median figure is a little less than Bryn's (1961) analysis of the granites of Østfold (dominated by the Iddefjord granite), but is considerably less than that reported for other Scandinavian granites, where the median yield is often quoted as around 1000 l/hr (Persson et al. 1979, 1985a,b). The mean and median yields for the Hvaler area are less than those for the Iddefjord granite as a whole. There is no observab-

Table 1. Correlation between depth and borehole yield. All boreholes in Iddefjord granite with yield information (N = 310), quartile intervals according to yield.

| | | Depth (m) | Yield (l/hr) |
|---|---|---|---|
| 1st quartile interval | Mean | 57.5 | 81.4 |
| | Max | 130 | 200 |
| | Min | 10 | 0 |
| 2nd quartile interval | Mean | 55.4 | 333.2 |
| | Max | 140 | 500 |
| | Min | 7 | 200 |
| 3rd quartile interval | Mean | 56.1 | 802.4 |
| | Max | 128 | 1400 |
| | Min | 10 | 500 |
| 4th quartile interval | Mean | 53.7 | 3261 |
| | Max | 126 | 10000 |
| | Min | 10 | 1440 |

Mean yield = 1124 l/hr
Median yield = 500 l/hr

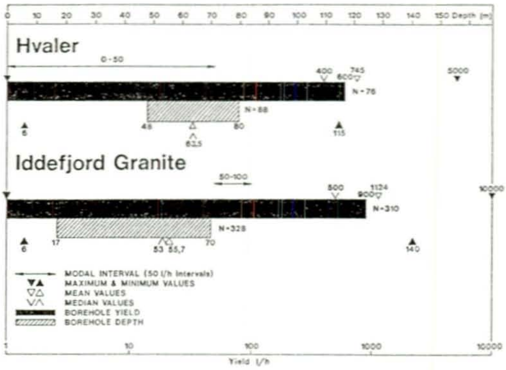



Fig. 4. Statistical distribution of yield and depth of boreholes registered on NGU's archive. Bars represent the smallest interval encompassing 67% of boreholes.

le correlation between borehole depth and yield (Table 1), perhaps due to a driller continuing to considerable depths if no good supply is found, but stopping if he meets a good supply at relatively shallow depth.

The statistical figures must be used with caution. The statistics are likely to be overestimated because:

(a) many drillers use crude methods of assessing a borehole's yield, or at best a short term pumping/recovery test. Long term capacity may be considerably less.
(b) the archive includes some (although by no means a majority) boreholes with capacities artificially increased by explosives or hydraulic fracturing.
(c) some unsuccessful, low-yield boreholes may not be reported to NGU.

## *Fracture mapping*

Fracture mapping in the Hvaler area has been carried out on three scales (Banks & Rohr-Torp 1991). Lineaments have been identified and measured from 1:50,000 topographical maps covering a large portion of the Iddefjord granite area and some of the adjoining gneiss area (Fig.2). The results show a very dominant NNE or NE direction over the entire granite area, with subsidiary NNW and N directions in some sub-areas. Lineaments on the well-exposed northern half of the island of Kirkeøy have also been mapped using aerial photos (Fig.5), and a field survey of fractures within the same area has been carried out (Fig.6). With increasingly detailed studies (i.e. maps to aerial photos to field measurements) one obtains increasingly complex results, the variation in fracture direction increases, and a general NW direction becomes increasingly prominent over the NE/NNE direction. At all scales, however, the NE/NNE direction can be identified in most sub-areas, as can the lack of E-W oriented fractures.

Despite the combined results of the field survey giving an apparently complex result, at each individual locality a well-defined fracture pattern consisting of two or three fracture sets (typically steeply dipping) could commonly be identified. A large amount of variation *between* localities did, however, occur.

On the NW peninsula of Kirkeøy (left half of Fig.5), the largest, aerial-photo-identifiable fracture zones have four major directions, (NW, NNW, NE-NNE and ENE), dividing the terrain into a mosaic of smaller blocks. Such orthogonal, or double-orthogonal patterns have frequently been observed in ancient granitic terrains (e.g. Tirén & Beckholmen 1989). Within each block, tectonic stresses would be significantly modified by the presence of the major fracture zones, and by the interaction of adjacent blocks, thus giving a plausible explanation for the increasingly complicated fracture pattern at smaller scales.

It appears that two types of fracture zone can be distinguished; those whose component fractures are approximately parallel to the zone and those comprised of fractures lying oblique to the main trend of the zone. This

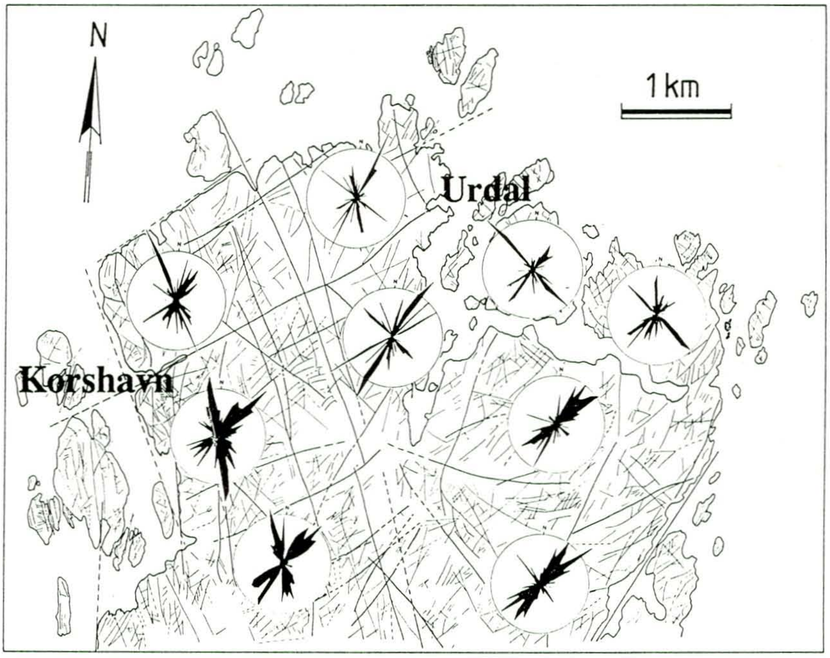



Fig. 5. The northern part of Kirkeøy, showing lineaments identified from areal photos. Rose diagrams show lineaments identified from the photos (N = 1637). Right hand side shows total lineaments, left hand side shows total length.

