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Drilled well yield and hydraulic properties in the Precambrian crystalline bedrock of Central Finland

by

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ABSTRACT

The drilled well yield and hydraulic properties and their relationships to different well factors related to the location of the wells were investigated in the Precambrian crystalline bedrock of Central Finland. Data from 2,352 private wells constituted the primary study material. Additional data from 73 test wells were utilized as a support material. Based on technical well data and single well pumping tests, estimates for hydraulic parameters (normalized yield, specific capacity, well productivity, transmissivity, bulk hydraulic conductivity) were statistically determined. Nearly 60 well factors were extracted. They were divided into five groups: construction, geologic, topographic, lineament and catchment factors. In addition, the role of seismotectonics was considered.

In Central Finland, drilled wells are most often used in domestic water supply by single households and farms. Some 30 villages and small towns use bedrock groundwater for their common water supply. Bedrock groundwater can provide a valuable source for water supply of large communities in various times of crisis. Hydraulic fracturing has proven to be a successful method in increasing the yield of low to medium yield wells. The median well depth and yield are 73 m and 700 Lhr^{-1}, respectively. The median hydraulic parameters are as follows: normalized yield Q/d_s 12 $Lhr^{-1}m^{-1}$, specific capacity Q/s 50 $Lhr^{-1}m^{-1}$, well productivity Q_w $2,1x10^{-7}$ ms^{-1}, transmissivity T $7,3x10^{-6}$ m^2s^{-1}, and hydraulic conductivity K $1,1x10^{-7}$ ms^{-1}. Much the same values have been reported in most unweathered crystalline rock settings around the world despite different lithologies, climate and tectonic histories. This is suspected to be largely due to similar rock fracturing characteristics. Q/d_s, Q/s and Q_w can be used as estimates for bulk T and K in regional studies of fractured rock aquifers. A prerequisite for this is, however, that their mutual relations have first been statistically adjusted.

The well yield and hydraulic conductivity decrease downwards in bedrock, at least to the drilled well depths. The soil type at well site and the thickness of overburden are not of any great importance to the well production properties. Lithological differences between well sites may be considered insignificant from the well production point of view. The drilled well yield and hydraulic properties are not statistically related to surface water bodies. Neither does the land uplift rate nor does the highest shore level indicate any clear trend in the well production properties. The productivity of a drilled well is clearly related to its topographic setting. Valley wells are most productive whereas hilltop wells, though deepest, yield the least amounts of water. The distance to the nearest (bedrock) hilltop and the relative height differences in a well's catchment area are statistically related to the well yield and hydraulic properties. The proximity of lineaments is considered the most important factor entity controlling the productivity of drilled wells in the study area. The most significant single lineament factors are the azimuth and prominence of lineaments and their perpendicular distance to the drilled wells. Lineament intersections have no statistical relations to the well production properties.

Tectonic reactivation of faults is considered of utmost importance for groundwater flow and drilled well hydraulics. Analogously to the large postglacial faults in northern Fennoscandia, the NE-SW and NW-SE lineaments have most probably been activated in the study area during the last phases of the Weichselian deglaciation some 10,000 years BP. This well explains their higher permeability compared to other lineament sets and may be considered as an implication of postglacial faulting in Central Finland. Elevated well yields elsewhere in glaciated terrains of Fennoscandia and northern North America are possibly related to similar fault reactivation during the deglacial time. Although much has been done, a better understanding of bedrock hydrogeology and paleohydrogeology is still needed in the arrangements of the future disposal of nuclear fuel waste in Fennoscandia.

High-yield well siting in crystalline rock areas should start with lineament mapping. The key point in obtaining the highest probability of success is to be able to identify those lineament sets, which are optimally orientated and critically stressed in a present-day stress regime and/or have been geologically recently activated. Hence, in a thrust fault regime, one should go for detecting lineaments that strike perpendicular to the maximum horizontal stress direction S_H. In normal fault regimes the most promising lineaments should lie parallel to the S_H, while in strike-slip regimes they either coincide with the S_H or diverge at various angles ($<<$ 45°) to it. Along with the lineament orientation, the prominence of lineaments must also be considered. In Central Finland, the short NW-SE and medium length to long NE-SW lineaments should be preferred to other lineament sets. Their central parts ($\leq$ 100 m) are optimal for high-yield well drilling. Staying in low-elevation areas away from hilltops and ridges statistically increases the chances to catch a high-yield well site. The siting of high-yield wells may also benefit from previous borehole information or water well inventories to be executed in the area of interest.

A conceptual hydrotectonic model was developed to guide the selection of the most favorable well sites in the study area. The lineament-stress analysis forms the fundamental basis of the model. In addition, the optimal field for bedrock seismic velocity have been included in the model. It is stressed that only the drilling and proper pumping test with water quality determinations provide the ultimate information required to decide whether a high-yield well site would suit for long-term withdrawal of groundwater.

Keywords: groundwater, crystalline bedrock, drilled wells, well yield, hydraulic properties, normalized yield, specific capacity, well productivity, transmissivity, hydraulic conductivity, lineaments, seismotectonics, rock stress, fault reactivation, postglacial faulting, high-yield well siting, hydrotectonic model, Precambrian, Central Finland.

CONTENTS

LIST OF SYMBOLS

A	m^2	area
A_f	m^2	cross-sectional area of flow
α	--	level of risk
B	sm^{-2}	aquifer constant
b	m	thickness of aquifer
C	s^2m^{-5}	well-loss constant
C_s	MPa	cohesive strength of a medium
C_p	--	Mallow´s statistic
CQ^2	%	well loss
°C	°	degree Celcius
d	m	drilled rock depth
d_s	m	saturated open well section
F	N	force
g	ms^{-2}	acceleration due to gravity
Δh	--	hydraulic gradient
θ	°	inclination angle to σ_1
K	ms^{-1}, md^{-1}	hydraulic conductivity
κ	m^2	permeability
λ_v	--	pore-fluid factor
max	--	maximum
Md	--	median
min	--	minimum
n	--	number of observations
n_e	--	effective porosity
p	--	level of significance
P_f	MPa	fluid pressure
Q	Lhr^{-1}	yield
Q/d	$Lhr^{-1}m^{-1}$	normalized yield (yield per drilled rock depth)
Q/d_s	$Lhr^{-1}m^{-1}$	normalized yield (yield per saturated open well section)
Q_f	m^2s^{-1}, m^2d^{-1}	normalized well productivity ('yield factor')
Q/s	$Lhr^{-1}m^{-1}$, m^2s^{-1}	specific capacity
Q/s_c	$Lhr^{-1}m^{-1}$, m^2s^{-1}	specific capacity corrected for well loss
Q/s_m	$Lhr^{-1}m^{-1}$, m^2s^{-1}	specific capacity determined with step test
Q_w	ms^{-1}, md^{-1}	well productivity (normalized specific capacity)
Q_1	--	lower quartile
Q_3	--	upper quartile
R^2	--	coefficient of determination (the goodness of fit)
r	--	correlation coefficient
r_c	m	radius of the unscreened part of a well
r_e	m	radius of influence of a well
r_w	m	well radius
ρ	kgm^{-3}	density of fluid
Σ	--	sum
S_H	MPa	maximum horizontal stress
S_h	MPa	minimum horizontal stress
S_V	MPa	vertical stress
s	--	standard deviation
s_w, s_m, s	m	drawdown
Δs_w	m	drawdown difference per log cycle of time

σ	MPa	magnitude of stress
σ_m	MPa	mean stress
σ_n	MPa	normal stress
σ_n'	MPa	effective normal stress
σ_1	MPa	maximum principal stress
σ_2	MPa	intermediate principal stress
σ_3	MPa	minimum principal stress
T	m^2s^{-1}, m^2d^{-1}	transmissivity
T_m	m^2s^{-1}, m^2d^{-1}	transmissivity determined with pumping test
T_n	ms^{-1}, md^{-1}	normalized transmissivity
T_s	MPa	tensile strength
T_t	--	slip tendency
t	min, day	time
τ_f	MPa	frictional strength
τ_{max}	MPa	maximum shear stress
τ_s	MPa	shear stress
μ	$kgm^{-1}s^{-1}$	dynamic viscosity
μ_s	--	static coefficient of friction
v	ms^{-1}	average linear flow velocity
W	--	Shapiro-Wilk statistic
$\bar{x}$	--	arithmetic mean
φ	°	slope of failure envelope
z_i	--	standard score

CONVERSION FACTORS

For readers who wish to convert measurements from the metric system of units to the inch system of units, the conversion factors are listed below:

Multiply	By	To obtain
meters (m)	3,281	feet (ft)
square meters (m^2)	10,76	square feet (ft^2)
kilometers (km)	0,6214	miles (mi)
square kilometers (km^2)	0,3861	square miles (mi^2)
liters per hour (Lhr^{-1})	$4{,}403 \times 10^{-3}$	gallons per minute ($galmin^{-1}$)
liters per hour per meter ($Lhr^{-1}m^{-1}$)	0,0144	gallons per minute per foot ($galmin^{-1}ft^{-1}$)
cubic meters per day (m^3d^{-1})	0,1835	gallons per minute ($galmin^{-1}$)

1 INTRODUCTION

1.1 General

Groundwater is a crucial source of fresh water throughout the world. Groundwater is an essential part of the hydrologic cycle and is important in sustaining streams, lakes, wetlands, and aquatic communities (Davis & DeWiest 1966, Freeze & Cherry 1979, Alley et al. 2002).

More than 1,5 billion people worldwide rely on groundwater for their primary source of drinking water. In places, groundwater is extensively used for agricultural irrigation, too. In India, for instance, more than 90% of rural and nearly 30% of urban population depend on groundwater for meeting their drinking and domestic requirements and more than 50% of groundwater is used for irrigation purposes (Gustafson & Krásný 1993, Foster et al. 2000). Approximately 700 billion liters of groundwater is extracted worldwide every year (Banks et al. 1994, Foster & Chilton 2003). Yet another 1,5 billion people do not have access to a safe water supply. In many countries in the developing world, access to a reliable supply of high-quality groundwater can mean the difference between life and death (Banks et al. 1996).

Certain factors make groundwater an attractive water supply source compared with surface water. Firstly, there is groundwater in shallow or moderate deep aquifers in most places around the year; secondly, groundwater is for most part protected naturally from evaporation and pollution (although persistence of pollution when it happens may be long-lasting); and thirdly, groundwater has frequently a good and stable microbiological and physico-chemical quality and requires minimal or no treatment at all (e.g. Davis & DeWiest 1966, Freeze & Cherry 1979, Foster et al. 2000, 2006, Gyau-Boakye et al. 2008).

Groundwater in fractured rocks has gained great international importance in recent years. There are several reasons for this. One main reason is the increased demand of safe drinking water coupled with periodic droughts as the world´s population increases, as well as the increased demand of water because of the improved standard of living (e.g. Wladis 1995, Krásný 2003, Faillace 2007). There is also an increased need for water in industry and agriculture. Other reasons are engineering projects such as road and tunnel constructions, foundation of houses, geothermal reservoirs and disposal of waste, especially radioactive and high-level nuclear fuel waste (e.g. Lyslo 2000, Limaye 2004, Knutsson 2008, Petrov et al. 2008, Vidstrand et al. 2008). Faster and more efficient methods of water-well drilling, like down-the-hole hammer (DTH), have also greatly promoted the construction of wells in crystalline rocks (Banks & Less 1999, MacDonald et al. 2002, Faillace 2003, 2007, Rebouças 2004).

The importance of crystalline rock aquifers for water supply issues differs from place to place, depending on various factors but mainly upon the overall availability of water and water demand (Gustafson & Krásný 1994). In northern Europe and North America, the native consolidated bedrock is mantled with a variable thickness of unconsolidated glacial deposits whose aquifers yield most of the groundwater used in the populated centers of these regions (Davis & DeWiest 1966, Zektser & Everett 2004). However, rural inhabitants largely abstract groundwater from boreholes in crystalline bedrock (Knutsson & Fagerlind 1977, Englund et al. 1988, Banks et al. 1996). Some 10 percent of the Finnish and Swedish population, for instance, uses their own private wells for domestic water supply; a major part of these wells are drilled in Precambrian bedrock (Gustafson 2002, Hatva et al. 2008). In Pennsylvania, USA, fractured bedrock aquifers with 3,000 public-supply wells provide the major source of water for nearly two million people (Risser & Barton 1995).

Large parts of other continents must rely still more heavily on water supplies from fractured-rock aquifers. In Brazil, for instance, more than 250,000 deep wells have been drilled during the last decades for municipalities and industry. In almost all the largest metropolitan areas (population from 1 to 16 million inhabitants), thousands of uncontrolled private wells provide groundwater for hotels, hospitals, residential buildings, and industries (Rebouças 1978, 1988). In most of the arid or semi-arid shield areas in Africa, India, and Australia, surface-water systems are meager, and groundwater is an important source for livestock and domestic use and irrigation (Habermehl 1985, Foster et al. 2000, Murty & Raghavan 2002, Giordano 2006). In these regions, a great emphasis has been given to crystalline rock terrains as drinking water sources for rural populations (Dijon 1977, Foster 1984, 2012, Chilton & Foster 1995, Faillace 2007).

While geology is an important factor in groundwater distribution, current and past climatic conditions also play a significant role. In terms of past climatic conditions, a relatively unique feature of arid regions is the large volume of non-renewable, fossil groundwater resources. These fossil groundwater supplies are most often situated deep underground, and they are an important water source for many arid countries (e.g. Davis & DeWiest 1966, Larsson et al. 1984, Lloyd 1999a, Giordano 2006).

1.2 Groundwater in crystalline rocks

Crystalline basement areas occur in many regions of the world. They are composed mainly of metamorphic and igneous (magmatic) rocks of Precambrian age and at present they form the geologically most stable parts of the continents. The seismicity of continental interiors is between two and three orders of magnitude below plate boundaries (Johnston 1989a, Crone & Machette 1994, Fenton et al. 2006).

Crystalline rocks are exposed principally in large areas called shields. The related term platform has been used to represent that part of the craton where sedimentary cover is exposed (e.g. Park & Jaroszewski 1994). The principal shields are Canadian, Guianan and Amazonian, Fennoscandian (Baltic), African, Indian, Australian, Angaran (Siberian) and Antarctic shields (Fig. 1). In addition, less extensive but important outcrops can be found outside these shield regions (e.g. Armorian Massif and Massif Central in France, Spanish Meseta and Bohemian Massif in the Czech Republic; e.g. Gustafson & Krásný 1993, 1994, Larsson & Tullborg 1993, Park & Jaroszewski 1994).

The total extent of shield-rock outcrops is estimated to be about 20% of the present land surface, i.e. 30 million km^2 (Davis & DeWiest 1966, Gustafson & Krásný 1993, 1994, Singhal & Gupta 1999, Krásný 2003). The area of the Canadian Shield is about 8 million km^2 (e.g. http://www.newworldencyclopedia.org/entry/Canadian_Shield). In Australia crystalline basement rocks occupy 3,3 million km^2 or 43% of the continent (Bestow 1990). In Sub-Saharan Africa they cover an area of 9,5 million km^2 making up 40% of the region. This means that 235 million people live in Sub-Saharan rural areas underlain by crystalline basement rocks (Foster 1984, Wright 1990a, 1992, Chilton & Foster 1995, Foster et al. 2000, MacDonald & Davies 2000, MacDonald et al. 2002, 2008, Adelana & MacDonald 2008). In India, over two third of the surface area totaling about 2,3 million km^2 is occupied by hard rock regions and nearly 50% of the replenishable resources of groundwater occur in these rocks (Singhal 2008). The Fennoscandian Shield, which is the largest Precambrian area in Europe, covers approximately one million km^2 (Gaál & Sundblad 1990).

Crystalline igneous and metamorphic rocks have practically no primary (matrix) porosity as sandstones or other sedimentary rocks do. That is, they are virtually impervious (total porosity 0,0001-0,1; Carlsson & Olsson 1977a, Gbürek et al. 1999, Lloyd 1999c, Olofsson et al. 2001, Healy & Cook 2002). Groundwater in crystalline rocks resides mainly in weathered zones of the bedrock and in fractures and other discontinuities, i.e., what is generally referred to as secondary porosity (e.g. Davis & DeWiest 1966, Larsson 1977, Freeze & Cherry 1979, Larsson et al. 1984, Greenbaum 1992, Gustafsson 1994, Lloyd 1999c, Singhal & Gupta 1999, Murty & Raghavan 2002, Tam et al. 2004, Henriksen & Braathen 2006). Groundwater in crystalline rock areas occurs under both water-table and confined conditions.