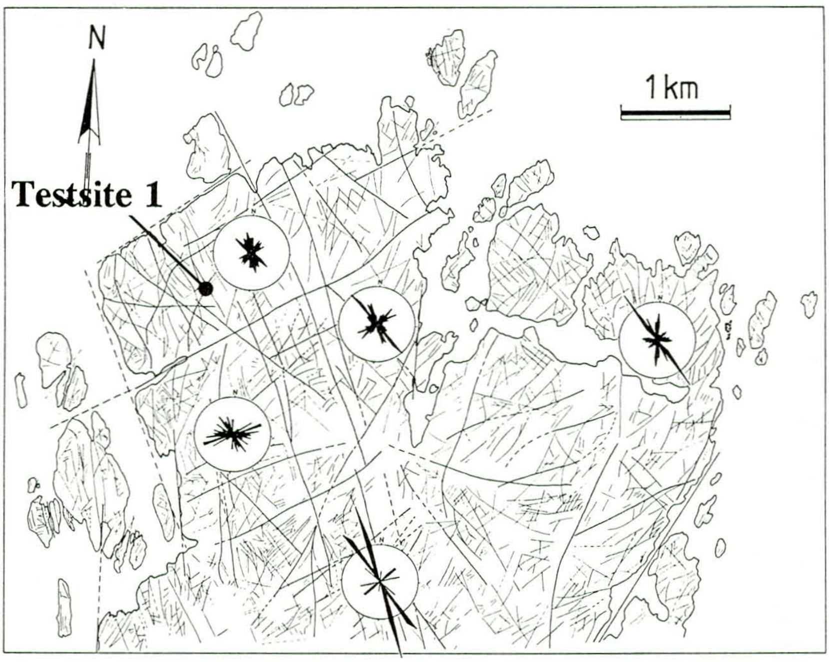



Fig. 6. The northern part of Kirkeøy, showing lineaments identified from aerial photos. Rose diagrams show fracture strikes identified from field measurements (only fractures with dip $\geqq$ 45° are included). N = 1167, 31 localities.

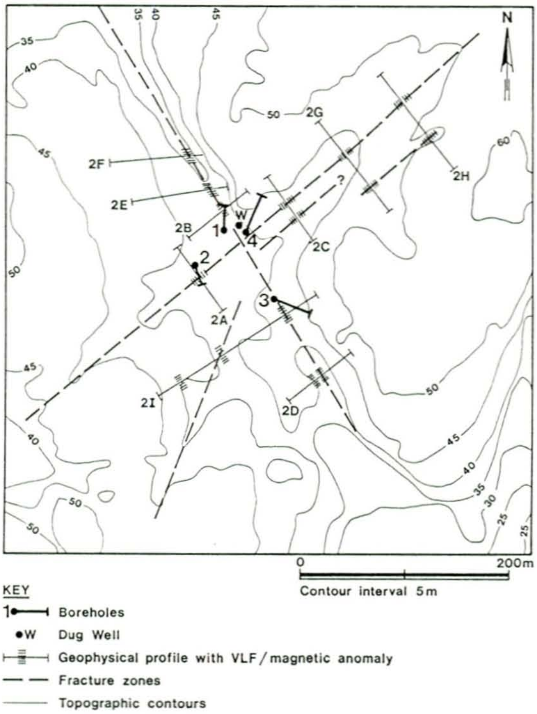



Fig. 7. Map of testsite number 1, SE of Pulservik.

latter type are another possible reason for the discrepancies between Figs. 5 & 6.

Fracture minerals observed during the fracture survey have included quartz, chlorite, epidote, calcite, fluorite, lepidolite and pyrite. Of these, epidote has a tendency to occur preferentially on fractures with strike 20-40° and c.130°, and fluorite (usually in association with calcite, and often with epidote) has only been recorded on fractures with strike 23-40° along the Korshavn-Urdal fracture zone (Banks & Rohr-Torp 1991).

Most fractures surveyed in the field were steeply dipping. The occurrence of near-horizontal unloading fractures appears to be very variable. In some locations, such as road cuttings on the small islands north of Vesterøy, horizontal unloading joints are well-developed to a depth of several metres, but at other locations, the joint set appears very poorly developed.

## Geophysical investigations

Several potential test-drilling areas have been identified on the NW peninsula of Kirkeøy. All the test areas have been investigated by geophysical methods including electrical resistivity profiling, total magnetic field measurements, georadar and very low frequency electromagnetic induction (VLF). All methods, except georadar, showed significant anomalies at the major topographical lineaments under some circumstances (Lauritsen & Rønning 1992). However, the VLF measurements were often disturbed by 'noise' from power-transmission cables. Total magnetic field measurements appeared to be the most consistently reliable of the various geophysical methods, relying on the oxidation of the granite's magnetite content to haematite along fracture zones.

## Drilling programme

Drilling is planned at the selected test sites, and has been completed at Testsite 1, SE of Pulservik (Fig.7). Here, four boreholes, each c. 73 m deep have been drilled. Details are given in Table 2 and Banks et al. (1991). Two holes, numbers 1 and 2, have been drilled into each of the two intersecting fracture zones at 73° from the horizontal. They were expected to cross the fracture zones at ca. 50 - 60 m depth, assuming the fracture zones to be approximately vertical. The other two holes (3 and 4) were drilled at 60° from the horizontal into relatively massive granite (i.e. away from the fracture zones). Observations during drilling (November 1990) were carefully recorded and the holes were geophysically logged (electical resistivity, self-potential, fluid resistivity and fluid temperature; point measurements at half-meter intervals) in September 1991 (Fig.8).

Hole number 1 encountered several minor water-bearing fractures, as well as some 'dry' fractures. The main fracture zone, characterised by fast drilling, reddish cuttings and a powerful anomaly in the resistivity log, was encountered, as expected, between 54 and 62 m. It is interpreted as a substantial crush-zone. The majority of the fracture zone appeared to be of rather low permeability, a major inflow only being met at c. 62 m (i.e. at the very base of the zone), and a very minor one at c.55½ m. The location of the main zone of water flow at the boundary of the fracture zone with relatively massive rock is a feature also observed by Ahlbom & Smellie (1989) in Sweden. The inflow position is confirmed by the fluid resistivity log, run in September 1991,

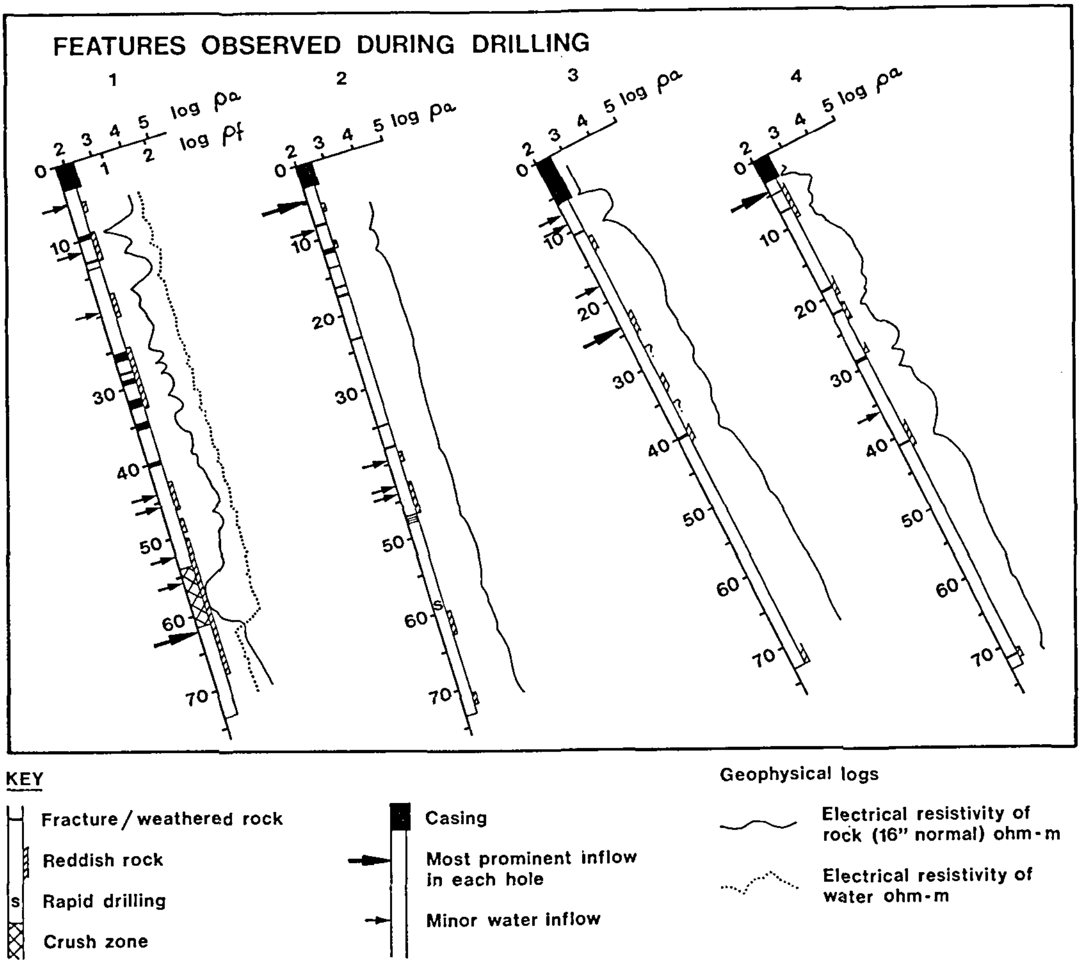



Fig. 8. Features observed during boring of testholes 1-4. Geophysical logs. Depth scale in m. $\rho_a$ = apparent rock resistivity; $\rho_f$ = fluid resistivity.

after the pumping tests of May 1991. The log shows clearly that 62 m is the lowest major flow horizon, and has yielded relatively fresh water with a resistivity about 30 ohm-m (c.330 µS/cm) at in-situ temperature. Below that level, the borehole contains a more saline 'residual' water with a resistivity of 7 ohm-m (1400 µS/cm). The temperature of the fluid column in borehole 1 ranges from a minimum of c. 7.0°C at 15-19 m up to 8.2°C at 71 m, corresponding to a vertical gradient of 0.022°C/m. The temperature shows a very small 'step' by the 62 m level, adjacent to the inflow horizon.

In hole 2, the only signs of a possible fracture zone at the expected depth (46-63 m) were slightly elevated boring rates and small minima in the electrical resistivity log at c.47 m and 58 m, together with minima in the self-potential log at c. 46 and 53 m. There was no sign of any major fracture zone as encountered in hole 1. Thus, either the fracture zone had a substantial deviation from the vertical (unlikely, given the zones' direct cross-cutting of the topography, the near-vertical nature of the zone in hole 1, and of the zones encountered in the Hvaler tunnel) or it died out at a relatively shallow depth. The only significant water inflows were from shallow, near-surface fractures.

Holes 3 and 4, as expected, encountered no significant fracture zones. Hole 3 appeared to meet a substantial water inflow at c.24 m,

Table 2: Borehole details, testsite 1

| Borehole no. | Depth (m) | Azimuth | Fall | Rest water level (m under well top) 25/5/91 | Yield (l/hr) with Pumping water level = 50m |
|---|---|---|---|---|---|
| 1 | 73.5 | 3° | 73° | 3.75 | 360 |
| 2 | 73.5 | 157° | 73° | 2.65 | 65 |
| 3 | 73 | 110° | 60° | 1.13 | 40 |
| 4 | 73 | 24° | 60° | 0.86 | 22 |
| Dug well | 1.7 | vertical | | 0.74 | |

but subsequent test-pumping failed to detect this, indicating it to be an unconnected fracture with limited storage capacity. Otherwise, in holes 3 & 4, the significant inflows (although relatively small) were from rather shallow fractures - subsequently confirmed by test-pumping.

The correlation between geophysical logs and drilling observations was rather good. The correlation of reddish cuttings with fractured zones should also be noted (the cuttings otherwise being grey in fresh granite), presumably due to the presence of oxidised Fe from preferential weathering along fracture planes.

## *Test pumping*

The four holes were capacity-tested in May 1991 (Banks et al. 1991), using four methods.

(a) low-rate step-test pumping using a Grundfos MP1, 1½" diameter, submersible pump, and subsequent recovery. Yield and water level were measured. Due to rapid water level decline, the pump frequency, but not the yield, was constant for each step. Results are marked ● (drawdown) and ○ (recovery) on Fig.9a.
(b) rapid emptying of the borehole down to c.50 m by a large capacity 4" diameter submersible pump, and subsequent pumping with a constant pumping water level of 50 m. The yield was measured. Suitable for higher capacity holes. Marked + on Figs.9.
(c) rapid emptying of the borehole down to 50 m, pump switched off for a given time (c. 10 - 30 mins.) followed by up-pumping of the amount of water accumulated in the hole during the interval. Suitable for low-capacity holes. Marked + on Figs.9.
(d) monitoring of water level's recovery after methods (b) or (c). Marked x on Figs.9.

As the contribution ($Q_B$) from the storage capacity of the borehole itself is significant in comparison with the total yield (Q), the contribution from the aquifer ($Q_A$) is calculated by:

$$Q_A = Q - Q_B = Q + \pi r^2(\delta h/\delta t)$$

where r = the borehole's radius (0.07 m) and $\delta h$ is the change in water level in a short time interval $\delta t$ (positive for rising water level)

Thus, one can plot $Q_A$ vs. water level as in Figs.9. In methods (a) and (d) above, both water level and $Q_A$ vary. In methods (b) and (c) $Q_A$ is measured for an approximately constant water level.

It is widely accepted that the specific capacity (F) of a borehole in bedrock is approximately proportional to the 'apparent, local transmissivity' (T) of the fractures feeding the hole. The Logan approximation (Kruseman & De Ridder 1989), the Moye (1967) and Banks (1972) methods, the Krasny (1975) method and the Carlsson and Carlssted (1977) method, all assume the following equation:

$T = F/c$ where c is a constant.

As the specific capacity of the borehole is merely the sum of the specific capacities of the individual 'feeder fractures', it should be possible to estimate the apparent transmissivity of these fractures from a simple $Q_A$ vs. water-level diagram (Fig.10). Such a technique (Banks 1991) appears to be most applicable to relatively low-capacity boreholes where the low transmissivity of the feeder fractures is the controlling factor for the borehole's yield, rather than the storage properties of the wider aquifer, and where a 'pseudo-equilibrium' is established relatively rapidly. In higher capacity holes, storage effects become important, and $Q_A$ vs. water-level plots typically display hysteresis between drawdown and recovery (Banks 1991). The gradient G of the $Q_A$ vs. h plots in Figs.9, at a given water level, is related to the total specific capacity F of all the fractures below the water level, by:

$$F = G/\sin \alpha$$

where $\alpha$ is the fall of the borehole.