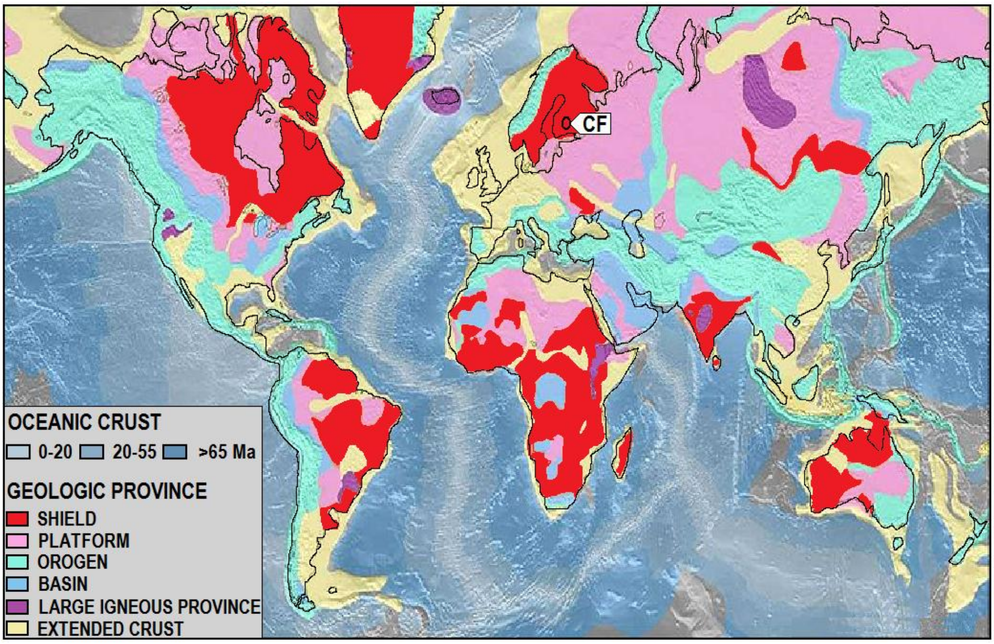

Fig. 1. World geologic provinces. CF = Central Finland. Modified from U.S. Geological Survey at http://earthquake.usgs.gov/research/structure/crust/maps.php.

Weathered crystalline bedrock aquifers are a common feature in equatorial Africa, Asia, South America and Australia (e.g. Chilton & Smith-Carrington 1984, Clark 1985, Houston & Lewis 1988, Howard et al. 1992, Taylor & Howard 2000). In the eastern districts of the Upper Region of Ghana, for example, the weathered mantle averages about 65 m thick and in places along major fault zones as much as 135 m (Taylor 1977). The basal part of a mantle of alteration products called regolith together with the deeply weathered and fractured bedrock (saprock) is likely to provide most of the yield to successful boreholes. Further, the presence of a relatively thick saturated regolith is of critical significance in terms of aquifer storage and available drawdown especially for village water-supply boreholes (Davis & DeWiest 1966, Chilton & Smith-Carington 1984, Clark 1985, Jones 1985, Acworth 1987, Waters 1989, Tennakon 1990, Barker et al. 1992, Hazell et al. 1992, McFarlane et al. 1992, Wright 1992, Chilton & Foster 1995, Edet et al. 1998, Taylor & Howard 1999, 2000, Ricolvi 1999, Dapaah-Siakwan & Gyau-Boakye 2000, Banks & Robins 2002, Maréchal et al. 2003, Dewandel et al. 2006, Owen et al. 2007, Lachassagne 2008, Lachassagne et al. 2008, 2009, Foster 2012).

In glacigenic areas, such as Fennoscandia and northern North America, the weathering zone is often non-existent or very thin. Preglacial weathering surfaces have survived in a few places only. The secondary porosity, which can vary considerably even over relatively short distances, is mainly determined by the intensity, orientation, connectivity, aperture, length, shape, roughness and infill of fracture and fault systems (e.g. Zhang & Sanderson 1995, Singhal & Gupta 1999, Aydin 2001). This means that the flow of water and potential for groundwater resources in the rock is largely determined by the distribution of interconnected fractures and the way, in which different factors control fracture apertures and the reactivation potential of existing faults and fractures (e.g. Sayers 1990, Odling 1993, Skjernaa & Jørgensen 1994, Barton et al. 1995a, 1995b, 1995c, 1997a, 1997c, Banks et al. 1996, Gudmundsson et al. 2003, Morin & Savage 2003, Denny et al. 2007).

Sometimes the term 'hard rocks' is used as a synonym for crystalline rocks. Larsson et al. (1984) defined hard rocks as igneous and metamorphic, non-volcanic and non-carbonate rocks. The same approach was accepted by Gustafson and Krásný (1993, 1994), Lloyd (1999a, 1999b) and Banks et al. (2005). Gustafsson (1993, 1994) has proposed that the term 'hard rock' should, from a groundwater exploration point of view, include all rocks without sufficient primary porosity and conductivity for feasible groundwater extraction. According to Krásný (1996a, 1996b, 1997), crystalline rocks and hard rocks cannot be considered synonyms from a hydrogeological point of view: the content of hard rocks has to be wider than that of crystalline rocks. In the review of Olofsson et al. (2001) the term hard rock has a wider meaning and incorporates the hard sedimentary and volcanic rock types, volcanic tuff among others. Recently Krásný (2003) defined hard rocks as crystalline igneous and metamorphic rocks and sedimentary highly cemented and/or folded rocks (see also Kellgren & Sander 2000, Kellgren 2002). According to Solomon and Quiel (2006), the term 'hard rock' commonly applies to hard and dense rocks with the main part of the groundwater flowing in secondary structures, mainly fractures.

In this study, the approach of Larsson et al. (1984) is adopted while at the same time the study takes advantage of the groundwater literature about hard rocks in a wider meaning.

1.3 Crystalline rocks as a common water supply

The role of crystalline rocks as a general water supply source is fairly limited due to low transmissivity and scanty well yields (e.g. Davis & DeWiest 1966, Knutsson & Fagerlind 1977, Wright 1990b, Wright & Burgess 1992). Typical (short-term) well yields in unweathered crystalline rocks usually range from some hundred to a couple of thousand liters per hour (e.g. Gustafson 2002, Banks et al. 2010). In basement areas there are, however, many bedrock wells producing more than 4,000 Lhr^{-1} of good quality groundwater, which shows that crystalline rocks can offer a substantial water source even for small towns and villages not to mention single or a few households and farms (e.g. Englund et al. 1988, Mäkelä 1990b, 1990c, 1990d, 2004).

In Finland, there are nearly 200 water works supplying bedrock groundwater for small villages and municipalities (Isomäki et al. 2007). For instance, the water supply of the municipality of Leppävirta in east-central Finland is based on bedrock groundwater pumped daily around 1,200 m^3 from eight drilled wells (Breilin et al. 2003). In Central Finland, the common water supply of the municipality of Luhanka is arranged on a single bedrock well whereas in the municipality of Konnevesi half of the groundwater needed is extracted from bedrock. In the municipality of Uurainen, bedrock groundwater is used whenever there is water shortage in the main water works. In addition to this, there are 24 private water works supplying bedrock groundwater to local consumers in small villages around Central Finland. The total amount of bedrock groundwater extracted per water work varies from a few thousand liters per day to more than 100 m^3d^{-1}.

To date, well drilling in most crystalline-rock areas for the needs of common water supply has been associated with a high risk of failure resembling sometimes a game of chance (Davis & DeWiest 1966, Braester & Barak 1991, Banks et al. 1994, 2010, Sander 1997, Banks & Robins 2002, Henriksen 2003a). Major problems in evaluating groundwater availability in fractured bedrock are its extreme variability in water-bearing properties and highly localized water-producing zones, which make geological and geophysical exploration challenging (e.g. Larsson et al. 1972, Dijon 1977, Banks 1992a, Sander 1999, Fernandes & Rudolph 2001, Moore et al. 2002a). It is difficult to predict consistently whether a borehole at a given location will yield respectable quantities of water. Neither are there any really reliable guidelines for the location of high-yield boreholes in crystalline bedrock aquifers (Banks 1992a, Banks et al. 1992a, 1992b, 1994, Sidle & Lee 1995). The practice of assuming that topographically and geophysically prominent fracture zones will always give elevated yields has been shown to be, at best, rather unreliable (Carlsson & Olsson 1977a, 1977b, Banks et al. 1992a, 1992b, 1993, Filho & Rebouças 1995).

Davis and DeWiest (1966) have suggested that few tasks in hydrogeology are more difficult than locating drilling sites for water boreholes in igneous and metamorphic rocks. Mahmood (1996) puts this as follows:"Nowhere is water scarcity felt more seriously than in those areas underlain by crystalline, igneous, and metamorphic rocks". According to Drew et al. (1999), crystalline bedrock aquifers are among the least understood and quantified groundwater resources in the United States. Faybishenko and Benson (2000) state that among the current problems that hydrogeologists face, perhaps there is none as challenging as the characterization of fractured rock. Titus et al. (2009) have proposed that basement aquifers are notoriously complex to develop, especially where a thinner weathered overburden is present or limited in extent. Hence, it is not surprising that the percentage of unsuccessful wells in some igneous and metamorphic regions may be high even though well sites have been located by experienced hydrogeologists (Davis & DeWiest 1966, Kath & Crawford 2001).

1.4 Purpose of the study

Groundwater exploration for a common water supply is a challenging task in crystalline rock areas. The key issue in defining exploration targets is to be able to combine various geologic, topographic and remote-sensing data to geophysical, hydrogeologic and borehole information (Anon 1989a, Sander 1997, Dapaah-Siakwan & Gyau-Boakye 2000, Srinivasa Rao et al. 2000, Murty & Raghavan 2002).

Groundwater in (unweathered) bedrock concentrates along lines of structural weakness, such as faults, fracture zones and dikes, which often show as linear features on satellite imagery and topographic maps, generally referred to as lineaments (e.g. Greenbaum 1992, Gustafsson 1994). Structural analysis and understanding of the tectonic evolution of a given area can provide useful information for hydrogeologists involved in regional-scale studies. This knowledge may help to identify promising lineaments according to their relation to the stress directions and origin (Fernandes & Rudolph 2001, Johansson 2005). On the other hand, in impoverished arid countries, the development of efficient, simple and low-cost methods for groundwater exploration is of utmost importance (Neves & Morales 2007a). It is also crucial to identify the reasons for success or failure encountered in the construction of wells in crystalline rock areas (Dijon 1971).

The hydrogeological characteristics of crystalline-rock aquifers have been intensively studied worldwide (e.g. Davis & Turk 1964, Larsson et a. 1984, Gale et al. 1985, Houston & Lewis 1988, Wright 1990a, 1992, Wright & Burgess 1992, Gustafson & Krásný 1994, Chilton & Foster 1995, Sander 1997, Lloyd 1999a, Taylor et al. 1999, Krásný et al. 2003, Krásný & Sharp 2007a). Numerous studies have shown the importance of both weathering, leading to the development of a storative weathering cover and to transmissive fissuration (e.g. McFarlane et al. 1992, Taylor & Howard 2000, Laghassagne et al. 2001, Maréchal et al. 2004, Courtois et al. 2010), and tectonic fracturation (e.g. Greenbaum 1992, Howard et al. 1992, Yin & Brook 1992a, 1992b, Bradbury & Muldoon 1994, Banks et al. 1996, Mabee & Hardcastle 1997, Sander et al. 1997, Gbürek et al. 1999, Mabee 1999, Magove & Carr 1999).

As Banks (1992a) and Banks et al. (1994) have earlier stated, relatively few studies have, however, attempted to characterize groundwater in fractured aquifers on a coarser regional scale, from a practical, water-resources point of view. This is despite the importance of such aquifers, not only to much of the Third World (e.g. Robins et al. 2006), but also to large areas of Europe, such as Fennoscandia, and North America. Additional information is especially needed on the hydrogeological factors that influence well productivity so that high-yielding well sites can be located more easily and the potential productivity of a site can be estimated before drilling begins. There is also a clear need for systematic research on basement aquifers to improve knowledge of the groundwater resource occurrence and methods of exploration and development (e.g. Brook 1988, Gustafson & Krásný 1994).

The overall objective of this study is to increase understanding of crystalline rock aquifers in unweathered basement areas. For this purpose a representative database with about 2,500 drilled wells (private wells, test wells) was compiled and well data analysis was executed. Well data analysis consisted of statistical examination of well construction parameters (drilled well depth, well diameter, casing depth, depth to the first (main) water strike, depth to the groundwater table, length of the saturated open borehole) and short-term well yield and that of various interactions and relationships between them. Based on well data analysis and single well pumping tests, estimates for hydraulic parameters, which include normalized yield, specific capacity, well productivity, transmissivity and bulk hydraulic conductivity, were statistically determined. Together with well yield these hydraulic parameters may be called as well production properties.

Another objective of the study is to identify local and regional (hydrogeological) factors, which may affect well yields and hydraulic properties of crystalline rock aquifers. Well yields can vary considerably from one location to the next and can exhibit extreme variations even within the same rock type. Many researchers have tried to explain these variations by examining statistically the specific geologic and other factors that influence the development and distribution of hydraulic properties in the bedrock (e.g. LeGrand 1954, Siddiqui & Parizek 1971, Daniel 1987, 1989, Morland 1997, Moore et al. 2002a, Henriksen 2003a, 2006b). The goal of these studies has been to search for a factor or a combination of factors that are most influential in controlling yield and to use this information to help locate high-yield bedrock well sites with greater reliability. Those hydrogeological factors, which correlate with well yield and hydraulic properties, may be further used to identify areas of enhanced permeability and potentially good aquifer zones at reduced exploration costs in the same or similar areas even when primary well-production data are lacking (e.g. Mabee 1992, 1999, Kellgren & Sander 1997, 2000, Edet et al. 1998).

According to various studies, factors affecting well yield and hydraulic properties in crystalline rock areas include geomorphologic situation, erosion surfaces, topographic setting, size of the drainage area up-gradient of a well, rock type, overburden type, overburden thickness, structural setting, joint and fracture characteristics, dip of bedrock strata, proximity to lineaments and lineament intersections, total well depth, borehole construction, depth to the groundwater table, saturated aquifer thickness, rock stress, tectonic history, neotectonic activity, annual crustal uplift rate in response to the last glaciation, climate, rainfall, the amount of recharge from precipitation, runoff, and the proximity to surface-water bodies (LeGrand 1954, 1967, Poth 1968, Siddiqui & Parizek 1971, 1974, Jammallo 1984, Brook 1985, 1988, Daniel 1987, 1989, 1990, Houston & Lewis 1988, Clarke & McFadden 1991, Huntley et al. 1991, Wikberg et al. 1991, Zewe & Rauch 1991, Henry 1992, Mabee 1992, 1999, McFarlane et al. 1992, Wright 1992, Yin & Brook 1992a, Briz-Kishore 1993, Helvey & Rauch 1993, Henriksen & Kyrkjeeide 1993, Knopman & Hollyday 1993, Kastrinos & Wilkinson 1994, Mabee et al. 1994, 2002, Rohr-Torp 1994a, 1995, 2000, Chilton & Foster 1995, Henriksen 1995, 2003a, 2003b, 2006b, 2008, Banks et al. 1996, Sami 1996, Sander et al. 1996, Morland 1997, Stibitz 1998, Singhal & Gupta

1999, Wladis & Gustafson 1999, Krásný 2000, Fernandes & Rudolph 2001, Eftimi 2003, Morin & Savage 2003, Kenny et al. 2006, Neves & Morales 2007a, 2007b, Verbovšek & Veselič 2008, Holland & Witthüser 2009).

Some of the factors are important at the local scale, others at the subregional scale, and still others at the regional scale (Henriksen 2003a, Henriksen & Braathen 2006). Despite the coherence of such factors local conditions may dominate yield and well response at a given site. Reliable predictions of potential well yield cannot be guaranteed even by putting considerable effort into the assessment of existing data and the use of a variety of siting techniques (Chilton & Foster 1995). Moreover, factors affecting well yield in one geologic setting may not be directly transferable to another (Mabee 1999).

A specific goal of this study is to unravel the relationships between well production properties and different seismo- and neotectonic processes. Seismotectonics is the synthesis of earthquake, geophysical, geodetic and geological data to deduce the tectonic framework of a region (Scholz 1990). It is concerned with the relationship between the seismological characteristics of (present-day) earthquakes and tectonics (Stewart & Hancock 1994). A neotectonic structure is interpreted as having been propagated or reactivated in a stress/strain field that has persisted without significant change of orientation to the present day, i.e. neotectonic structures have developed in the current tectonic regime (Muir Wood 1993a, Stewart & Hancock 1994). For instance, in the onshore Canning Bain, Australia, neotectonic fracture systems in granite pavements cut aboriginal petroglyphs (Hillis 2008). For the purpose of this paper, neotectonic processes are defined as those which have occurred during the last 30,000 years including the last glacial maximum (LGM), which was about 20,000 years ago (Lundqvist 1992, Donner 1995, Svendsen et al. 2004, Forsström 2005, Lokrantz & Sohlenius 2006, Lunkka 2007). Rock stresses are an essential element in seismotectonic and neotectonic processes.