As regards the constant c, this has been calculated by Moye's (1967) method as around 1.4 for Lugeon testing of short sections of a borehole, by Logan (see Kruseman & De Ridder 1989) as 0.82 (ostensibly for porous flow aquifers), and by Krasny (1975) as 0.91. Carls-

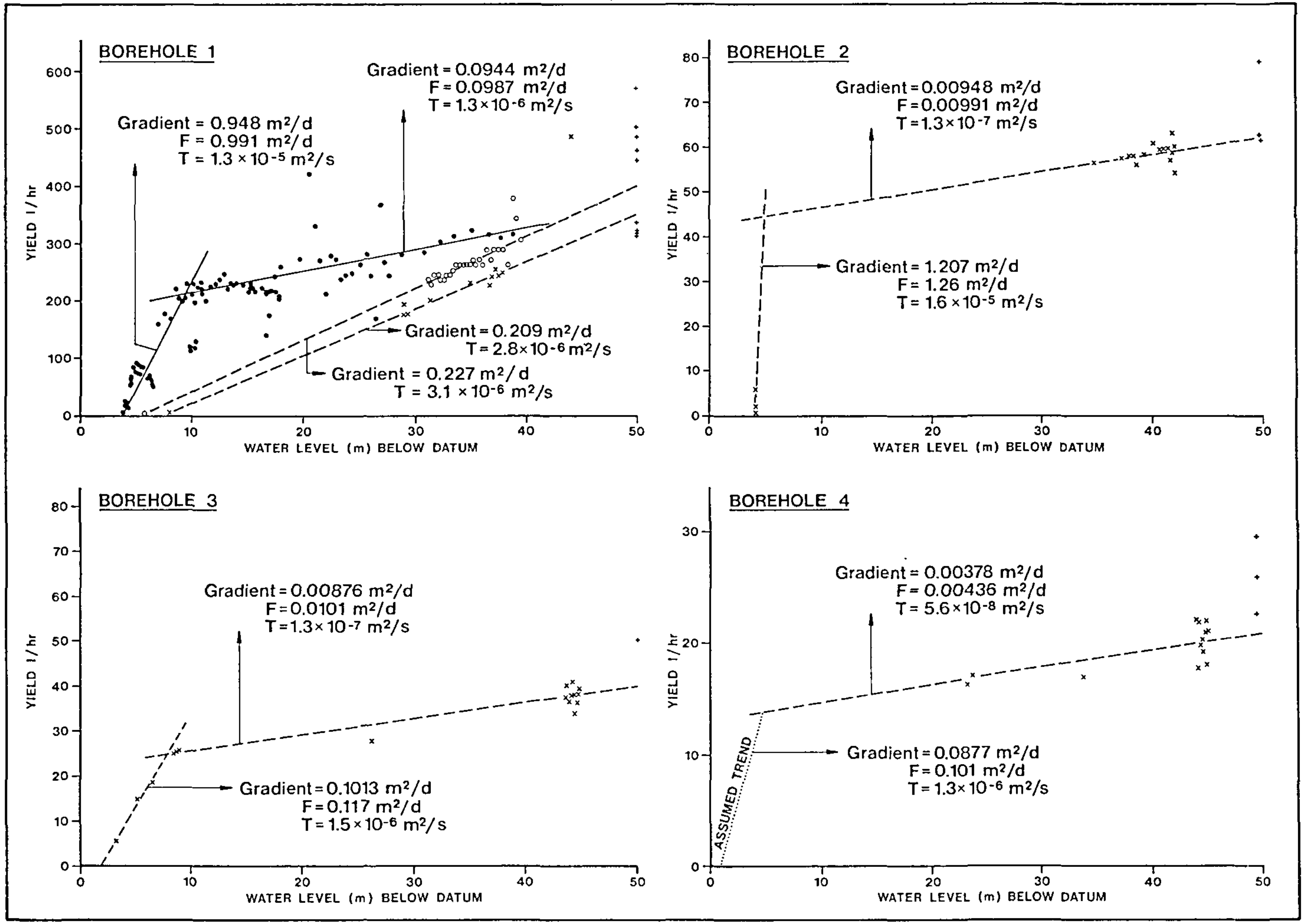



Fig. 9. Results of capacity testing carried out on testholes 1-4. See text for explanation of symbols. Plots show yield from aquifer ($Q_A$) vs. water level below well-top.

Table 3 : Hydrogeological parameters calculated from Figs. 9 a-d.

| Borehole section | | Borehole 1 0-12 m | 1 12-73½ m | 2 0-12 m | 2 12-73½ m | 3 0-12 m | 3 12-73 m | 4 0-12 m | 4 12-73 m |
|---|---|---|---|---|---|---|---|---|---|
| Gradient G | $m^3/d$ | 0.948+ | 0.218 | 1.21 | 0.00948 | 0.1013 | 0.00876 | 0.0877* | 0.00378 |
| Specific capacity F | $m^3/d$ | 0.991+ | 0.228 | 1.26 | 0.00991 | 0.117 | 0.0101 | 0.101 | 0.00436 |
| **Entire borehole** | | | | | | | | | |
| Transmissivity T | $m^2/d$ | 1.10+ | | 1.40 | | 0.130 | | 0.113 | |
| | $m^2/s$ | $1.3 \times 10^{-5}$ | | $1.6 \times 10^{-5}$ | | $1.5 \times 10^{-6}$ | | $1.3 \times 10^{-6}$ | |
| **Section-wise** | | | | | | | | | |
| Transmissivity T | $m^2/d$ | 0.848+ | 0.253 | 1.39 | 0.0110 | 0.119 | 0.0112 | 0.108 | 0.00485 |
| | $m^2/s$ | $9.8 \times 10^{-6}$ | $2.9 \times 10^{-6}$ | $1.6 \times 10^{-5}$ | $1.3 \times 10^{-7}$ | $1.4 \times 10^{-6}$ | $1.3 \times 10^{-7}$ | $1.2 \times 10^{-6}$ | $5.6 \times 10^{-8}$ |
| Saturated depth | m | 8.25 | 69.75 | 9.24 | 61.5 | 10.83 | 61 | 11.08 | 61 |
| Hydraulic conductivity K | m/s | $1.2 \times 10^{-6}$ | $4.2 \times 10^{-8}$ | $1.7 \times 10^{-6}$ | $2.1 \times 10^{-9}$ | $1.3 \times 10^{-7}$ | $2.1 \times 10^{-9}$ | $1.1 \times 10^{-7}$ | $9.2 \times 10^{-10}$ |

+ shallow fracture appears to have dried up during testing
* based on assumed gradient, Fig. 9d.

son and Carlstedt (1977) developed the method further for non-steady-state pumping and contend that, for normal Swedish 110 mm diameter holes, with pumping periods under 1 day, the c value will lie between 0.9 and 1.1 (they then proceed to use values of around 0.84 to evaluate apparent T for four Swedish bedrock areas!). A value of 0.9 is used here (Banks 1991), and the calculated apparent transmissivities are shown in Table 3.