The stress-dependency of permeability has been a known issue in hydrogeological sciences and in petroleum engineering. Because the fractures constitute the main conduits for the fluid flow and fracture geometric variations are sensitive to the stress changes, the study of the coupled stress-permeability relationships have attracted significant attention, particularly in the last two or three decades (e.g. Stephansson 1986, Amadei & Stephansson 1997, Chen & Bai 1998, Jolly et al. 2000, Grollimund et al. 2001, Meyer 2002, Lu et al. 2006, Tamagawa & Pollard 2008, Zoback 2011). According to Banks et al. (1996), there are two schools of thought with respect to stress influence on permeability: some workers have considered the relationship between the orientations of the past tectonic stresses and the permeability of the fracture zones, while others have studied permeability from the point of view of current in-situ stress field.

As early as 1951, Meier and Petersson suggested that the reason for the remarkably high average yield of drilled wells in certain areas of western Sweden was reactivation of faults in connection with postglacial and even recent earthquakes. Thus the joints and fissures have been kept open, contrary to other areas where joints may be regarded as healed (Meier & Petersson 1951). Later also Mörner and Tröften (1989) have presented the idea to get groundwater from reactivated postglacial fault fractures in Sweden. Banks et al. (1992a) have for their part pointed out that evaluation of most important geological processes affecting the water-yielding properties of crystalline bedrock should include, in addition to secondary mineralization and overlying deposits, earlier and current stress fields and neotectonic (postglacial) fault movement.

Morland (1997) has argued that the Quaternary, with several glaciations and deglaciations, has been the single most hydrogeologically important period of fracture formation and reactivation in Fennoscandia. According to Olesen et al. (2000a, 2000b), there was most likely a major seismic pulse in mainland Fennoscandia accompanying each of the deglaciations that followed the multiple glaciation cycles during the last 600,000 years.

They propose that neotectonic reactivation of regional fault zones may consequently have resulted in contraction and dilation of fissures in the bedrock.

In Canada, Martini and Bowlby (1991) have supposed that the same zones of crustal weakness in the Lake Ontario Basin were most probably reactivated under the influence of the repeated cycles of supracrustal glacial ice loading and unloading. Thorson (1996, 2000) and James et al. (2000) have suggested that glacial loading and unloading in the Puget Sound region in the Pacific Northwest may have controlled the timing of movement on crustal faults, namely the Seattle fault, situated above the Cascadia subduction zone. Kohut (2006) has likewise proposed that the periodic glacial loading and rebounding in Vancouver Island, southwestern Canada, resulted in some changes to fracture permeability in the bedrock with differential closing and opening of some faults and fractures.

However, as Olofsson et al. (2001) state, it is not known whether such neotectonic disturbance have increased the flow possibilities of groundwater or improved the well yields in crystalline bedrock. Neither does there appear to be any consistent regional association between groundwater and postglacial faulting (Munier & Fenton 2004). According to Chen and Bai (1998), the shortage of sufficient representations of such dependency consistent with the field conditions is primarily due to the lack of measured data. Yet, if relationships between seismo- and neotectonic processes and bedrock well yield and hydraulic properties can be shown to exist, then this can provide a valuable tool to improve the selection of borehole drilling localities (Banks et al. 1992a, 1996, Sander et al. 1996, 1997, Edet et al. 1998, Nascimento da Silva & Jardim de Sá 2000, Owen et al. 2003, Henriksen 2006b).

Thirdly, the study aims at finding general guidelines that may be useful in locating sites for high-yield drilled wells. High-yield wells are defined as those that can be used for the common water supply of villages and municipalities in crystalline rock areas. The yield of these wells is commonly a few thousand liters per hour, at least. High-yield well siting for long-term groundwater resource development and management requires identification and delineation of zones of bedrock with high storativity and transmissivity and those with a good recharge potential (Daniel 1989, Howard et al. 1992, Greenbaum et al. 1993, Clark et al. 1996, Lyslo 2000, Moore et al. 2002a, Robinson 2002). An intent of this study is to develop a hydrotectonic model in which various available data can be assessed and integrated to form a valuable well-siting tool and a decision support for the selection of targets for detailed field investigations and which can be used in crystalline rock areas for quantifying bedrock aquifer characteristics in order to increase the drilling success rate of high-yield bedrock wells (Sander et al. 1996, Sander 1997, Johansson 2005).

It is not only the quantity but also the quality of groundwater that is of interest while exploring the potential of water supply (e.g. Banks et al. 1998, Johansson 2005). Considerable research in this respect has been done in Central Finland, too (Mäkelä 1990d, 1993, 1994b, Mäkelä & Rönkä 1992, Sara 2001; Appendix I). In this thesis, however, the main point is focused on bedrock well yield and hydraulic properties and factors affecting on them. Hence, groundwater quality issues are not objectives of this paper. Yet, groundwater quality is briefly discussed, from a water consumption/supply point of view, in connection with general information on bedrock wells in Central Finland.

2 HYDRAULIC PROPERTIES

2.1 General

Hydrogeologists and water engineers need to be able to estimate the safe, or reliable, yield of a water-supply well or well field. This information is essential for proper groundwater planning and resource management (e.g. water balance), yet it can be difficult to acquire. The reliable yield of a well depends on a complex interplay of many factors including, for instance, the hydraulic properties of the aquifer and the construction and present condition of the well (Misstear & Beeson 2000). Knowledge of the statistical distribution of hydraulic properties of crystalline bedrock is also a prerequisite for a cost-effective drilling strategy in such aquifers (Banks et al. 2010).

The well yield and hydraulic properties of crystalline rocks largely depend on the physical and geometrical properties of their fracture network. They commonly display extreme variations over short distances, which may be referred to as hydraulic heterogeneity (Freeze & Cherry 1979, Anon 1996, Sander et al. 1996). This may be demonstrated, for example, by the common observation that only a few percent of interconnected fractures noted in boreholes are hydraulically conductive to any significant extent, that the few producing fractures or the water table in closely neighboring wells can be at significantly different elevations, and by the lack of response in some observation wells during aquifer tests (e.g. Ahlbom & Smellie 1991, Caine & Tomusiak 2003).

A large number of analytical methods are available for determining hydraulic properties from pumping-test data. However, analyses of pumping tests in crystalline rock aquifers do not satisfy many of the assumptions inherent in different analytical methods: crystalline rock aquifers are neither homogeneous nor isotropic, flow in fractures is usually channelized with fractional flow dimension, well storage is not negligible relative to well yield, groundwater flow is complicated by great drawdown and partial penetration of well etc. (e.g. Barker 1988, Jones 2001, Leveinen 2001, Lods & Gouze 2004, 2007, 2008). Thus, it is not surprising that convincing field examples of thoroughly characterized bedrock aquifers are lacking in the literature (Snow 1968, Uhl & Sharma 1978, Black 1994, Gernand & Heidtman 1997, Misstear et al. 2006).

Maréchal et al. (2003, 2008) have introduced four different methods, namely the Neuman, Gringarten, Warren & Root and Barker methods, well adapted to complexities of groundwater flow in hard-rock aquifers. These methods consider the heterogeneity and the anisotropy of the medium: anisotropy of permeability, single horizontal fracture intersecting the pumping well, double-porosity behavior or connectivity, and fractional dimension flow, respectively. Maréchal et al. (2003, 2008) suggest that these methods should be widely used by scientists and engineers for fractured aquifer characterization and evaluation. However, the methods need good test data (involving step- and constant-rate tests, with water-level measurements in both abstraction and observation wells), which are not found as often as one would wish. For the majority of private wells, test data (if any) may be restricted to short-term yield and specific capacity values (e.g. Misstear 2001).

Single-well tests are tests in which no piezometers are used. Water-level changes during pumping or recovery are measured only in the pumped well itself (e.g. Kruseman & de Ridder 2000). Simple equilibrium approximation formulae can be used in regional studies to provide initial estimates of aquifer hydraulic properties in the many situations where only simple air-lift yield data or limited time-drawdown data are available from single well pumping tests. Of course, these equilibrium approaches have significant limitations: they cannot be used to determine the aquifer storage coefficient and may lead to large errors in estimating hydraulic parameters where equilibrium conditions have not been reached, or have been achieved due to leaky-aquifer conditions or the occurrence of recharge-boundary effects (Misstear & Beeson 2000, Misstear 2001, Maréchal et al. 2003, Misstear et al. 2006).

Water well drillers in Finland do not usually conduct pumping tests with drawdown measurements or even measure the water level in the borehole after drilling has been completed. Neither do well owners nor do operators carry out these tests afterwards. For these reasons, hydraulic parameters other than short-term well yield are seldom available in different data sets. According to Henriksen (1995), well yield reflects the state of the hydromechanical system in the nearest vicinity of the borehole and specific capacity, transmissivity and hydraulic conductivity relate in some way to well yields in bedrock aquifers. Hence, the well yield should be considered an adequate statistical test observator for a comparative study of the hydraulic systems in different hydrogeological settings (Henriksen 2006b).

2.2 Hydraulic parameters

Well yields have proved to be of great importance in characterizing groundwater potential and hydraulic properties of crystalline rocks in different regions (e.g. Rohr-Torp 1994a, Henriksen 1995, 2003a, Cohen et al. 1996, Morland 1997, Wladis & Gustafsson 1999).

The yield of a drilled well (Q Lhr^{-1}) is generally obtained by air-lift or pumping test carried out by the driller at the termination of the drilling. It is a short-term yield, i.e. maximum recharge rate that can be sustained from the well during a relatively short period of time. Information concerning the associated drawdown is generally lacking. The air-lift yield is also known as flushing yield or air-blow yield.

Such short-term air-lift or pumping tests are not representative of a large aquifer volume, but of the relatively small number of fractures that transmit water to the well from some form of broader fracture network/aquifer storage. The yields would thus be expected to be determined by the transmissivity of these “feeder fractures” in the limited zone of rock intersected by the well (Banks et al. 2010). The short-term yield still gives a rough measure of the actual sustainable well yield.

There is a very wide distribution of borehole yields in crystalline rocks, the distribution typically being heavily skewed towards low yields and approximately log-normally or power-law distributed (e.g. Henry 1992, Huntley et al. 1992, Banks et al. 1993, 2010, Holland & Witthüser 2009, 2011).

2.2.1 Normalized yield

Normalized (linearized) yield (Q/d $Lhr^{-1}m^{-1}$) is the yield of a drilled well (Q) divided by the drilled rock depth (d), i.e. total well depth – thickness of soil cover (overburden; e.g. Morland 1997). Davis and Turk (1964) have defined the normalized yield as ‘yield per foot’, which is calculated by dividing the total yield of the well by the depth of the well below the water table.

Owing to the overall availability of depth and simple air-lift yield data, normalized yield can facilitate a regional characterization of hydraulic properties of rock in an initial stage and can form the basis for more detailed studies. Indeed, normalized yield is the most common estimate for hydraulic bedrock properties (Randolph et al. 1985, Daniel 1989, Rohr-Torp 1994a, Wladis & Rosenbaum 1994, Henriksen 1995, 2003a, 2008, Wladis 1995, Krásný 1996c, Morland 1997, Dagestad et al. 2003) and it has been successfully applied in geostatistical studies of transmissivity and hydraulic conductivity in crystalline rock areas (Wallroth & Rosenbaum 1996, Wladis et al. 1997, Follin et al. 1999, Wladis & Gustafson 1999).

2.2.2 Specific capacity and well productivity

Specific capacity (Q/s $Lhr^{-1}m^{-1}$, m^2s^{-1}) is a measure of the productivity of a well. It is defined as the volume of water pumped per unit time (Q) per unit drawdown (s) (e.g. Freeze & Cherry 1979, Mace 2001) or as the sustainable pumping rate divided by the drawdown in the well at a quasi steady state (Singhal & Gupta 1999, Rayne et al. 2001).

Specific capacity in fractured aquifers is approximately log-normally distributed (Jetel 1968, Carlsson & Carlstedt 1977, Davis 1986, Knopman 1990, Huntley et al. 1992, Knopman & Hollyday 1993, Gustafson & Krásny 1994, Fabbri 1997, Wladis & Gustafson 1999, Jalludin & Razack 2004, Razack & Lasm 2006).

Specific capacity is preferred to yield as a measure of well productivity because it accounts for the loss in head that is associated with pumping of water. Specific capacity thus normalizes the effects of pumping rate on drawdown (Knopman & Hollyday 1993, Naik et al. 2001). Specific capacity is a function of aquifer setting, well setting and pumping setting. That is, specific capacity depends not only on aquifer transmissivity and storativity and boundary conditions within the aquifer but also on well diameter, well condition, open well section, partial penetration ratio of the well, well loss correction and pumping rate and time (Lattman & Parizek 1964, Jetel 1968, Summers 1972, Carlsson & Carlstedt 1977, Clark 1977, Viswanathiah & Sastri 1978, Theis et al. 1983, Daniel 1989, Carlsson & Gustafson 1991, Yin & Brook 1992a, Knopman & Hollyday 1993, Olofsson 1994, Mace 1997, 2000, 2001, Cesano et al. 2000).

Normalized yield Q/d has been used as a conservative estimate for specific capacity Q/s, where s is the drawdown, because $d \geq s$ (Wallroth & Rosenbaum 1996, Wladis & Gustafson 1999, Johansson 2005). Henriksen (2006b, 2008) has suggested that the saturated thickness of a well, that is available drawdown, is a better estimate for drawdown than the drilled rock depth, and the approximation for specific capacity obtained by this method should be considered a better estimate of the specific capacity than the normalized yield.

The specific capacity derived from normalized yield may be biased to some extent due to the following reasons (Rhén et al. 1997): drilling stops after the requested amount of water is achieved, drawdown in high-yield wells is generally less than the depth of the wells, dry or low-capacity wells may have not been reported and somewhat more wells are drilled close to lineaments. These factors largely compensate each other in a large data set, so there may not be any significant bias left in the well data (Rhén et al. 1997).

Well productivity or specific capacity index (Q_w ms^{-1}, md^{-1}) is the specific capacity normalized to aquifer thickness (e.g. Davis & DeWiest 1966). It is computed by dividing specific capacity by the saturated thickness of the well, that is, by the available drawdown (Lattman & Parizek 1964, Siddiqui & Parizek 1971, 1974, Walton 1977, Brook 1988, Singhal & Gupta 1999). In this respect it may be used as an approximate for bulk hydraulic conductivity (Singhal 1977, Follin et al. 1999, Henriksen 2006b, Rotzoll & El-Kadi 2008, Verbovšek 2008, Banks et al. 2010).

The well productivity can be further normalized to a uniform aquifer thickness by multiplying it with say 100 m. The result is the 'yield factor' (Q_f m^2s^{-1}, m^2d^{-1}), which is the specific capacity to a 100 m thick aquifer (Poland 1959, Thomasson et al. 1960).

2.2.3 Transmissivity

Transmissivity (T m^2s^{-1}, m^2d^{-1}) is the rate at which fluid, such as water, of prevailing kinematic viscosity is transmitted through a unit width of the aquifer under a unit hydraulic gradient. It is equal to Kb, where K is the hydraulic conductivity and b is the thickness of the aquifer (Lohman 1972).

Transmissivity is a transmission property, and variations in T are due to variations in media properties, the saturated thickness, and the properties of transmitting fluid (e.g. Acheampong & Hess 1999, Raj 1999). Therefore, transmissivity represents the hydraulic characteristic of the aquifer whereas specific capacity can be a function of how efficiently a well is completed and developed (De La Garza & Slade 1986). Transmissivity is often regarded as the primary hydraulic property that can be estimated from consideration of drilled well yields (e.g. Banks et al. 2010).

The lognormality of the transmissivity is well recognized in the literature for fractured aquifers (Krásný 1975, Huntley et al. 1992, Fabbri 1997, Banks 1998, Mace et al. 1999, Moore et al. 2002a, Jalludin & Razack 2004, Razack & Lasm 2006, Lasm et al. 2008). Gustafson and Franson (2006) argue for a power law distribution of fracture transmissivities. Normalized transmissivity [T_n ms^{-1}] is the transmissivity divided by the length of uncased saturated borehole (Mabee 1992).

The transmissivity typically ranges over several orders of magnitude in a single rock type. In addition, a general decrease in the transmissivity of fracture zones with depth has been noted (Davis & Turk 1964, Carlsson & Olsson 1977a, 1977b, Black 1987, Houston & Lewis 1988, Krásný 1996c, Zhang & Sanderson 1996a, Havlík & Krásný 1998, Taylor & Howard 2000, Darko & Krásný 2003, Krásný & Sharp 2007b, Rhén et al. 2007, Lasm et al. 2008). Darko and Krásný (2000) argue that this seemingly inconsistent trend may be explained by the fact that many of the wells are terminated just after enough yields are obtained. According to Banks et al. (2010), the transmissivities of bedrock aquifers increase with drilled depth, although the transmissivities of individual fractures decrease with drilled rock depth. This means that the bulk transmissivity of the portion of the aquifer down to say 100 m only accounts for around half of that of the total transmissivity of the full depth of the rock mass (Banks et al. 2010 and their Fig. 5).