It should be noted from Fig.9a that the drawdown curve indicates the presence of a shallow contributing fracture in borehole 1 at c. 9-11 m (presumably that encountered during drilling at c.11½ m - see Fig.8). This appears to have dried up during pumping, the recovery curve being a straight line, dominated entirely by the major deep inflow at 62 m. In boreholes 2, 3 and 4 the yield is very low, and what little water there is is derived from shallow, relatively transmissive fractures at around 5, 8 and (presumably) 4.75 m respectively (corresponding well with drilling logs). Apparent permeability in the boreholes can be calculated by dividing the transmissivity by the boreholes' saturated length. Hole 1 gives an average apparent permeability of 4 x $10^{-8}$ m/s, relatively low for a borehole crossing a prominent fracture zone. In the other boreholes apparent permeabilities in the region of $10^{-9}$ m/s are obtained for the borehole sections below the transmissive near-surface fractures, agreeing well with the results derived from Lugeon testing in the lower permeability sections of the Hvaler tunnel (Banks et al. 1992). The average permeability of Precambrian gneissic and granitic aquifers is reported to lie around $10^{-6}$ - $10^{-8}$ m/s, with many 'water yielding' fracture zones having permeabilities of $10^{-6}$ m/s and higher, and more massive portions of bedrock $10^{-9}$ m/s or less (Hult et al. 1978, Olsson 1985). The Iddefjord granite at the test-site thus lies towards the lower end of this scale.

In defence of the «prominent fracture zones equals elevated borehole yields» theory, the one borehole (no.1) which clearly crossed a significant fracture zone did have a substantially higher yield (c. 360 l/hr with pumping water level (PWL) at 50 m) than the other boreholes 2, 3, 4 (65, 40 and 22 l/hr respectively). It must, however, be pointed out that the yield of 360 l/hr is still below the median (400 l/hr) and mean (745 l/hr) yields for the Iddefjord granite in Hvaler. This could be due to:

(a) the fracture zone having a low-permeability filling of clay minerals.

(b) the statistics from the borehole archive being artificially exaggerated.

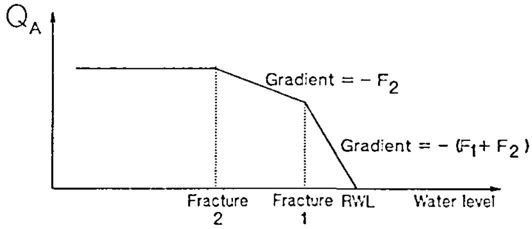



Fig. 10. Diagram illustrating theory behind capacity-test analysis for a vertical borehole. $F_1$ = specific capacity of fracture 1 etc. RWL = rest water level.

Table 4 : Summary of hydrochemical samples. Sample numbers correspond with figs. 11-14. Samples 1-9 analysed in connection with tunnel construction. Samples 10-35 & 6b analysed by NGU (d = bore deviated from vertical).

| Sample | Name | Sampling date | Source Type | Borehole depth m | Chloride mg/l | Water Type |
|---|---|---|---|---|---|---|
| 1 | Birkeland | 23/11/87 | Borehole - granite | 25 | 11.2 | III |
| 2 | Berg | 23/11/87 | Borehole - granite | > 50 | 4.2 | I |
| 3 | Hansen | 23/11/87 | Borehole - granite | > 31 | 20 | II |
| 4 | Sørensen | 23/11/87 | Borehole - granite | > 43 | 19 | III |
| 5 | Marstrander | 23/11/87 | Borehole - granite | 78 | 270 | IV |
| 6a | Bombua | 20/10/89 | Borehole - granite | ? | 37 | |
| 6b | Bombua | 23/03/90 | Borehole - granite | ? | 22 | III (II) |
| 7a | Sørensen 2 | 25/10/88 | Borehole - granite | 60 | 33 | II or III |
| 7b | Sørensen 2 | 15/11/88 | Borehole - granite | 60 | 35 | II or III |
| 7c | Sørensen 2 | 24/02/89 | Borehole - granite | 60 | 37 | II or III |
| 7d | Sørensen 2 | 27/06/89 | Borehole - granite | 60 | 938 | IV |
| 8 | Sørensen 3 | 28/04/89 | Borehole - granite | 70 d | 800 | IV |
| 9 | Chain.2140 | 28/04/89 | Leakage in tunnel | | 56 | |
| 10a | Chain.4120 | 23/03/90 | Leakage in tunnel | | 11600 | Modified sea |
| 10b | Chain.4120 | 29/05/91 | Leakage in tunnel | | 17300 | Modified sea |
| 11 | Urdal | 09/05/90 | Leakage from granite cliff | | 21 | II |
| 12 | Melhuus | 15/05/90 | Well - Quaternary | 3 | 22 | II |
| 13 | Melhuus | 15/05/90 | Borehole - granite | 45 | 57 | IV |
| 14 | Bølingshavn | 15/05/90 | Well - granite | 6 | 96 | IV |
| 15 | Bølingshavn | 15/05/90 | Borehole - granite | 27 | 202 | IV |
| 16 | Dauløkkene N | 15/05/90 | Borehole - granite | ? | 20 | III |
| 17 | Dauløkkene S | 15/05/90 | Borehole - granite | ? | 56 | IV |
| 18 | Sandbrekke | 16/05/90 | Borehole - granite | 80 | 35 | III |
| 19 | Testsite 1 | 08/11/90 | Well - Quaternary | 1.7 | 35 | II |
| 20 | Bølingshavn | 08/11/90 | Seawater | | 7500 | Seawater |
| 21a | Testhole 1 | 25/05/91 | Borehole - granite | 73.5 d | 32 | III* |
| 21b | Testhole 1 | 25/05/91 | Borehole - granite | 73.5 d | 57 | IV* |
| 22 | Knausen | 26/05/91 | Borehole - granite | 90 | 283 | IV |
| 23 | Skartlien | 26/05/91 | Well - Quaternary | 3.8 | 31 | II |
| 24 | Testhole 2 | 27/05/91 | Borehole - granite | 73.5 d | 17 | III* |
| 25 | Granli | 28/05/91 | Borehole - granite | 101 | 98 | IV |
| 26 | Testhole 3 | 28/05/91 | Borehole - granite | 73 d | 32 | III* |
| 27 | Chain.3615 | 29/05/91 | Leakage in tunnel | | 17000 | Modified sea |
| 28 | Testhole 4 | 29/05/91 | Borehole - granite | 73 d | 103 | IV* |
| 29 | Solhell | 30/05/91 | Well - Quaternary | ? | 15 | II |
| 30 | Andresen | 30/05/91 | Borehole - granite | 40 | 26 | III |
| 31 | Heyerdahl | 30/05/91 | Borehole - granite | 65 | 468 | IV |
| 32 | Granlie | 30/05/91 | Borehole - granite | 70 d | 92 | IV* |
| 33 | Dahl | 30/05/91 | Borehole - granite | 76 d | 87 | IV |
| 34 | Testsite 1 | 31/09/91 | Precipitation | | 3.0 | |
| 35 | Testsite 1 | 31/09/91 | Storm run-off from granite | | 42 | |

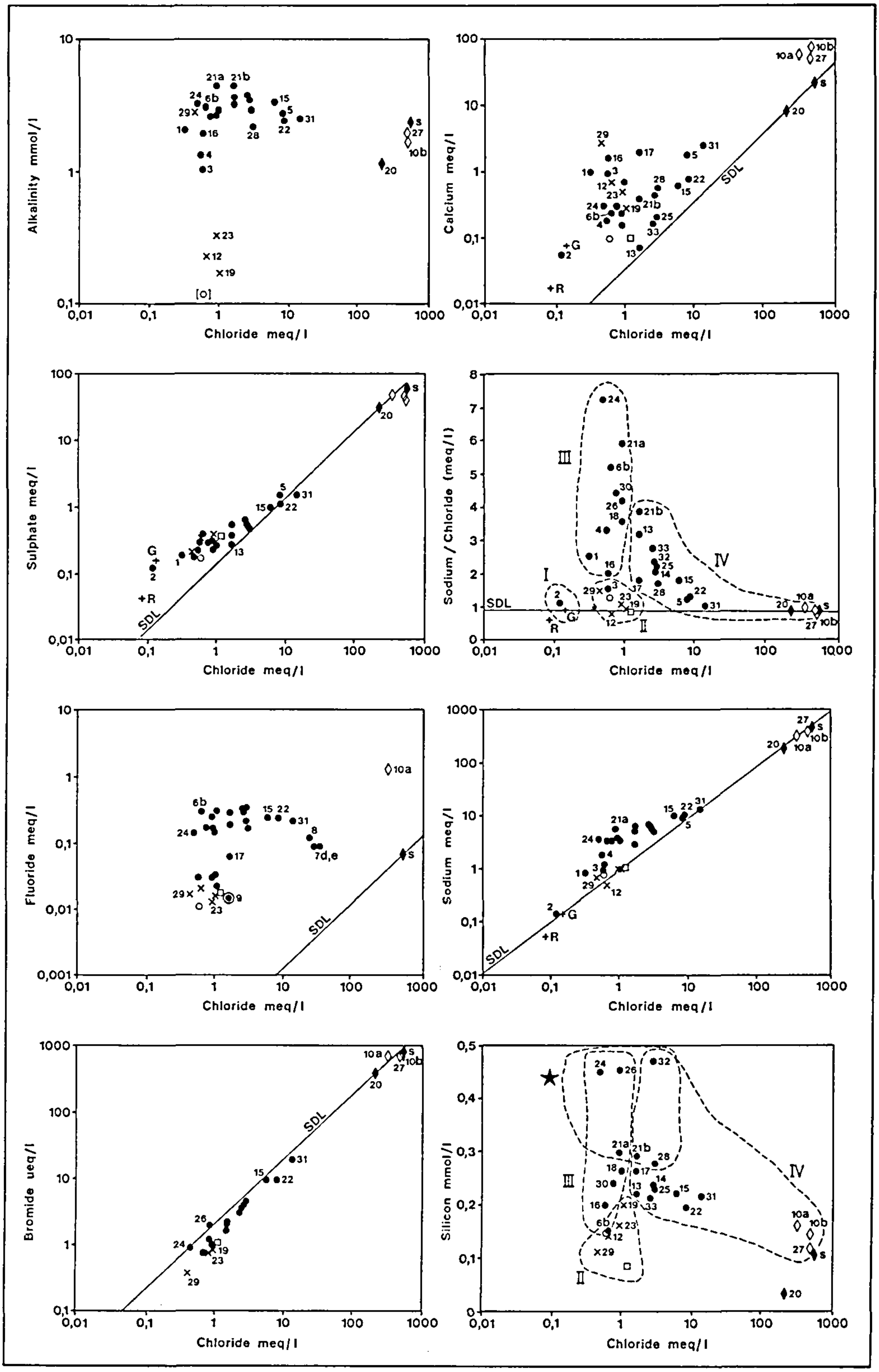
Drilling cuttings from the boreholes were analysed by X-ray diffraction. In one sample, from a 'dry' fracture in borehole 3 at $10\frac{1}{2}$ m, where it was possible to extract individual clots of clay minerals from the cuttings, smectites were positively identified.

The test pumping results give an overall picture of a 'two-layered' fractured aquifer. The upper layer, down to approximately 12 m, contains relatively transmissive fractures ($10^{-6}$ - $10^{-5}$ $m^2/s$), which do not, however, yield large quantities of water due to the small head gradients that are achievable along such shallow fractures. The deeper layer is typically of very low permeability, c. $10^{-9}$ m/s, except where a major fracture zone crosses borehole 1, having a significantly, though not dramatically, elevated apparent transmissivity of $3 \times 10^{-6}$ $m^2/s$ (corresponding to an average apparent permeability for the lower section of the borehole of $4 \times 10^{-8}$ m/s).

# Groundwater chemistry

A series of water samples has been taken from boreholes in the Iddefjord Granite, both in connection with the excavation of the Hvaler tunnel, and with NGU's project. In addition, samples of seawater (Bølingshavn), rainwater (in an open area near testhole 3 during a storm), and storm run-off (from a granitic 'massif' near testhole 3) have been taken, together with samples from water leakages in the Hvaler tunnel and from wells in Quaternary deposits. The samples are detailed in Table 4.

The NGU samples have been analysed for cations (samples filtered by a 0.45 μm Millipore filter) using inductively coupled emission spectroscopy, anions by ion chromatography, alkalinity by standard titration, and for pH and electrical conductivity in the laboratory using standard electrodes. The relationships between the most significant parameters are detailed in Figs. 11, 12 & 13. Four main water ty-


Alkalinity mmol/l
Calcium meq/l
Sulphate meq/l
Sodium / Chloride (meq/l)
Fluoride meq/l
Sodium meq/l
Bromide ueq/l
Silicon mmol/l
Chloride meq/l
SDL

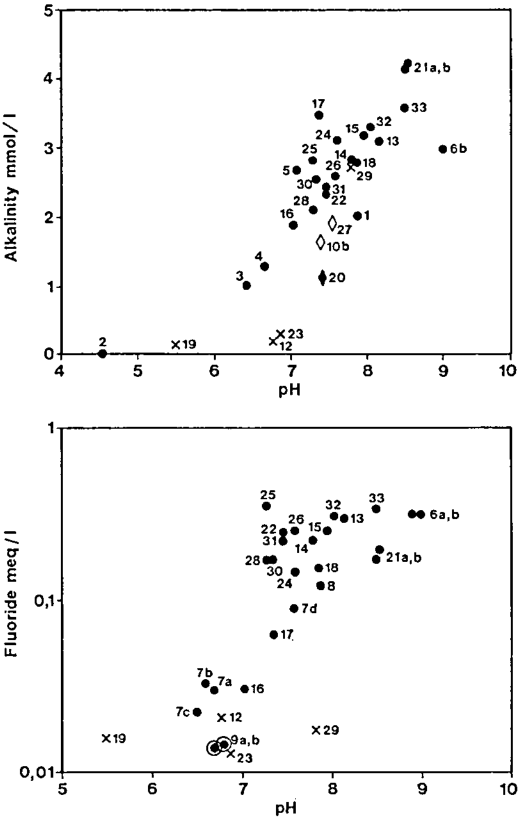



Fig. 12. Fluoride and alkalinity plotted against pH. Symbols as for Fig. 11.