In order to compare objectively transmissivity values on local and regional scales, Krásný (1993a) introduced a classification of transmissivity magnitude and variation. The classification consists of six classes of transmissivity magnitude (I – very high, more than 1,000 m^2d^{-1}, to VI –imperceptible, less than 0,1 m^2d^{-1}) and of six classes of transmissivity variation (class a - insignificant variation, indicating almost homogeneous hydrogeologic environment) to class f – extremely large transmissivity variation, i.e. extremely heterogeneous hydrogeologic environment). For instance, in the Bohemian Massif, the regionally prevailing transmissivity ranges commonly between classes V to III (0,1-100 m^2d^{-1}), the most frequent being the class IV(-III)c,d (units to slightly more than 10 m^2d^{-1}) with moderate to large transmissivity variation (Krásný 1996a). Comparison of regional prevailing transmissivity indicates only small differences in distinct bedrock areas and can be expressed by the classes IV(V, III) c,d (Carlsson & Carlstedt 1977, Krásný 1996c, 1997, 2002, Staśko & Tarka 1996, Darko & Krásný 1998, Chambel et al. 2003, Krásný & Sharp 2007b).

To approximate transmissivity in different aquifers, the famous Thiem´s equilibrium equation has widely been used. It can be written as

$$T = (2{,}3Q/2\pi s_w) \log r_e/r_w \qquad (1)$$

where T is the aquifer transmissivity (m^2d^{-1}), Q is the well discharge rate (m^3d^{-1}), s_w is the maximum well drawdown (m), r_e is the radius of influence of the well (m), and r_w is the well radius (m) (e.g. Singhal 1977, Singhal & Gupta 1999).

Various simplifications of the Thiem´s equilibrium formula have been developed. Thomasson et al. (1960) modified the Thiem equation for confined-aquifer conditions in valley-fill sediments as

$$T = A_1 (Q/s) \quad (2)$$

where Q/s is the specific capacity and A_1 is a constant which ranges from 0,9 to 1,52 with a mean of 1,18.

Logan (1964) proposed a value of 3,32 as "typical" for the log ratio r_e/r_w and gave the approximation of the Thiem equation for confined sand and gravel aquifers as

$$T = 1{,}22 (Q/s) \quad (3)$$

Specific capacity (Q/s) has commonly been employed to estimate transmissivity of hard rock aquifers, because of the general availability of specific-capacity data from driller´s logs and the relative expense in obtaining transmissivity through aquifer testing (Jetel 1968, Krásný 1975, Carlsson & Carlstedt 1977, Theis et al. 1983, Banks 1992b, Huntley et al. 1992, Knopman & Hollyday 1993, El-Naqa 1994, Fabbri 1997, Mace 1997, 2001, Rhén et al. 1997, Acheampong & Hess 1998, Meier et al. 1999, Singhal & Gupta 1999, Hamm et al. 2005, Razack & Lasm 2006, Rotzoll & El-Kadi 2008). Banks (1992b) has set forth the term 'apparent transmissivity', because specific capacity derived from short-term capacity testing only yields information about the aquifer in the vicinity of the borehole (see also Meier et al. 1999, Mace 2001).

2.2.4 Hydraulic conductivity

Hydraulic conductivity (K ms^{-1}, md^{-1}) of a rock is a measure of its ability to transmit a unit volume of fluid, such as water, in unit time at the prevailing viscosity through a cross section of unit area, measured at right angles to the direction of flow, under a hydraulic gradient of unit change in head through unit length of flow (Lohman 1972).

Hydraulic conductivity depends both on the properties of the medium (rock) as well as of the fluid (e.g. Singhal & Gupta 1999). The permeability (κ m^2) is a property of the medium only, not the fluid. It is a portion of K and can be calculated as

$$\kappa = K (\mu/\rho g) \quad (4)$$

where μ is the dynamic viscosity ($kgm^{-1}s^{-1}$), ρ is the density of the fluid (kgm^{-3}) and g is the acceleration due to gravity (ms^{-2}) (e.g. Fetter 1994).

Hydraulic conductivity of crystalline rock aquifers depends on fracture characteristics: aperture distribution, surface roughness, contact asperities area, shape, infilling, and length of fractures and on the degree of interconnection between them (e.g. Long & Witherspoon 1985, Houston 1992, Brown & Bruhn 1998, Muldoon et al. 2001). Few fractures are not always an indication of low hydraulic conductivity, or vice versa, high fracture frequency does not always mean high hydraulic conductivity (e.g. Anttila 1988, Johnson 1999). Rock gouge filling or cementation of fractures reduces the K-value even in the case of dense fracturing (Huntley et al. 1991). Field and laboratory hydraulic conductivity measurements for many rock types predominantly have a log-normal probability distribution (e.g. Bedinger et al. 1989).

Hydraulic conductivity can vary within several orders of magnitude within the same lithologic unit and within short distances. The variation is usually much higher within rock types and groups than between them. Fracture-network and fault-zone properties appear much more important than lithology when considering storage capacity and permeability of crystalline bedrock aquifers (e.g. Davis & Turk 1964, Hult et al. 1978, Brace 1980, Clauser 1992, Gustafson & Krásný 1994, Follin et al. 1999, Caine & Tomusiak 2003, Nastev et al. 2005).

Bedrock hydraulic conductivity shows a decreasing trend with depth (Ellis 1909, LeGrand 1954, Davis & Turk 1964, Davis & DeWiest 1966, Snow 1968, Landers & Turk 1973, Carlsson & Olsson 1976, 1977a, 1977b, 1981, Olsson 1979, Davis 1981, Gale 1982, 1983, Woolley 1982, Carlsson & Gidlund 1983, Carlsson et al. 1983, Gale et al. 1985, Johansson 1985, Ericsson & Ronge 1986, Daniel 1987, Hansson 1989, Soulios et al. 1990, Anon 1992, Stevenson et al. 1995, Stober 1996, Berggren 1998, Zhao 1998, Rutqvist & Stephansson 2003, Nastev et al. 2005, Rhén et al. 2007, Vaittinen et al. 2009, Boutt et al. 2010). A high hydraulic conductivity is usually found close to ground surface. This is assumed to depend on surface fracturing due to stress release, weathering and dissolution of fracture minerals (Gustafson & Krásný 1994, Wallroth & Rosenbaum 1996, Zhang & Sanderson 1996b, Singhal & Gupta 1999, Mazurek 2000). In addition, due to the general reduction of the aperture of joints with depth (weight of overlying rock, stress) and often the increase of spacing, the conductivity decreases with depth below the surface (Davis & DeWiest 1966, Carlsson & Olsson 1981, Gale 1983, Nilsen & Palmström 2000). According to Brusila (1983), decreasing hydraulic conductivity with increasing depth is a common finding of water loss measurements made in Finland.

There is, however, a large scatter of results, and even at considerable depth zones having high hydraulic conductivity have been reported (e.g. Carlsson & Olsson 1977a, Brace 1980). In the Outokumpu Deep Drill Hole R2500 (depth 2516 m), eastern Finland, high hydraulic conductivity was observed in the two uppermost sections at the depths of 480–550 m and 957–997 m (K-values $7{,}5 \times 10^{-6}$ ms^{-1} and $5{,}3 \times 10^{-7}$ ms^{-1}, respectively), whereas measured sections at deeper levels had very low hydraulic conductivity (Ahonen et al. 2011). According to Knutsson (1997), highly water-bearing fractures and fracture zones of regional dimensions down to a depth of 600-700 m have been documented at several sites in Sweden. Most of these zones are sub-vertical, but several sub-horizontal and horizontal zones are also found (Ahlbom & Tirén 1991). In Äspö granitoid rock area in Sweden the frequency of wet fractures peaks at depths of 100-200 m (Talbot & Sirat 2001) and the hydraulic conductivity below 100-200 m is fairly constant down to a depth of 500 m (Rhén et al. 1997). Fetter (1994) has reported higher permeabilities from fractures at depths of 664 to 1669 m in granitic rocks of Illinois, USA. The discovery of flowing water in fractures at depths of 11 km in Kola Peninsula, Russia, is direct evidence for fracture permeability at great depth (Kozlovsky 1982). On the other hand, Huntley et al. (1992) did not found any significant change in the permeability of crystalline rock with depth in San Diego, California, USA.

Leveinen (2001) has suggested that, instead of steadily decreasing hydraulic conductivity, the channeling of groundwater flow into the fracture zones in deeper parts of the bedrock might be possible. According to Hsieh (1996), only subsurface characterization on the scale of hundreds of meters may reveal the presence of highly permeable fracture zones or clusters of connected fractures. In tunnels most of the leakage normally occurs in the shallowest parts, but the most difficult leakages, due to the high pressures, are often experienced in the deeper parts (Nilsen & Palmström 2000). In the Central Alps, Switzerland, Masset and Loew (2010) describe several exceptionally high tunnel inflow rates (> 100 Ls^{-1}) at depths of several hundred meters. According to Singhal (2008), it may well be that although generally in fractured rocks a decrease in permeability with depth is observed, there is not much justification of such a universal rule. Jiang et al. (2009) have proposed that the depth decay of K may affect the regional groundwater flow in bedrock.

2.3 Well loss, skin effect and scale effect

In single well tests, the drawdowns are measured inside the pumped boreholes. Because of this, they are made up of two components: aquifer loss and well loss (e.g. Jacob 1947, Clark 1977, Kruseman & de Ridder 2000). The aquifer loss is that part of the drawdown caused by resistance to laminar flow within the aquifer. The well loss results from resistance to turbulent flow in the zone adjacent to the well, and through the possible screen. An additional component of well loss, which is important in deep wells, is the frictional head loss during flow up the well (Clark 1977). Well efficiency is the ratio of the drawdown in the aquifer to the drawdown in the well (e.g. Kruseman & de Ridder 2000). Well loss is directly proportional to discharge, and well loss coefficient is inversely related to well diameter, transmissivity and specific capacity. The well losses are constant in time but increase with increasing discharge rate (Gustafson 1974, Raj 1999, Mace 2001).

There are often significant well losses in wells completed into fractured-rock aquifers (Hahn 1977, Talley & Hahn 1978, Uhl & Sharma 1978, Atkinson et al. 1994, Tam et al. 2004, Izinyon & Anyata 2006). Well losses, which increase drawdown in the pumping well, decrease the specific capacity and can thus deteriorate considerably the simple analytical relationship between T and Q/s leading to underestimation of T (Razack & Huntley 1991, Huntley et al. 1992, Fetter 1994). The results of Jalludin and Razack (2004) indicate that correcting specific capacity values for turbulent head losses markedly improves the empirical relationships between transmissivity and specific capacity. Step drawdown tests or constant-head drawdown tests enable determination of well losses (e.g. Kruseman & de Ridder 2000).

The skin effect exists when the hydraulic conductivity around a pumped well is different from the overall hydraulic conductivity of the aquifer. The skin factor can be positive or negative (e.g. Almén et al. 1986, Andersson et al. 1993, Carlsson & Ntsatsi 2000). A positive skin factor indicates that the drawdown in the pumped well is more than is expressed by the drawdown in the aquifer at the borehole radius. A negative skin factor occurs when a zone of increased hydraulic conductivity is developed around the borehole (Andersson et al. 1993, Kaehler & Hsieh 1994, Carlsson & Ntsatsi 2000). The skin factor only affects estimates of transmissivity to a level of accuracy that is less than an order of magnitude (Wladis & Gustafson 1999).

In fractured rocks hydraulic properties are highly heterogeneous and flow is localized in a small part of the rock volume (Anon 1996). Because of this, hydraulic conductivity and transmissivity are quantities that vary depending on the scale of the investigation (Brace 1984, Clauser 1992, Rovey & Cherkauer 1995, Sánchez-Vila et al. 1996, Cook 2003). The scale dependence of the hydraulic properties in the rock mass is connected to the existence and the size of a potential representative elementary volume (REV). The REV can be regarded as the volume/dimension at which there are only small changes in bulk hydraulic conductivity for small changes in sample size or sample location (e.g. Banks et al. 2010). Below the REV, a medium appears as heterogeneous with a high variability of its properties. Above the REV it can be considered as a statistically homogeneous medium (Krásný 1996c, Olofsson et al. 2001, Zimmermann et al. 2003, Banks et al. 2010).

Brace (1980, 1984) pointed out that both porous sedimentary rocks and fractured crystalline rocks exhibit a scale effect, which means that the larger the measurement scale or volume the greater the hydraulic conductivity (Fig. 2). Large-scale permeability of granitic rocks is qualitatively different and quantitatively larger than small-scale permeability because large-scale permeability is controlled by a few fractures, which provide high connectivity to the system but are intersected by few testing intervals (Martinez-Landa & Carrera 2005). According to Clauser (1992), this is obvious in small distances or volumes (laboratory, borehole) but does not extend to a regional (crustal) scale, such as above 100 m, where the permeability becomes approximately constant (see also Hsieh et al. 1994, Rovey & Cherkauer 1995, Guéguen et al. 1996, Margolin et al. 1998, Follin et al. 1999, Schulze-Makuch et al. 1999).

Renshaw (1998) states that conductivity above borehole scale such as 100 m decreases with increasing scale. According to Rovey and Cherkauer (1995), scaling effects vary consistently with the type of geologic medium and degree of secondary porosity. The scale effect persists down to 9 km depth, at least, which seems to be in contradiction to the effect of pressure permeability that would be expected to cause downhole permeability and matrix permeability to converge (Huenges et al. 1997). Shapiro (2003) has shown that the magnitude of the bulk hydraulic conductivity of the rock mass is the same from aquifer tests over tens of meters and kilometer-scale estimates inferred from groundwater modeling. Le Borgne et al. (2006) have found that the borehole scale variability of transmissivity estimates vanishes at larger scale and that the transmissivity converges towards the high values of the transmissivity distribution.

Butler and Healey (1998) state that the observed scale effect is actually caused by different methods of measurement and by effects related to well installation and development. However, Schulze-Makuch and Cherkauer (1998) observed the same value of exponential scale increase whether using one or several methods of measurements within a given geologic unit. Their results indicate that hydraulic conductivity generally increases during an individual test as the volume of aquifer affected increases (see also Nastev et al. 2004) and that scale dependence of hydraulic conductivity during single tests does not depend on the method of measurement.

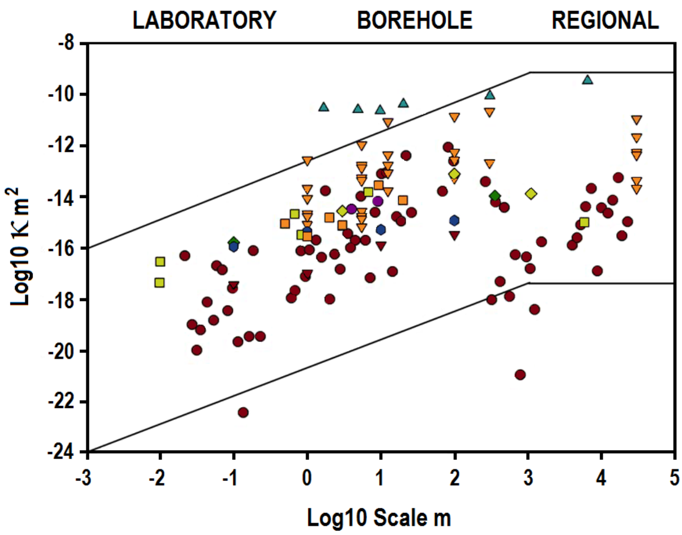

Fig. 2. Worldwide permeability (κ) data plotted against measurement scale. The data are compiled for various fractured geologic media, e.g. granite, schist, carbonates and karstic limestone; see Illman (2006) for legend of symbols. A general increasing trend is observed for data in the range of $-2<\log_{10}$ scale<3. Permeability does not appear to increase further with scale beyond the regional scale ($\log_{10}$ scale>3) for the available data. Modified from Illman (2006).

2.4 Groundwater potential indexes and maps

Different groundwater potential indexes and maps have been constructed in order to identify favorable areas for exploratory well positioning and drilling. They are based on various integrated hydraulic, geomorphologic, geologic, hydrologic, and geophysical features (e.g. hydraulic properties, topography, physiography, lithology, rock fracturing, lineaments, drainage, rainfall, geophysical properties) possibly weighted according to their believed importance (Venkateswara Rao & Briz Kishore 1991, Gustafsson 1994, Hardcastle 1995, Edet & Okereke 1997, Lachassagne et al. 2001, Sreedevi et al. 2005, Chandra et al. 2006, Ettazarini 2007, Adabanija et al. 2008, Prasad et al. 2008, Oh et al. 2011).

With statistical assessment of transmissivity Darko and Krásný (2003, 2007) and Darko (2005) prepared regional transmissivity anomaly maps for the quantitative characterization and prediction of borehole yields in hard rock aquifers in Ghana. According to them, detailed investigations in regional zones of positive transmissivity anomalies could reveal potential aquifers where large volumes of groundwater could be abstracted. The well yields expected from areas of positive anomalies were 6-15 times higher than in areas of negative anomalies.

The comparison between the calculated groundwater potential index and the corresponding yields of drilled wells in Morocco showed a positive correlation coefficient of 0,89 (Ettazarini 2007). Fracturing and lithology were the principal factors whereas drainage, topography and rainfall were secondary factors that had no influence unless the principal factors were favorable (Ettazarini 2007). The geomorphologic and hydrogeologic interpretations and the productivity data of drilled wells were integrated to make a map of the potential zones for obtaining groundwater in India (Sreedevi et al. 2005). The map portrayed three zones as poor, moderate and good.

The Geographic Information System (GIS) with its capability of map overlaying, reclassification, proximity analysis, and other mathematical operations can help to carry out criteria-based analysis of groundwater resources (Chesley et al. 1995, Kamaraju et al. 1996, Sander 1996, 2007, Ringrose et al. 1998, Banoeng-Yakubo & Skjernaa 2000, Kellgren & Sander 2000, Lachassagne et al. 2001, Anon 2002, Hung et al. 2002, Kellgren 2002, Singh & Ravi Prakash 2003, Solomon & Quiel 2006, Srivastava & Bhattacharya 2006, Prasad et al. 2008, Solomon & Ghebreab 2008).

Sander (1996) integrated his remote sensing interpretations with geologic and topographic maps, borehole yield data, and geophysical investigations in a GIS environment. Based on this he evaluated a probabilistic approach to groundwater exploration in hard rock areas of Ghana and Botswana. However, as Sander (1996) states, owing to the large heterogeneity of hard rock areas, groundwater potential maps may often be of limited use and the siting and construction of wells cannot be based on hydrogeological maps alone (see also Lachassagne et al. 2001). That is, a few tenths of meters may in some areas mean the difference between a high capacity well and a dry well (Lachassagne 2008), and a regional groundwater potential map is then not much help (Sander 2007). In addition, improved methods for assessing the fracture zone characteristics are needed (Struckmeier 1993, Sander 1999).

3 ROCK FRACTURES AND LINEAMENTS

3.1 Structural concepts and definitions

A fracture is a surface expression and a general term for a discontinuity, with fractures in rock including faults, veins and joints (e.g. Peacock & Mann 2005).

Fractures along which there has been no appreciable movement (shear) parallel to the fracture plane and only slight movement normal to the fracture plane are called mode I (opening-mode) fractures or joints (Fig. 3; e.g. Price 1966, Pollard & Aydin 1988, Palmström 1995, Park 2000). When there is sliding or tearing parallel to fracture plane, the fracture is called mode II or mode III shear fracture, respectively (e.g. Hatcher 1990, Twiss & Moores 2000).

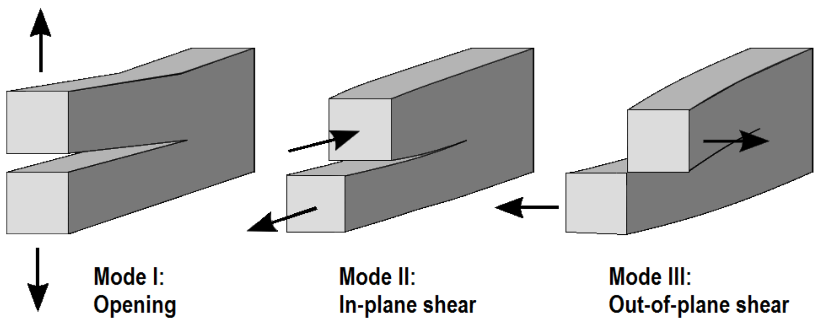

Fig. 3. Fracture modes. Modified from http://en.wikipedia.org/wiki/File:Fracture_modes_v2.svg.

A special kind of joint that forms subparallel to surface topography, generally in massive rocks, is called sheet joint or sheeting (LeGrand 1949, Price 1966, Carlsson 1979, Larsson 1987, Greenbaum 1990a, Hatcher 1990, Park 2000). Sheet jointing usually occurs in the superficial parts of the rock and decreases significantly with depth (Clapp 1911b, Barker et al. 1992). Sheeting is thought to form by unloading over long periods of time as erosion removes large quantities of overburden from a rock mass (Davis & DeWiest 1966, Hatcher 1990). Sheet joints are also suggested to form in response to torsion or compression, i.e. to tectonic stress (Vidal Romaní & Twidale 1999). Schoeller (1975) describes them as pressure relief fractures, which form along the sides and underneath valleys as a result of removal of rock load. There may also be surface-parallel fractures known as exfoliation (Roberts & Myrvang 2004). These are extensional fractures caused by high horizontal compressive stresses. They are a near-surface phenomenon and exfoliation joints are rarely seen deeper than 20-30 m below the present-day surface. Tensile (release) fractures are often created parallel to the fold axes (Einstein & Dershowitz 1990).

According to Peacock and Mann (2005), the geometry, frequency and distribution of fractures within rocks are controlled by rock characteristics and diagenesis (e.g. lithology, sedimentary structures, bed thickness, mechanical stratigraphy, and the mechanics of bedding planes), structural geology (e.g. tectonic setting, palaeostresses, subsidence and uplift history, proximity to faults, position in a fold, timing of structural events, mineralization, and the angle between bedding and fractures) and present-day factors (e.g. orientations of in situ stresses, fluid pressure, perturbation of in situ stresses, and depth).

Different trends in the number of fractures with depth have been noted. In the Fjällveden study site in Sweden the fracture frequency decreases with depth until 200 m. From 200 m and down to 700 m the average fracture frequency is more or less constant, about 2 fractures per meter (Ahlbom et al. 1991a). Johnson (1999) found that the distribution of fractures with respect to depth in New Hampshire crystalline rocks, USA, do not follow a linear distribution. Instead, the best fit was obtained by the power-law function. Highly fractured intervals typically occur in the top 100 m of bedrock. Although less numerous, some highly fractured zones exist at depths greater than 100 m (Johnson 1999). In the fractured-crystalline rock aquifer of the Turkey Creek in Colorado, USA, Poeter et al. (2002) found that both fracture density and well productivity are nearly constant with depth, not decreasing substantially until 210 m. According to Seeburger and Zoback (1982), the number of observable fractures does not decrease markedly with increasing depth in test wells drilled in crystalline rocks in the United States. According to Huntley et al. (1991), neither does the frequency of fractures nor the frequency of water-bearing fractures decrease with depth in crystalline rocks of San Diego, California, USA. In crystalline rocks of Massachusetts, USA, boreholes show decreasing fracture frequency up to 300 meters depth, with hydraulically active fractures showing a similar trend (Boutt et al. 2010).

A fault is a fracture having appreciable shear movement parallel to the plane of the fracture (Price 1966, Billings 1972, Hobbs et al. 1976, Palmström 1995, Singhal & Gupta 1999, Nilsen & Palmström 2000, Park 2000, Twiss & Moores 2000). Faults are usually revealed by their morphological expression, because fault zones are generally much more eroded than surrounding rocks due to their weakness and partly because of groundwater circulation and increased weathering are concentrated along them (Angelier 1994). Faults occur in many forms and dimensions. They may be hundreds of kilometers long or only a few centimeters. Their outcrop traces may be straight or sinuous. They may occur as knife-sharp boundaries or as fault or shear zones millimeters to several kilometers thick (Hatcher 1990, Twiss & Moores 2000).

The fault zone is composed of two basic structural components: a fault gouge core and a fractured damage zone. The undeformed rock adjacent to the fault is termed intact protolith or host rock (Fig. 4; Sibson 1977, Chester & Logan 1986, Smith et al. 1990, Chester et al. 1993, Bruhn et al. 1994, Forster et al. 1994, Scholz & Anders 1994, Caine et al. 1996, Evans et al. 1997, Seront et al. 1998, Gudmundsson & Homberg 1999, Gudmundsson et al. 2001, 2010, Ganerød et al. 2008). Fault cores in crystalline rocks are commonly narrow and homogeneous zones of fine-grained crushed rock (cataclasite) and clays encased in wider highly fractured and often hydrothermally altered damage zones (Smith et al. 1990, Stewart & Hancock 1994, Hickman et al. 1995, Evans et al. 1997). Boundaries between the damage zone and fault core are typically sharp, whereas the damage zone to protolith transition is usually gradational (Caine et al. 1996).

The intensity of fracturing associated with faulting appears to be a function of lithology, distance from fault plane, amount of displacement along the fault, total strain in the rock mass, depth of burial, and possibly the type of fault (Nelson 2001). Chester and Logan (1986) suggest that there is a variation in properties within the damaged host rock along strike and a significant variation in mechanical properties across the fault zone transverse to strike. According to Palmström (1995), fracture zones can vary in composition from mostly brecciated or crushed material with relatively small amounts of clay to highly weathered or hydrothermally altered, highly plastic, swelling clay gouge.

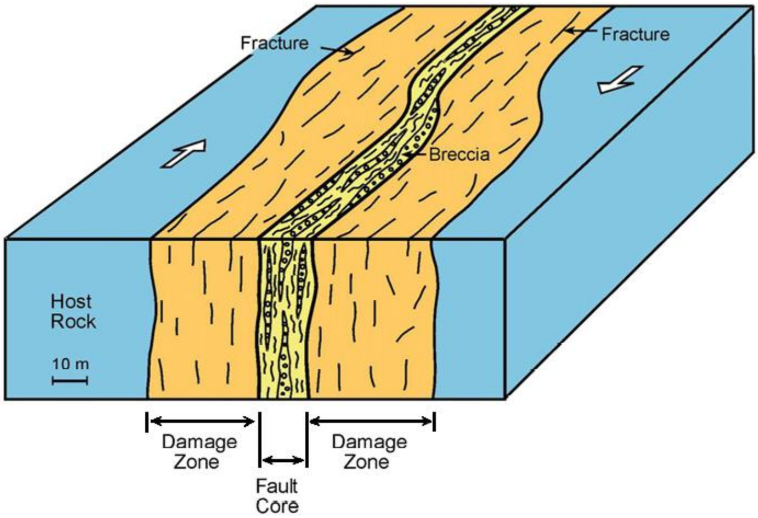

Fig. 4. Structural components of a fault zone (Gudmundsson et al. 2010).

In Norway, Braathen and Gabrielsen (1998, 2000) have found that an area within ca. 250-300 m of any large lineament is affected by enhanced fracturing, and the fracture orientations and frequencies vary as a function of the distance from the lineaments. They divide this 250-300 m wide deformation zone into central, proximal and distal parts. The topographic expression of the lineaments in general coincides with the margin of the proximal part, that is, 70 m from the lineament centre at its widest (Braathen & Gabrielsen 1998, 2000). Palmström (1995) states that many weakness zones do not have a well-defined width, but show a gradual transition from the central part to the surrounding rock masses. Vertical and subvertical fracture lineaments tend to have a symmetrical distribution of their deformation zones (Braathen et al. 1999), whereas inclined faults have accommodated most of the strains in their hanging walls (Gabrielsen et al. 2002).

The term lineament should be strictly reserved for linear or curvilinear features identified by remote sensing methods, and fracture lineament for those lineaments that are assumed to reflect a zone of stress-induced mechanical weakness in the bedrock (Braathen & Gabrielsen 1998, 2000, Gabrielsen et al. 2002). During field investigations, lineaments are commonly identified as zones of enhanced fracture frequency as compared to the surrounding areas. If identified as such by field study, the term fracture swarm could be applied (Bäckblom & Munier 2002, Gabrielsen et al. 2002). When any movement parallel to fracture swarm has proven or disproven, the term fault/fault zone or joint zone should be used. Joint zones differ from fault zones in that they do not show any signs of fault-parallel displacements; hence, they consist dominantly of tensile fractures (Gabrielsen et al. 2002).

3.2 Fault classification

Two end-member fault types exist: dip-slip and strike-slip faults (e.g. Billings 1972, Hatcher 1990, Park 2000). The slip along a fault describes the movement parallel to the fault plane. Movement down or up parallel to the dip direction of the fault is called dip-slip. The term strike-slip applies where movement is parallel to the strike of the fault plane. A combination of these is referred to as oblique-slip (Price 1966, Billings 1972, Davis 1984).

Normal faults are steeply dipping (60º or more) dip-slip faults in which the hanging wall has moved down relative to the footwall (Fig. 5; Price 1966, Davis 1984, Hatcher 1990, Park 2000, Twiss & Moores 2000). Normal faults frequently exhibit listric (concave up) geometry so that they have steep dips near the surface but flatten with depth. Normal faults have been called gravity faults, implying that the primary motive force is gravity. They have also been called extensional faults because they extend the crust (Hatcher 1990).

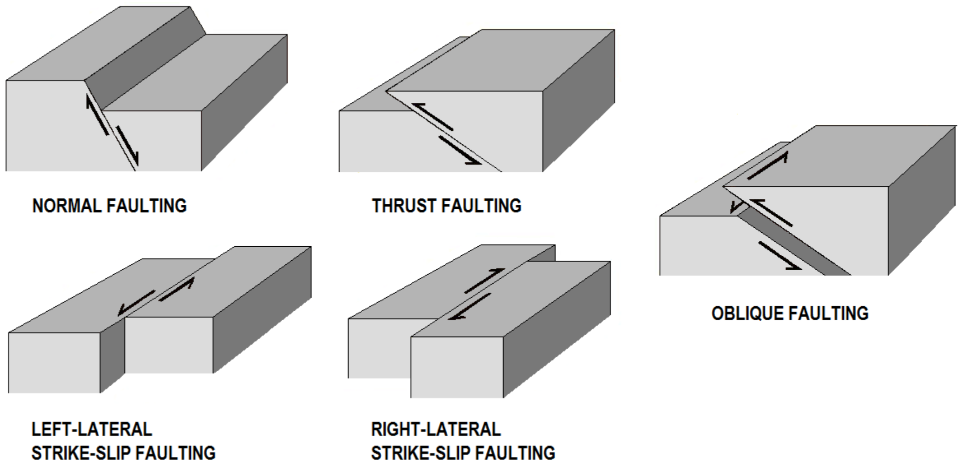

Fig. 5. Different faulting styles.

In thrust and reverse faults the hanging wall has moved up relative to the footwall (Fig. 5). Thrust faults are defined as those with a low angle of dip (30° or less, Hatcher 1990; <45°, Price 1966, Twiss & Moores 2000), and reverse faults as those with a moderate to steep dip (45° or more) and having the same sense of motion as thrusts (Price 1966). Both steeply dipping reverse faults and gently dipping thrust faults are common (Cloos 2009). Moreover, on many thrust faults there exist high- and low-angle segments. Low-angle segments may likewise be found along many reverse faults (Stewart & Hancock 1994). Generally, reverse faulting warps the ground surface to produce a steep slope with an opposite sense of dip to the underlying fault plane, whereas normal faulting mimics the geometry of the underlying faults. Many thrust and normal faults are bedding faults because the fault propagates along a weak zone parallel to bedding (Stewart & Hancock 1994). The term overthrust is used for a thrust fault that dips less than 10º and has a large net slip (Billings 1972).

Strike-slip faults, where the right side moves toward the observer, are called dextral or right-lateral (right–handed) strike-slip faults. Those in which the left side moves relatively toward the observer are called sinistral or left-lateral (left-handed) strike-slip faults (Fig. 5; Anderson 1972, Davis 1984, Hatcher 1990, Park 2000). Strike-slip faults commonly have steep dips – near the vertical – but moderate and shallow segments have also been noted. Wrench faults are high-angle strike-slip faults of great linear extent (Wilcox et al. 1973). Many strike-slip faults are known worldwide as spectacularly active earthquake generators. Thrust faults and strike-slip faults often occur in association. Many faults are actually oblique-slip faults with a dominant thrust, normal, or strike-slip component of motion (Fig. 5; Anderson 1972, Cloos 2009).

Transform faults are a type of strike-slip fault compensating relative motion between lithospheric plates (Billings 1972, Twiss & Moores 2000). One widely known transform is the San Andreas fault, which is a major intracontinental transform (Park 2000). Special kinds of strike-slip faults are tear faults, which bound the edges of thrust sheets (Hatcher 1990). Where strike-slip movement is combined with extension the process is termed transtension (divergent strike-slip); where strike-slip and compression are combined, the result is transpression (convergent strike-slip; Reading 1980, Woodcock & Schubert 1994, Park 2000). Transpressive regimes are marked by thrust faulting, reverse faulting, folding and uplift (Reading 1980). Transcurrent faults are large-displacement strike-slip faults that cut continental basement and the sedimentary cover (Moody & Hill 1956, Reading 1980).

According to Anderson (1972), thrust faults are often quite straight on maps and are seldom much curved in the direction of motion. Normal faults might be expected to map with sinuous outlines whereas strike-slip faults should appear on the map as nearly straight lines. According to Anderson (1972), fault-breccias occur along normal and strike-slip faults, but not in association with thrust-planes. Along thrust-planes there is usually pronounced shearing, leading in extreme cases to the production of mylonite. According to Anderson (1972), normal faulting will increase the transverse horizontal pressure in its neighborhood, while thrust faulting will diminish this pressure.