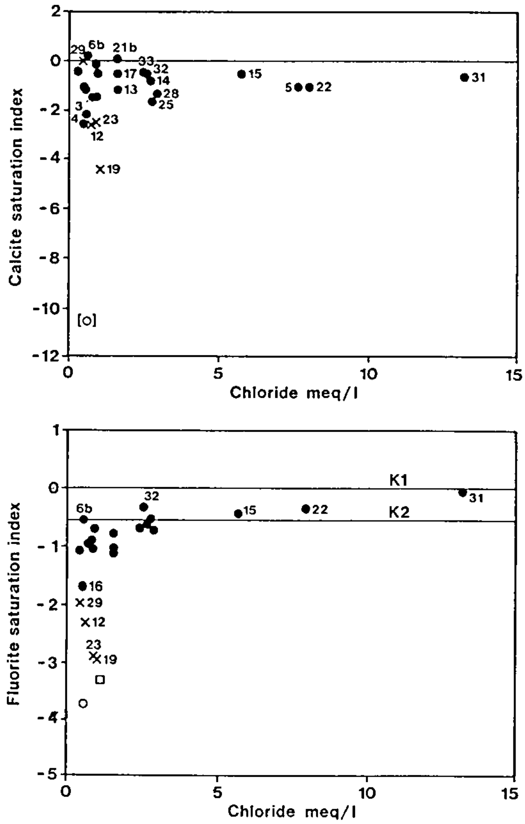



Fig. 13. Estimated calcite and fluorite saturation indices (SI) plotted against chloride. Calcite index estimated from the simplified equilibrium : $K=(Ca^{++})(HCO_3^-)/(H^+)$; log K = 2.17 at 7°C (Lloyd & Heathcote 1985). For fluorite saturation index, $K_1$ shows SI relative to log $K_S$ = -10.4 (Krauskopf 1977); $K_2$ shows SI relative to Nordstrom & Jenne's (1977) value of -10.96 (both at 25°C). K values are temperature-adjusted to 7°C using the Van't Hoff isotherm, and activities are calculated using the Debye-Huckel formula. Speciation modelling of water has not been included in calculations. Symbols as for Fig. 11.

pes, and one subsidiary type, can be identified on the basis of sodium, chloride and silicon content, as follows.

Type I has sodium and chloride concentrations very similar to those in rainwater, and is only recorded in one borehole (number 2). This indicates little interaction in the soil zone or rock, and probably implies very rapid recharge/pipe flow. The possibility of direct surface run-off into a badly constucted borehole cannot be excluded. Unfortunately, sample collection was not, in this instance, performed by NGU, and the borehole is no longer available for examination.

Type II waters show significantly (typically 4 to 10 times, agreeing well with Jacks 1973) higher sodium and chloride concentrations

←

Fig. 11. Various hydrochemical constituents plotted against chloride concentration. Numbers refer to well numbers in Fig.14 and Table 4.

● = borehole in bedrock
x = dug well in drift
♦ = seawater sample
□ = storm run-off (no.35)
SDL = seawater dilution line
○ = fracture at Urdal (nr.11)
⊙ = freshwater leakage in tunnel (no.9)
◇ = saltwater leakage in tunnel
+ R = rainfall sample (no.34)
★ = groups III*, IV*

[o] = alkalinity calculated from ion-balance (unreliable)
♦S= average seawater after Horne (1969) & Lloyd and Heathcote (1985)
+G = precipitation at Gøteborg (further south along Skaggerak coast) after Jacks (1973)

than precipitation, although similar Na/Cl ratios. This is indicative of evapotranspirative up-concentration on vegetation and in the soil zone (together with dry fallout of salt), but relatively little interaction with geological materials (low Si concentrations). Such waters are typical of wells in drift deposits. Few bedrock waters fall in this group.

Type III waters include the majority of waters from bedrock boreholes. They show elevated Na/Cl ratios, elevated Si levels but similar Cl concentrations with respect to type II waters. This is indicative of mineral-water interaction (leading to elevated Si and Na), but with very little salt-water mixing. These tend to be sodium (or occasionally calcium) bicarbonate dominated waters.

Type IV waters reflect some degree of mixing between Type III water and saline water (elevated Cl concentrations relative to type III, but reduced Si concentrations, due to seawater's relatively low Si content).

The four water types represent an attempt to systematise the groundwater chemistry. In reality there are no sharp boundaries between the water types, and the progression from groups I to IV represents the evolution of waters with an increasing degree of water-rock interaction (probably reflecting retention time in the aquifer) and mixing with salt water.

Type III* and IV* waters form subsets of Types III and IV. They have elevated concentrations of most cationic species and very high contents of silicon, aluminium, iron and titanium. These samples come from newly drilled boreholes (Testholes 1 to 4, drilled Nov.1990 and sampled during test-pumping May 1991; and Granlie's new hole, sampled in May 1991 the day after drilling completion). It is believed that interaction between the water and the newly exposed drilling cuttings, with high specific surface area, has led to these elevated concentrations, despite filtering of the samples. The anions appear relatively unaffected.

The origin of saline waters in bedrock aquifers is a subject much debated. While seawater intrusion is the obvious source for an island situation such as Hvaler, it must be noted that relatively saline waters occur in the centre of Kirkeøy (Fig.14), in boreholes which, at least according to the Ghyben-Herzberg relation, are not deep enough to approach the base of the presumed freshwater lens. Nordstrom et al. (1989) propose that the deep saline waters in the Stripa mine, also in Precambrian granite, may be derived from fluid inclusion leakage. They also note, however, that such 'exotic' saline groundwaters tend to have elevated bromide concentrations relative to sea water. The saline waters from Hvaler do not exhibit this, but lie approximately on, or a little below, the seawater dilution line, with respect to bromide. Edmunds & Savage (1991) believe that chloride can also be derived from the weathering of some minerals such as biotite. They also note, however, that in such cases, there is usually a surplus of $Cl^-$ ions over $Na^+$; the opposite to the situation observed on Hvaler. Less exotic derivations for the salinity of Type IV water must therefore be accepted, and can include: intrusion of present-day 'seawater' (the water surrounding Kirkeøy is in fact significantly less saline than ocean water, the island lying just south of the estuary of the Glåma, Norway's largest river); pumping-induced upconing of salt water from below the freshwater lens; residual pre-emergence seawater in 'blind' or low-permeablity fracture systems; downward leaching of residual saline pore water from the island's marine Quaternary deposits (this last explanation is unlikely, shallow wells in Quaternary deposits exhibiting no increased salinity).