3.3 Lineament interpretation

According to Waters (1989), the first usage of the term 'lineament' in geology was probably in a paper by Hobbs (1904), who defined lineaments as "significant lines of landscape" caused by joints and faults, revealing the architecture of the rock basement (see also Lattman 1958). O'Leary et al. (1976) proposed a lineament to be a mappable linear feature of a surface, such as a straight stream or ridge that commonly reflects a subsurface structure. Waters et al. (1990) suggested that the term ‘lineament’ is used for any straight, or slightly curved feature or alignment of discontinuous features identified on a map, image or photograph of an area. Clark et al. (1996) defined lineament as a linear pattern, seen on aerial photographs or other remotely sensed imagery that meet established criteria for features that may be the result of underlying zones of fractured bedrock. Braathen and Gabrielsen (1998) and Lie and Gudmundsson (2002) defined lineaments as all mappable linear or curvilinear features that may represent major discontinuities (mechanical breaks) in the bedrock.

Remote sensing is a method of collecting information indirectly from aircraft or satellite-born observation system. In geology, remote sensing is often used for the detection of tectonic structures. Satellite data have been used since the early seventies in regional lineament studies (e.g. Talvitie 1974, Deutsch et al. 1981, Farnsworth et al. 1984, Waters 1989, Wladis 1995, Sander 1996, Kellgren & Sander 1997, Koch & Mather 1997, Anon 2002, Kellgren 2002, Travaglia & Dainelli 2003, Kovalevsky et al. 2004, Hoffmann 2005, Meijerink et al. 2007). Aerial photography has been used for more detailed interpretations (e.g. Brown 1961, Mäkelä 1981, Wise et al. 1985, Waters et al. 1990, Gustafsson 1994). Some lineaments are clearly defined on gravity and aeromagnetic data sets, too, possibly indicating that they extend to deeper crustal levels (van Overmeeren 1981, Kukkonen 1984, Astier & Paterson 1989, Mattson & Salmi 1991, Roberts et al. 1997, Airo 1999, 2005, Murty & Raghavan 2002, Lamontagne et al. 2003, Olesen et al. 2006, Airo & Wennerström 2010).

Although satellite image or air-photo interpretation do not allow to reliably interpret the physical characteristics of fracture zones, such as dip and width, remote sensing is considered to be a very valuable means of investigation. Through remote sensing an overview can be obtained quickly and cheaply for a large area (Waters 1989, Waters et al. 1990, Mabee

1992, Gustafsson 1994, Van Dongen & Woodhouse 1994, Sander et al. 1996, Singhal & Gupta 1999).

In a strict definition, remote sensing does not include map interpretation. However, topographic maps, contoured every 2,5 or 5 m, are based on aerial photographs and they are well suited for lineament interpretation (e.g. Mäkelä 1989b, 1990b, 1990c, Tirén & Beckholmen 1989). The interpretation is based on the idea that the deformation zones in bedrock have eroded more easily than the surrounding intact bedrock. Hence, mapped topographic lineaments (topolineaments) are negative, linear to slightly curved relief features that are represented on topographic maps by surface contour lines, bedrock escarpments, stream segments, longitudinal water courses with depth contours, dry valley channels etc. (Figs. 6 and 7; Rautavuoma 1967, Vähäsarja 1971, Niini 1977, Robinson 2002, Korhonen et al. 2005, Rhén et al. 2007). The advantage of map interpretation is elimination of man-made lineaments (e.g. Waters 1989).

A digital elevation model (DEM) is an ASCII or binary file that contains only spatial elevation data in a regular gridded pattern in raster format (e.g. http://www.terrainmap.com/). These data sets allow the generation of 3D renderings of any location within the area of interest (Fig. 8). The 3D visualization software can be used to enhance landform irregularities by topographic exaggeration and altering illumination angles and direction.

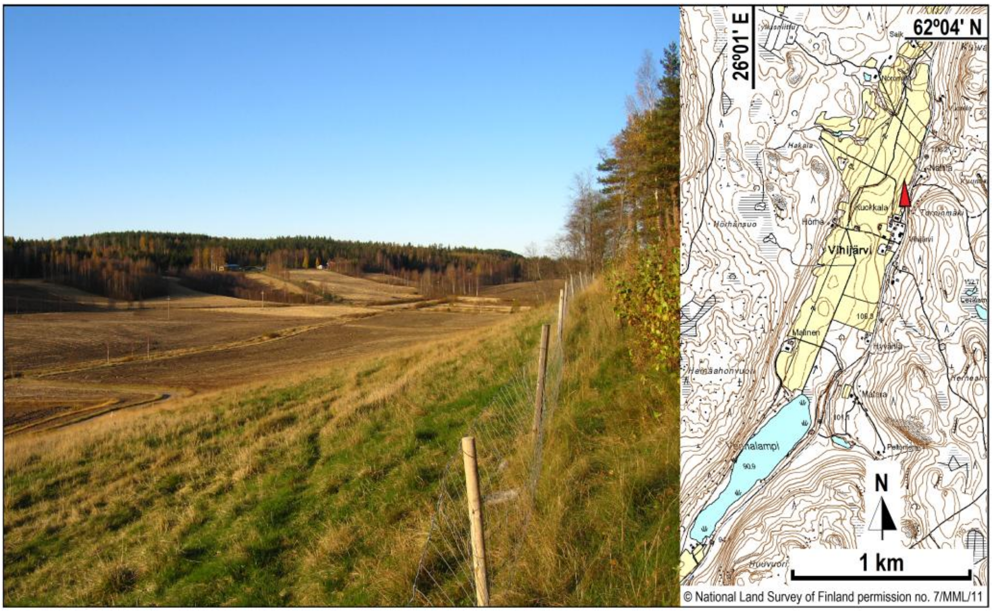

Fig. 6. The NNE-SSW orientated Vihijärvi fault zone in the municipality of Toivakka, Central Finland. Photo shooting place and direction are indicated with a red arrow in the map. The contour lines are marked at 2,5- (dashed lines) and 5-meter intervals. Photo Ossi Alho/CETECF 14/10/2009.

Fig. 7. Topographic map of the Korpilahti area in the municipality of Jyväskylä, Central Finland, depicting the shape of the land surface and showing numerous lineaments of varying trends. The contour lines are marked at 2,5- (dashed lines) and 5-meter intervals. The Korospohja bay is part of Lake Päijänne (+ 79 m a.s.l.). The location of the Oittila water work (test well no. 179330) and that of the Korospohja test well (no. 179324) are indicated with red circles.

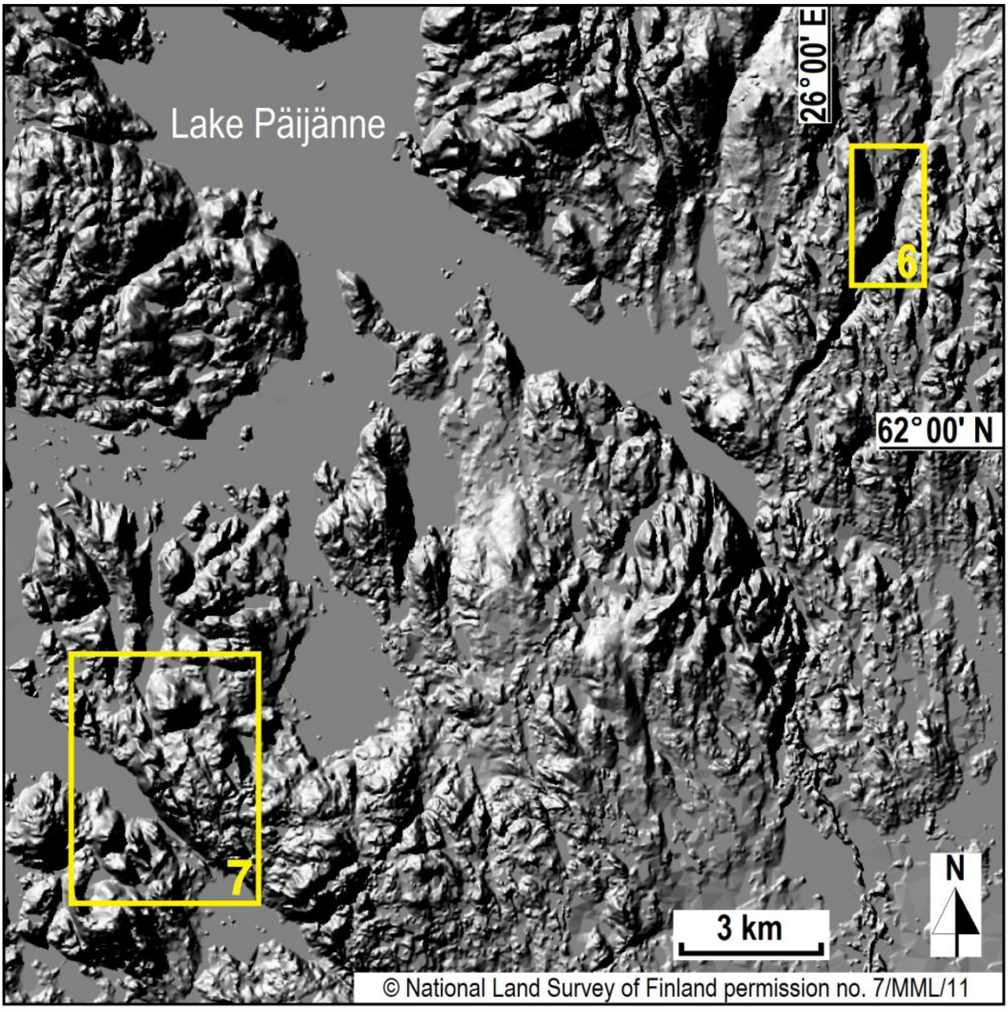

Fig. 8. A shaded relief model in a 25 x 25 m gridded raster format from the southeastern part of Central Finland. The location of the Figures 6 and 7 are outlined in yellow. ArcInfo hillshade grid, illumination angle 45° and direction 315º (NW-SE).

Lineaments related to bedrock features are most easily observed in areas where the surficial material is thin. Problems encountered during interpretation are typically wide valleys or smooth areas with thick soil cover, where it is difficult to position lineaments with satisfactory certainty (Gustafsson 1994, Tam et al. 2004). However, obscured and buried structures may be reflected at the surface by structurally controlled rivers and appear as lineaments (Edet et al. 1998). In glaciated terrain covered with soil certain lineament azimuths may be more pronounced than others. Superficial deposits tend to fill in negative relief features perpendicular to former ice sheet flow directions and etch out features parallel to them, introducing bias in the lineament map (Gustafsson 1994, Mabee at al. 1994, Wladis 1995). While certain true fracture zones can be obscured by the direction of the ice flow, more often the problem lies in interpreting long valleys running in the direction of ice flow as fracture zones.

Lineaments are mainly restricted to steeply dipping fractures or fracture zones that produce straight-line intersections with the earth´s surface (Clark et al. 1996, Moore et al.

2002a). Some lineaments are curved and may be shallow-dipping faults or consist of two or more structures with different trends (Lie & Gudmundsson 2002). Gently dipping fracture zone when intersecting ground surface may also appear as a straight valley and could be interpreted wrongly as vertical or sub-vertical fracture zone (Jääskeläinen et al. 2007). Although gently dipping structures are difficult to recognize with remote sensing techniques, there is an acute need to recognize such zones from the surface because of their importance, for instance, in controlling bedrock stability, rock stress regimes and the lateral flow of groundwater (Greenbaum 1990c, Ahlbom & Smellie 1991, Olofsson 1994, Morland 1997).

Mapped lineaments are most commonly classified based on a geometric trait such as linearity, density, or length, or on a measure of relative prominence (Brown 1994, Sander 1996, 1999, Sander et al. 1996, Koch & Mather 1997). Lineament distributions are often displayed in various types of azimuth-frequency plots (rose diagrams; Brown 1994). Lineament distribution is not homogeneous from place to place, and is dependent upon lithological, geomorphological, and tectonic features of the site under investigation (Tam et al. 2004). Tsuchida et al. (1990) have found that the longer the segments of a lineament, the higher the probability to be extracted by visual interpretation of remote sensing images.

Concurrently one must remember, too, that all interpretations are based on two-dimensional representations of a three-dimensional geological architecture (Brown 1994). The rosegrams can provide important information about surface lineament patterns, but, unfortunately, have little meaning for subsurface structural orientations unless the geological structures represented by the lineaments are vertical to subvertical. This is the main limiting assumption of lineament interpretation in most geological environments. However, combined structural analysis and lineament interpretation may provide a method for describing aquifers, which are important for groundwater exploration in crystalline bedrock areas (Brown 1994). According to Tirén and Beckholmen (1989), rock block maps give an additional (third) dimension to the information of structures as compared to lineament maps.

The actual interpretation of lineaments is a subjective process (Vuorela 1982, Wise 1982, Gustafsson 1994, Sander 1999, Kellgren 2002, Tam et al. 2004). A greater attention thus must be paid to testing reproducibility and reliability than is customary in traditional geologic research (Wise et al. 1985, Mabee et al. 1994). Suitability of the area for lineament interpretation, good data and skilful geological interpretation are vital components of any useful lineament study. The experience of the interpreter is very important (Peacock & Mann 2005). Field reconnaissance and ground truthing should be carried out to gain an adequate knowledge of the area interpreted on maps sheets. Mabee et al. (1994) have summarized the lineament-data-collection technique and describe techniques for filtering the data to produce a subset of lineaments that are more likely related to transmissive zones in bedrock.

3.4 Lineaments in Fennoscandia

The Fennoscandian Shield is known to consist of a mosaic of bedrock blocks separated by numerous lineaments of different sizes (from hundreds of meters to hundreds of kilometers) occupying topographical depressions. These lineaments are commonly surface expressions of Precambrian shear zones and faults, which have been formed because the physical and chemical processes of erosion and the Pleistocene continental ice sheets have acted most effectively on weakened fracture zones (Sederholm 1913, Marmo 1959, Härme 1961, 1966, Paarma & Marmo 1961, Paarma 1963, Larsson 1968, 1977, Teisseyre et al. 1968, Petrov 1970, Tuominen et al. 1973, Strömberg 1976, Simonen 1980, Koskiahde et al. 1985, Kuivamäki & Tuominen 1985, Saari 1992, Vuorela & Paulamäki 1992, Kuivamäki & Vuorela 1994, 2002, Vuorela et al. 1996, Kuivamäki et al. 1998, Ojala et al. 2004, 2006).

Several studies of lineament and rock blocks have been carried out in Fennoscandia using different types of remote sensing and map interpretation (Fig. 9; Talvitie 1971, 1975, 1977a, 1977b, 1978, 1979, Vuorela 1974, 1982, Aarnisalo 1977, 1978, Mikkola & Vuorela 1977, Röshoff & Lagerlund 1977, Röshoff 1979, 1988, Vuorela & Hakkarainen 1982, Vuorela & Niini 1982, Kuivamäki 1985, Salmi et al. 1985, Berthelsen & Marker 1986, Tirén & Beckholmen 1989, 1990, 1992, Tirén 1991, Tossavainen 1992, Veriö et al. 1993, Braathen & Gabrielsen 2000, Dehls et al. 2000a, Gabrielsen et al. 2002, Henriksen & Braathen 2006, Beckholmen & Tirén 2009). The lineaments of Central Finland are discussed more thoroughly in Chapter 5.2.3.

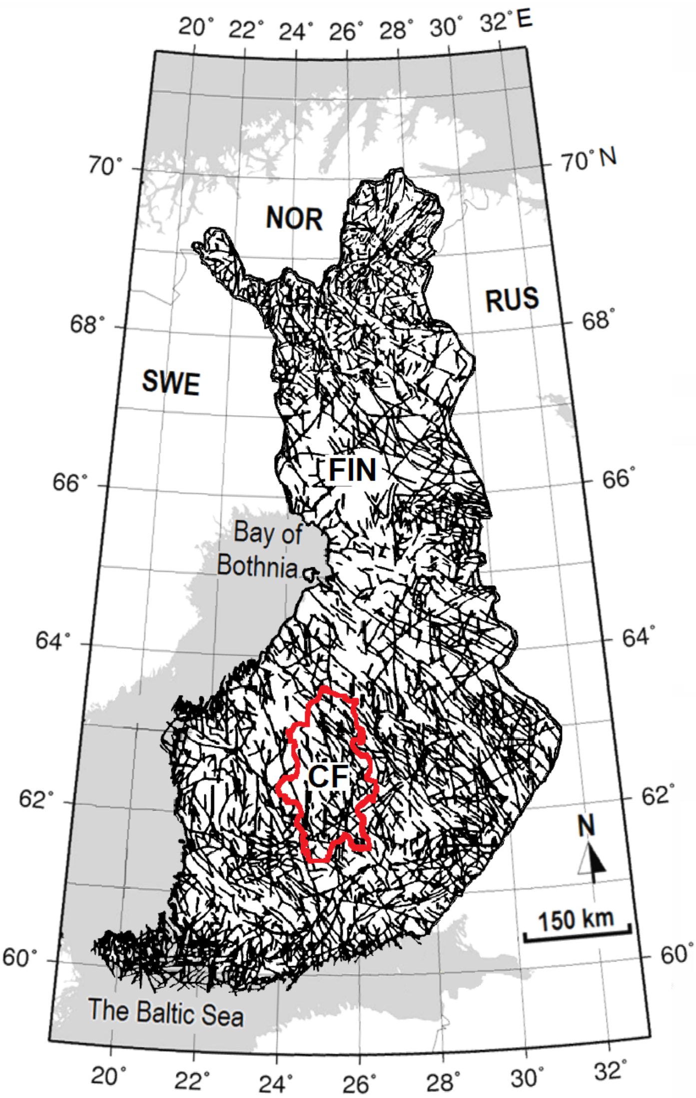

Fig. 9. Lineament map of Finland. SWE=Sweden, NOR=Norway, FIN=Finland, RUS=Russia. Central Finland (CF) is outlined in red. Modified from Vuorela and Niini (1982), base map from Saaranen (2005).