The dug wells in Quaternary deposits (except for Solhell) typically yield relatively low pH, low alkalinity water. This indicates short residence time and/or little buffer capacity in the Quaternary aquifers against rather acidic recharge waters. Although recharge waters are often acidic for other reasons than acidity in precipitation (pH of rainfall in SE Norway = c. 4.3, Henriksen et al. 1987), the effects of contaminated 'acid' rain are revealed in groundwater sulphate concentrations. Most of the waters lie approximately on the seawater - precipitation mixing line, indicating non-geological sources for the majority of the sulphate. The rainfall, run-off and lower water types lie above the seawater dilution line, however, due to precipitation being contaminated by anthropogenic sulphur oxides. There is reported an average of c.4 mg/l sulphate in the Hvaler area's precipitation (Soveri 1982). This value lies between the single point precipitation sample from Hvaler (sample 34) and the reported values for Gøteborg (Jacks 1973) further south along the Skaggerak coast. The smell of hydrogen sulphide has been detected in at least one of the boreholes, number 18, indicating some degree of sulphate reduction

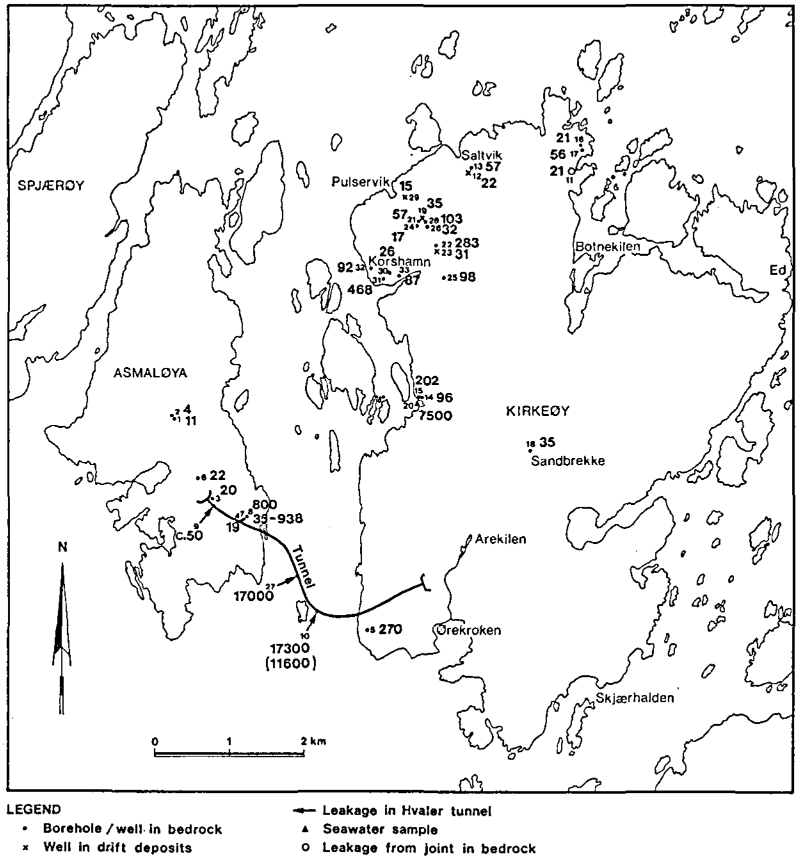



Fig. 14. Map of Kirkeøy and Asmaløy showing location of sampled groundwaters. Small figures show sample number. Large figures show Cl-concentration in mg/l.

in the aquifer.

The precipitation also contains a significant nitrate content (4.0 mg $NO_3^-$/l from the point sample no.34, c. 2.6 mg $NO_3^-$/l on average according to Storrø 1990). Very few groundwater samples display concentrations above 0.05 mg/l, and the surface run-off (sample 35) contains only 0.22 mg/l. This indicates nitrate uptake or denitrification in the soil zone.

A plot of alkalinity vs. pH (Fig.12) reveals a typical titration-type curve. Calcium and bicarbonate concentrations are also controlled by calcite saturation (Fig.13), many of the Type III and IV waters apparently being saturated with respect to calcite.

The granitic waters contain very high concentrations (up to 6 mg/l, compared with the SIFF (1987) drinking water limit of 1.5 mg/l) of fluoride. Possible sources for this fluoride include fluorite (observed on joint surfaces), sheet silicates (micas, chlorite), amphiboles and apatite. Assessing the importance of fluorite precipitation as a control on fluoride concentrations is complicated by the uncertainty surrounding fluorite's solubility product. Estimates of log $K_s$ at 25°C range from -8.27 to -11.19 (Nordstrom & Jenne 1977). Krauskopf (1979) gives a value of -10.4, while Nordstrom & Jenne (1977) conclude a value of -10.96. Fig.13 shows saturations calculated from these values (corrected to 7°C using the Van't Hoff isotherm, and a $\Delta H_s$ of 4.7 Kcal/mole - Krauskopf 1979). It appears that the type IV waters, at least, appear to approach saturation with respect to fluorite, though many of the waters in lower water types do not. In the latter case it appears that pH is the controlling factor, fluoride concentration increasing with elevated pH (Fig. 12). Such a trend is noted by Englund & Myhrstad (1980), and is explained by $OH^-$ versus $F^-$ ion exchange on illites, micas, chlorites and amphiboles at high pH values.

## Conclusions

Mapping of fractures and fracture zones in the Iddefjord granite by means of topographical maps, aerial photos and field measurements has revealed an increasing degree of complexity with decreasing scale. Due to the probable reactivation of the fracture pattern many times during the granite's history, it has not been possible, hitherto, to produce a simple tectonic interpretation of the pattern, or its significance for the granite's hydrogeology.

Magnetic and VLF measurements have proved the most powerful tools for identifying fracture zones in the granite. Observations in the Hvaler subsea road tunnel and in connection with test-pumping of NGU's boreholes on land confirm that the presence of a geophysical anomaly, or a fracture zone, does not guarantee a highly increased transmissivity in the zone. Possible reasons for this include the tightening of fracture zones by secondary clay minerals, or the non-persistence, or 'closure', of some fracture zones with depth.

Statistical analysis of 310 boreholes in the Iddefjord granite reveals a distribution of yield strongly skewed towards low yields. A median yield of 500 l/hr contrasts with a mean of 1124 l/hr. The median, rather than the mean, yield should be used when assessing the probable yield of a planned borehole in a bedrock aquifer.

Test-pumping results indicate a background permeability in the granite at NGU's testsite of $10^{-9}$ m/s. In the top c.12 m of the granite, however, transmissive fractures are more frequent, leading to apparent permeabilities of $10^{-6}$ - $10^{-7}$ m/s. Where one of the test-boreholes crossed a prominent crush-zone, a significant, though not spectacular, water-inflow was noted at the base of the zone, and a transmissivity of $3 \times 10^{-6}$ m$^2$d was calculated.

The groundwater chemistry on Hvaler is controlled by mixing between up-concentrated precipitation and seawater end-members, coupled with water-rock interactions. Bicarbonate buffering, anion exchange and calcite/fluorite saturation appear to be important processes controlling pH, bicarbonate, fluoride and calcium concentrations. No evidence of cation exchange has, as yet, come to light, the effect of seawater mixing overshadowing any traces of such a process.

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