Deep and long tectonic lineaments running in a NW direction appear to traverse the Fennoscandian Shield at regular intervals 50-80 km apart each other (Talvitie 1975, 1979). However, extensive NW-lineaments are missing in the mainland of southern Sweden (Tirén & Beckholmen 1992). In Finland, lineaments trending NW-SE and approaching the direction of the glacial drift, are continuous and sharp (Vuorela 1982). There are four main deformation zones striking right-handed NW-SE, namely (1) the Salla-Kittilä zone in Lapland, (2) the Ladoga-Bothnian Bay zone in central Finland, (3) the Central Finland zone and (4) the Loviisa-Pori zone in southern Finland (Mikkola & Vuorela 1977, Talvitie 1978, 1979). According to Talvitie (1977b), the Ladoga-Bothnian Bay fracture zone can be regarded principally as the result of the wrench faulting of two dynamic stages in the directions of 300-305º and 325-330º. Some of the NW-directed tectonic zones may have constituted boundaries for younger orogenies, with thrusts and deformations against a cratonic margin (Strömberg 1976, Talvitie 1979). The large faults tend to cut across different rock type areas.

Another important lineament direction is NE, which is represented by some very long faults and shear zones. Together with NW-lineaments, these represent an old megafracture pattern that in the eastern parts of the shield may date as far back as the Archean (Strömberg 1976, Mikkola & Vuorela 1977, Gabrielsen et al. 2002). From eight to ten strike frequency maxima have been identified in Finland, and schistosity, fractures and gravity anomalies all indicate statistically similar orientations (Tuominen et al. 1973, Aarnisalo 1977, Talvitie 1977a, Vuorela 1982, 1992, Vuorela & Hakkarainen 1982, Kuivamäki 1985).

In Finland, bedrock fracture zones have been divided into four classes of different size categories (Salmi et al. 1985, Veriö et al. 1993): I – width 1 km, length from tens to hundreds of kilometers, II – width hundreds of meters, length from five to tens of kilometers, III – width from tens to one hundred meters, length less than five kilometers, IV – minor fracture zones. In Norway, the length of the first order lineament is 50-100 km, the length of the second order lineament is 5-50 km and the length of the third order lineament is 1-5 km (Braathen & Gabrielsen 2000).

3.5 Lineaments and groundwater

Lineament studies have developed into one of the primary techniques for well siting in crystalline rocks, where groundwater occurrence is primarily a function of a fracture-induced secondary permeability. Lineaments are interpreted to represent surface manifestations of enhanced permeability zones in the bedrock (e.g. Barton et al. 1995a, Caine et al. 1996, Evans et al. 1997, Braathen et al. 1999, Solomon & Quiel 2006).

The technique of mapping lineaments from topographic maps, aerial photographs and satellite data for locating zones of higher permeability in hard rock terrains has been used in groundwater exploration studies from the early sixties. Although the use of lineaments to locate bedrock wells was known for a long time, Lattman and Parizek (1964) first successfully mapped lineaments ('fracture traces') in connection with groundwater exploration and found a positive correlation between them and high-yield wells in a carbonate rock terrain in the United States. Since then a great number of articles have been published on the subject (e.g. Setzer 1966, Rautavuoma 1967, Poth 1968, Siddiqui & Parizek 1971, Hanson 1972, Woodruff et al. 1974, Moore & Hollyday 1977, LeGrand 1979, Greeman 1981, Krothe & Bergeron 1981, Schowengerdt et al. 1981, Zall & Russell 1981,Welby & Wilson 1982, Kogbe 1983, Buckley & Zeil 1984, Jammallo 1984, Brook 1985, Odeyemi et al. 1985, Waters 1989, Greenbaum 1990a, 1992, Waters et al. 1990, Clarke & McFadden 1991, Zewe & Rauch 1991, Banks et al. 1992b, 1993, Boeckh 1992, Henry 1992, Mabee 1992, 1999, Yin & Brook 1992a, Tossavainen 1992, Briz-Kishore 1993, Gustafsson 1993, 1994, Helvey & Rauch 1993, Lonka et al. 1993, Brown 1994, Kaehler & Hsieh 1994, Kastrinos & Wilkinson 1994, Mabee et al. 1994, 2002, Wladis & Rosenbaum 1994, 1995, Chesley et al. 1995, Hardcastle 1995, Henriksen 1995, Kresic 1995, Loiselle & Evans 1995, Rosenbaum & Wladis 1995, Teeuw 1995, Edet 1996, Mahmood 1996, Sami 1996, Sander 1996, 1997, 1999, Sander et al. 1996, 1997, Stibitz & Fleischmann 1996, Edet & Okereke 1997, Koch & Mather 1997, Morland 1997, Edet et al. 1998, Braathen et al. 1999, Magowe & Carr 1999, Mälkki 1999, Singhal & Gupta 1999, Elfouly 2000, Kellgren & Sander 2000, Krishnamurthy et al. 2000, Moore et al. 2001, 2002a, 2002b, Naik et al. 2001, Degnan & Clark 2002, Degnan & Moore 2002, Degnan et al. 2002, Dinger et al. 2002, Kellgren 2002, Lie & Gudmundsson 2002, Robinson 2002, Domoney et al. 2003, Owen et al. 2003, Tam et al. 2004, Johansson 2005, Penner & Mollard 2005, Chandra et al. 2006, Kenny et al. 2006, Masoud & Koike 2006, Meijerink et al. 2007, Neves & Morales 2007a, 2007b, Mukherjee 2008, Ranganai & Ebinger 2008, Linn 2009, Jia & Lin 2010, Holland & Witthüser 2011, Jasmin & Mallikarjuna 2011, Tessema et al. 2012).

The central part of a lineament, the fault core, has usually a high fracture connectivity and density (e.g. Braathen & Gabrielsen 1998). In spite of this, it often represents a low permeability zone due to heavy weathering (e.g. clay gouge) and the presence of fault rocks such as breccias, as well as the common presence of secondary mineral fillings, which tend to clog or seal fractures (Cederstrom 1972, Sibson 1977, Chester & Logan 1986, Forster & Evans 1991, Banks et al. 1992a, 1992b, 1994, Scholz & Anders 1994, Chilton & Foster 1995, Caine et al. 1996, Forster et al. 1997, Sander 1997, Sander et al. 1997, Braathen & Gabrielsen 1998, Olesen et al. 2006, Ganerød et al. 2007, 2008, Mattila et al. 2008). Field and laboratory measurements generally show a very low core hydraulic conductivity (K ~10^{-10}...10^{-13} ms^{-1}; Evans et al. 1997, Seront et al. 1998, Braathen et al. 1999).

The damage (proximal) zone has longer individual fractures of diverse orientations and also higher fracture frequencies than the fault core, resulting in a fracture network of high connectivity (e.g. Odling 1997, Braathen et al. 1999). Secondary mineral fillings in fractures are in general absent, making the proximal part of the lineament a zone of potential enhanced permeability (e.g. Caine et al. 1996).

According to Smith et al. (1990), the permeability of the fault zone reflects the interplay of four elements: pods of undeformed protolith, open damage zones, damage zones partially or fully sealed by mineral precipitates, and fault core gouge zones. Lineament-perpendicular conductivity in fracture zones can be regarded as potential low in the fault core and high in the damage zone, whereas the lineament-parallel conductivity is more uniform along the fracture zone (Braathen & Gabrielsen 1998, Braathen et al. 1999, Henriksen & Braathen 2006). Fracture zones with good hydrodynamic properties often show a strong anisotropy, e.g. lateral and longitudinal differences in drawdown (Dewandel et al. 2005). Several geologic factors can account for anisotropy, e.g., the lithology, the fracture pattern, and the current stress field (Wladis & Gustafson 1999, Baghbanan & Jing 2008).

The overall permeability of a single fault may range over six orders of magnitude (Forster et al. 1994, Evans et al. 1997). This permeability variability may result in the hydrogeologic behavior of a lineament as a conduit, a barrier, or a combined conduit-barrier to groundwater flow (Scholz & Anders 1994, Caine et al. 1996, Evans et al. 1997, Zhang et al. 1999).

Some dikes, particularly thick and massive ones, are barriers to transverse flow of groundwater (Caine et al. 1996, Davis & DeWiest 1966, Evans et al. 1997, Gudmundsson 2000a, 2001, Babiker & Gudmundsson 2004). This follows because the dike rock is commonly of low matrix permeability (Gudmundsson et al. 2003). However, fractured dikes may act as sources of groundwater (Singhal & Gupta 1999). Some dikes are filled with secondary minerals, in which case even thin dikes may be barriers to transverse groundwater flow (Gudmundsson et al. 2003). Because the dike rock has commonly very different mechanical properties from that of the host rock, stresses tend to concentrate at the contacts between dikes and host rocks and generate groundwater conduits. Thus, dikes, even those that are barriers to transverse flow, are commonly good conduits of dike-parallel flow of groundwater (Gudmundsson et al. 2003).

Prominent large-scale lineaments are associated with transmissive zones that may extend as much as a few hundred meters perpendicular to the strike of the feature (Chesley et al. 1995, Teeuw 1995, Sander et al.1996). Wladis and Rosenbaum (1994) suggest that the favorable zone of influence around a lineament extends to a distance of about 500 meters; any correlation of yield and proximity to a lineament vanishes beyond distances of 200-300 meters. However, the groundwater potential does not always increase when approaching lineaments (Gustafsson 1994). To the contrary, Lie and Gudmundsson (2002) and Henriksen (2006a, 2006b) have found that groundwater flow rates in the innermost zone of the fault decrease towards the centre.

In the region of high lineament frequencies and densities the exploitation potential for groundwater is most often considered high. For example, in the Precambrian basement rocks of Sudan, Ahmed et al. (1984) have indicated groundwater target zones by the overlap of the high-intensity lineament contours and the low-intensity drainage contours, and by intersection of a stream channel and lineament structure. Most wells lying within the defined targets have proven to be successful.

Lineaments of a particular orientation may also be more potential than another (Greenbaum 1992, Kellgren & Sander 1997, Sander 1999, Lie & Gudmundsson 2002). For example, Bromley et al. (1994) observed that WNW trending lineaments deciphered from EM-VLF surveys gave the best results when identifying such lineaments in the buried Karoo sandstones in Zimbabwe. According to Tirén and Beckholmen (1989), lineaments between rock blocks of different elevation are potential aquifers for water supply.

Groundwater exploration programs based on lineament analysis commonly place particular importance on lineament intersections. The intersections form the locked area of a fault and are the locations of stress build-up, which can generate major earthquakes even in intraplate conditions (Slunga 1989c, Talwani 1989, 1999, Ojala et al. 2006). Indeed, many drilling results support the notion that these intersections, especially at the low altitude, promote favorable conditions for groundwater extraction (e.g. Meier & Sund 1952, Woodruff et al. 1974, Greeman 1981, Krothe & Bergeron 1981, Buckley & Zeil 1984, Larsson et al. 1984, Talwani 1989, Teme & Oni 1991, Boeckh 1992, Struckmeier 1993, Brown 1994, Magowe & Carr 1999, Elfouly 2000, Naik et al. 2001). Their important role in hydrocarbon systems has also been documented (Gartrell et al. 2003, 2004, Ligtenberg 2004, 2005).

On the other hand, many low-productive wells locating on or close to fracture lineaments demonstrate that lineaments alone cannot be used for water-well siting (Cederstrom 1972, Dijon 1990, Banks et al. 1993, Greenbaum et al. 1993, Filho & Rebouças 1995, Kellgren & Sander 1997, Sander 1997, 1999, Sander et al. 1997, Edet et al. 1998, Magove & Carr 1999, Park et al. 2000, Mabee et al. 2002, Chandra et al. 2006, Galanos & Rokos 2006, Boutaleb et al. 2008).

Norwegian experiences have tended to indicate that, even in prominent lineament zones in granitic lithologies with promising borehole statistics, one cannot guarantee a high yield (Banks et al. 1994, see also Kaehler & Hsieh 1994). Fracture lineaments may just as well coincide with zones of reduced permeability (Banks et al. 1992b, Gustafsson 1994) or even with barriers to groundwater flow (Gleeson & Novakowski 2009). At its worst, best well sites may be far away from lineaments, or lineaments should be avoided, or there is really nothing 'linear' available in the area where it is needed (Sander 2007). High yielding wells may also be drilled in areas where thick deposits of unconsolidated material obscure underlying transmissive features, which consequently go undetected in the lineament mapping

(Chilton & Smith-Carington 1984, Greenbaum et al. 1993, Kellgren & Sander 1997, Sander et al. 1997).

The occurrence of gently dipping fracture zones in crystalline rock has generally been underestimated (e.g. Tirén 1991). However, from a hydrogeological point of view, subhorizontal fracture zones may be important even though they might be difficult to recognize with remote sensing techniques (e.g. Morland 1997). In the majority of crystalline rock areas worldwide where subsurface investigations for radioactive waste disposal have been performed, the occurrence of horizontal to gently dipping fracture zones has been demonstrated (Andersson 1993, Hsieh 1996). According to Olofsson (1994), gently dipping fracture zones, formed as shear features, can exist down to considerable depths.

Hence, for different reasons, lineaments in the terrain may be highly variable in their hydrogeological significance (Banks et al. 1994, Sander et al. 1997, Banks & Robins 2002) and other factors than lineament characteristics may also be important from the water supply point of view.

4 SEISMOTECTONICS

4.1 Rock stress

There has been a great deal of interest in mapping the magnitude and orientation of in situ stress in the upper crust across the globe over the last 40 years (e.g. Hast 1969, 1974, Sbar & Sykes 1973, Ranalli & Chandler 1975, McGarr & Gay 1978, Richardson et al. 1979, Stephansson 1986, Zoback et al. 1989, Müller et al. 1992, Zoback 1992a, Reinecker et al. 2005, Anon 2008a).

Knowledge of the in situ state of stress in the Earth´s crust is very important in many problems dealing with rocks in civil, mining and petroleum engineering and energy development, as well as in geology and geophysics. It has improved, for example, our understanding of stability problems in underground excavations, of fluid flow and geothermal management and that of crustal deformation processes (e.g. earthquakes) and the influence of plate tectonics on these processes (Stephansson et al. 1987, Engelder 1993, Coblentz et al. 1995, Amadei & Stephansson 1997, Fuchs & Müller 2001, White & Hillis 2004, Heidbach 2009).

4.1.1 Stress concepts and definitions

Stress in rock is usually described within the context of continuum mechanics. It is defined at a point and is represented by a second-order Cartesian tensor with three normal stress and three shear stress components (Price 1966, Jaeger & Cook 1979, Ranalli 1995, Amadei & Stephansson 1997, Lund 2000, Zoback 2011). The magnitude of stress (σ) is a function of the force (F) and the area (A) on which the force acts (Davis 1984, Park 2000, Twiss & Moores 2000)

$$\sigma = F/A \qquad (5)$$

The S.I. unit for measuring stress is the Pascal (Pa). The megapascal (MPa), i.e. 10^6 Pa, is usually used due to the large stresses involved in geological processes (e.g. Reynolds 2001).

Stresses in rock cannot be measured directly and can only be inferred by disturbing the rock. Furthermore, rock stresses cannot be determined accurately due to the complex nature of rocks and rock masses. Different stress measurement techniques give on average comparable stress values for same rock volumes within the uncertainty expected in stress measurements, i.e. an error of ±10-20% for the stress magnitude and an error of ±10-20% for the stress orientation (Amadei & Stephansson 1997).

Stresses in rock can be divided into in situ stresses and induced stresses. In situ stresses, also called natural, primitive or virgin stresses, are the stresses that exist in the rock prior to any disturbance. On the other hand, induced stresses are associated with artificial disturbance (excavation, drilling, pumping, loading etc.) or are induced by changes in natural conditions (drying, swelling, consolidation, etc.; Amadei & Stephansson 1997). Residual (locked) stresses are usually defined as stresses which exist in the Earth´s crust as residues from former tectonic deformations to the contrary of tectonic stresses which are a result of recent tectonics. However, if residual stresses are present they have to be very small compared with the applied tectonic stresses (Greiner & Illies 1977). According to Zoback (1992a), residual stresses from past orogenic events do not appear to contribute in any substantial way to the modern stress field. Reconstruction of the stress regime responsible for a particular fracture set may indicate its original tensile or compressional nature, but this will not necessarily represent its present-day open or closed character, which will depend also on its history of reactivation and the current stress field (Greenbaum 1990b).

Stress tends to deform the rock mass. The amount of deformation of a rock body is expressed by strain. Strain is the change in size and shape of the body resulting from the action of an applied stress field (Park 2000). The presence of continuing seismicity is an important indicator of accumulating strain in the crust and upper mantle (Lambert & Vaníček 1979). Failure of material takes place when permanent deformation occurs. Material fails when stress reaches a certain critical value called the yield strength or the yield stress. Failure can occur as discontinuous deformation, i.e. fracturing, or continuous irrecoverable deformation, i.e. plastic flow (Ranalli 1995). Material that fails by fracturing is called brittle and material that fails by a plastic flow is termed ductile. Earthquakes, faulting and folding are thought to be examples of fracturing and plastic flow, which thus have a major role in geodynamics, especially in the upper lithosphere (Moisio 2005).

The state of stress can be represented by three principal stresses and their orientation in the x, y, z coordinate system. By definition, any stress acting perpendicular to a surface along which the shear stress is zero is a principal stress (Price 1966). By convention the z-axis is usually taken as vertical. Other principal stress axes may be orientated parallel to ground surface and perpendicular to each other and would be designated as x and y. If the relative intensities of the principal stresses are known, they may be termed the maximum (or greatest), intermediate and minimum (or least) principal stresses, i.e. σ_1, σ_2 and σ_3, respectively (compressive stresses are considered positive). In general, all principal stress magnitudes are shown to increase with depth (Price 1966, Anderson 1972, Amadei & Stephansson 1997, Park 2000). The present-day horizontal tectonic stresses in the upper crust (at < 1 km depth) equal 10-40 MPa (Fuchs & Müller 2001).

Shear fractures form when all three principal stresses are compressive. Shear fractures form at some acute angle to σ_1. The acute angle between shear fractures is called the conjugate angle (Nelson 2001). To form a tension fracture, at least one principal stress must be tensile, unless fluid pressure is high enough. Studies by Secor (1965) indicate that joints form more readily in the presence of fluid, generating fractures by hydraulic fracturing. The majority of fractures developed in the vicinity of faults are shear fractures parallel to the fault, shear fractures conjugate to the fault, or extension fractures bisecting the acute angle between these two shear directions (Nelson 2001).

The fault classification devised by Anderson (1905, 1972) defines three fundamental categories based on mutual relationships between fault planes and principal stresses. In thrust faults, the maximum principal stress σ_1 and the intermediate principal stress σ_2 are horizontal and the minimum principal stress σ_3 is vertical. A state of horizontal compression is thus defined for thrust faulting. In strike-slip faults, maximum and minimum principal stresses σ_1 and σ_3 are horizontal and the intermediate stress σ_2 is vertical. In normal faults, the maximum principal stress σ_1 is vertical and σ_2 and σ_3 are horizontal. It should be noted that the resulting shear planes are compressive features in every stress regime. This is true even in the case of normal faulting despite the fact that this is usually associated with lateral extension in one horizontal direction (Greenbaum 1990a).

For the standard Andersonian stress states in intact isotropic rocks, the shear fractures should form at acute angles (typically between 20º and 35º) to the orientation of σ_1 such that the fault plane contains the orientation of σ_2. In thrust faults commonly only one dominant shear plane becomes a fault, while in normal faults both fault planes frequently develop (Anderson 1905, 1972, Cox et al. 2001, Sibson 2001).

In that what follows, the maximum and minimum horizontal stress and the vertical stress are designated as S_H, S_h and S_V, respectively. Although S_H, S_h and S_V are treated as principal stresses, they have different relative magnitudes depending on the geological setting and, therefore, have no a priori relationship to the principal compressive stresses designated as $\sigma_1>\sigma_2>\sigma_3$ (Engelder 1993).

4.1.2 Stress origin

Two primary categories of forces are responsible for the state of stress in the upper, elastic part of the Earth's lithosphere. These are tectonic (continental) stresses (e.g. plate-boundary forces) and local (second-order) stresses (e.g. effects of geological structures, glacial rebound and topography; Klein & Barr 1986, Hickman 1991, Engelder 1993, Miller & Dunne 1996, Fejerskov & Lindholm 2000). Tectonic stresses typically are uniform over distances many times the thickness of the elastic part of the lithosphere, whereas local stresses have wavelengths that are a fraction of this thickness (Zoback et al. 1989).

The tectonic stress fields are believed to be largely the result of compressional forces applied at plate boundaries, primarily ridge push and continental collision but also slab pull and trench suction (Husebye et al. 1978, Mendiguren & Richter 1978, Richardson et al. 1979, Klein & Barr 1986, Skordas 1992, Zoback 1992a, Engelder 1993, Reynolds et al. 2003). For instance, the Mid Atlantic Ridge push force resulting from the excess thermal elevation of young oceanic lithosphere is an integrated tectonic force acting over the cooling oceanic lithosphere and generally acts normal to the plate boundary, except where spreading is oblique to the ridge (Zoback et al. 1989). Sea-floor spreading between Greenland and Europe commenced in the Early Tertiary about 54 Ma ago (Muir Wood 1995, Torsvik & Cocks 2005). Since the crust cools down and becomes denser as it moves away from the ridge, gravitational sliding occurs and causes high stresses in the direction of increasing plate age and thickness (e.g. Sykes 1978, Grollimund & Zoback 2000). Fejerskov and Lindholm (2000) have proposed ridge push stresses in order of 10-30 MPa. Slab pull and trench suction are not relevant in Fennoscandia (Husebye et al. 1978).

Plate boundary forces are also responsible for the plate motions (e.g. Engelder 1993). Thus, there is a coincidence between plate motion trajectories and stress orientations in most continents (Zoback et al. 1989, Müller et al. 1992, Spassov 1998, Heidbach et al. 2010). The orientation of the intraplate stress field is largely controlled by the geometry of the plate boundaries. In areas with large differential stress, such as western Europe or northern America, the orientation of the maximum horizontal stress is most likely to be fairly consistent and parallel to regional trends (Sbar & Sykes 1973, Engelder 1982, Zoback 1992a, 1992b). However, large-scale faulting, which separates areas of the field into distinct fault blocks, can locally rotate the regional stress significantly (Heidbach et al. 2008a, 2010). Local variations in magnitude and orientation of rock stresses occur, for example, around folds, steep valley sides and with depth, too (Miller & Dunne 1996, Su & Stephansson 1999, Ameen 2003, Yale 2003, Peacock & Mann 2005). In Australia, a number of locally anomalous stress orientations appear influenced by second-order sources of stress such as structure, topography and density heterogeneities (Hillis & Reynolds 2003).

4.1.3 Mohr stress diagram

The stress states in rocks and the relationships between stress magnitudes, stress differences, fracture type, and fracture orientation can be illustrated by the Mohr stress diagram (Fig. 10; e.g. Jaeger & Cook 1979, Ranalli 1995, Twiss & Moores 2000, Cox et al. 2001). In the diagram, the normal stress (σ_n) on the horizontal axis is plotted against the shear stress (τ_s) on the vertical axis. The Mohr circle has its center on the horizontal axis; its intersections with this axis define the maximum and minimum principal stresses σ_1 and σ_3, respectively. Then the center of the Mohr circle is at $(\sigma_1 + \sigma_3)/2$ and its diameter is $(\sigma_1 - \sigma_3)$ and the radius $(\sigma_1 - \sigma_3)/2$. The normal and shear stresses acting on a plane inclined at an angle θ to σ_1 are given by the coordinates of the point P, which is situated on a radius inclined at 2θ to the horizontal normal axis (Fig. 10).

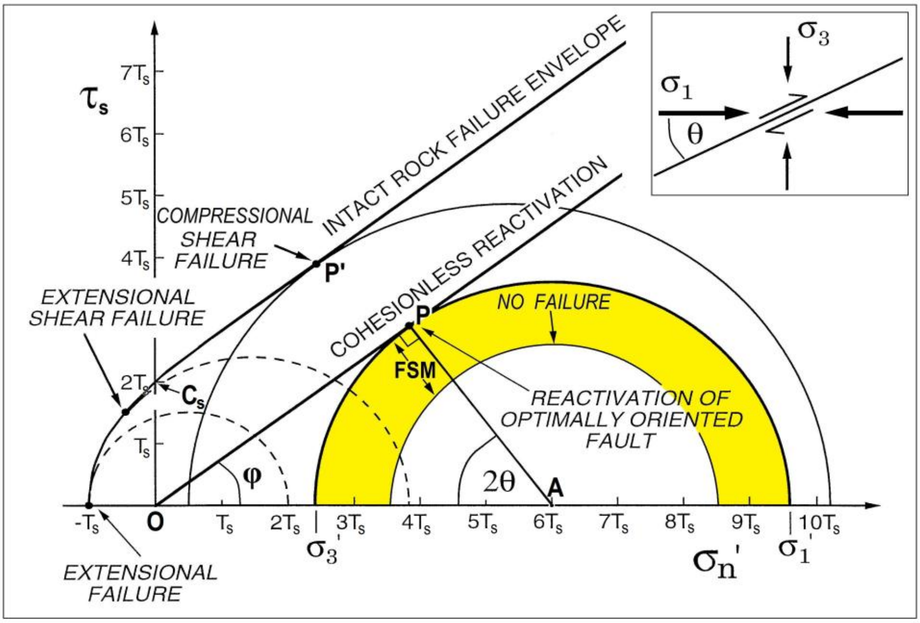

Fig. 10. Mohr stress diagram with composite failure envelopes for intact isotropic rock with tensile strength, T_s, and for frictional reactivation of an existing cohesionless fault, illustrating the stress conditions and orientations with respect to the stress field of extensional failure, hybrid extensional-shear failure, compressional shear failure and reactivation of an optimally oriented existing fault. Modified from Sibson and Scott (1998).

The magnitudes of the normal stress and shear stress are given by the relationships

$$\sigma_n = 1/2(\sigma_1 + \sigma_3) + 1/2(\sigma_1 - \sigma_3)\cos 2\theta \qquad (6)$$

and

$$\tau_s = 1/2(\sigma_1 - \sigma_3)\sin 2\theta \qquad (7)$$

where the $(\sigma_1 - \sigma_3)$ is the differential stress (the diameter of the Mohr circle; Sibson 2001). Maximum shear stress is defined as

$$\tau_{max} = 1/2(\sigma_1 - \sigma_3) \qquad (8)$$

and it lies along planes containing the σ_2 axis oriented at ±45° to the σ_1 direction (Sibson 2001).

With a rock subject to a high deviatoric stress one means that there is a large difference between a particular component of the stress tensor, for instance σ_1, and the mean stress (Engelder 1993), which is defined as

$$\sigma_m = (\sigma_1 + \sigma_2 + \sigma_3)/3 \tag{9}$$

Zoback et al. (1993) have estimated that the maximum differential stress ($\sigma_1 - \sigma_3$) reaches values of ~300-400 MPa in the mid-crust.

The line OP in Figure 10 indicates the Mohr-Coulomb shear failure criterion. This line is called the failure envelope and it specifies the brittle shear failure strength of the medium as a function of normal stress. The shortest perpendicular distance from the failure envelope OP to the Mohr circle is defined as the fault stability margin (FSM; e.g. Johnston 1987, 1989a). So, brittle shear failure (i.e. fault reactivation) occurs where and when the Mohr circle contacts the failure envelope OP, i.e. when the FSM becomes to zero. Decreasing σ_3 and/or increasing the value of σ_1 may, therefore, induce shear failure. That is, changes in the deviatoric or differential stresses or changes in the mean stress will move the Mohr circle (e.g. Sauber & Molnia 2004, Lund 2005a). Stress states, which exceed the failure line, are not allowed because failure of the rock would have occurred prior to the rock having achieved such a stress state (Zoback 2011).

Under fluid-absent conditions, the stress state that causes compressional brittle shear failure in an intact, isotropic medium is approximated by the relationship

$$\tau_s = C_s + \mu_s\sigma_n \tag{10}$$

where C_s is the cohesive strength of the medium (point where failure envelope intersects y-axis) and μ_s is the static coefficient of friction (tangent of slope of failure envelope, i.e. $\tan(\varphi)$), which commonly lies in the range $0{,}60 \leq \mu_s \leq 1{,}00$ (Byerlee 1968, 1978, Zoback 2011).

The coefficient of friction is found by dividing the shear stress required to cause sliding by the normal stress across the surfaces (Price 1966, Sibson 2004). The slip tendency, T_t, of a surface is the ratio of shear stress to normal stress acting on that surface (Morris et al. 1996)

$$T_t = \tau_s/\sigma_n \tag{11}$$

Because the angle between the Mohr circle radius (AP) and the normal stress axis is 2θ, the angle θ between the shear failure plane and the orientation of the maximum principal stress is given by the relationship (e.g. Cox et al. 2001)

$$\theta = (90^{\circ} - \tan^{-1}\mu_s)/2 \tag{12}$$

Then, for typical friction coefficients of approximately 0,70, shear fractures are inclined at nearly ±30° to σ_1. Shear fractures obeying the aforementioned relationship are termed 'optimally oriented' faults (Sibson 1985). When preexisting faults with fault gouge or other low strength material are present, μ_s may be as low as 0,2-0,4, and slip on faults over a range of orientations could occur (Richardson & Solomon 1977, Sauber et al. 2000).

Extension failure and extensional shear failure occur if the normal stress and shear stress state (Mohr circle) contacts the nonlinear part of the failure envelope at negative (i.e. tensile) normal stresses (Fig. 10). Pure extension failure occurs where the normal stress equals the tensile strength (T_s) of the rock. The angle 2θ is zero, as expected, for extensional failure along the σ_1-σ_2 plane, which is perpendicular to σ_3. Hybrid extensional shear failure occurs where the stress circle contacts the failure envelope between $-T_s$ and C_s (Sibson & Scott 1998, Twiss & Moores 2000, Cox et al. 2001).

Where pore and fracture space is saturated with aqueous fluid and freely interconnected up to the water table, the state of fluid pressure (P_f) is said to be hydrostatic (Price 1966, Hobbs et al. 1976, Sibson 2001). The level of fluid pressure, P_f, is related to the vertical stress, S_V, by the pore-fluid factor λ_v as

$$\lambda_v = P_f/S_V \tag{13}$$

so that for typical rock densities, hydrostatic fluid pressures are represented by λ_v ~ 0,4 (the ratio of water to rock densities; Hubbert & Rubey 1959, Secor 1965, Sibson 2004). Overpressured fluids may be described as suprahydrostatic (0,4 < λ_v < 1,0), lithostatic (λ_v = 1,0) or even supralithostatic (λ_v > 1,0; Sibson 2001).

Pore fluid pressure modifies stress states at depth in the Earth´s crust (Hubbert & Rubey 1959). Fluid pressure for shallow midplate earthquakes may exceed hydrostatic pressure, thus reducing the friction that must be overcome for faulting to occur (Sibson 1977). The effect of pore fluid pressure is to reduce the effective normal stress (σ_n') according to the relationship (Secor 1965, Hobbs et al. 1976)

$$\sigma_n' = \sigma_n - P_f \tag{14}$$

Although fluid pressure modifies normal stress, it does not influence shear stress (e.g. Cox et al. 2001). In terms of the Mohr circle representation of stress states, the role of fluid pressure is to move the stress circle to the left.

Mohr diagrams can also be extended to three-dimensional stress fields (e.g. Twiss & Moores 2000). In 3D Mohr representations the vertical axis is denoted as τ_s/S_V and the horizontal axis is given by $(\sigma_n\text{-}P_f)/S_V$. That is, shear stress and effective normal stress are normalized by vertical stress. All the principal stresses are marked in the horizontal axis and joined with stress circles (e.g. Barton et al. 1995a, 1995b, 1995c).

Under fluid-present conditions, frictional strength (τ_f) of a fault under normal stress may be represented by (Price 1966, Byerlee 1978, Sibson 2004)

$$\tau_f = C_s + \mu_s(\sigma_n - P_f) \tag{15}$$

According to the Mohr-Coulomb failure criteria, brittle shear failure occurs where and when shear stress τ_s is greater than the rock shear strength τ_f. That is, where

$$\tau_s \geq C_s + \mu_s(\sigma_n - P_f) \tag{16}$$

The Coulomb failure stress (CFS, e.g. Lund et al. 2009) is defined as

$$CFS = \tau_f - \mu_s(\sigma_n - P_f) - C_s \tag{17}$$

For frictional reactivation of existing faults (Fig. 10) it is often considered that $C_s \rightarrow 0$ so that frictional strength then reduces to

$$\tau_f = \mu_s(\sigma_n - P_f) \tag{18}$$

Consideration of this failure criterion suggests that fault instability could be induced either by an increase in shear stress τ_s due to elastic strain accumulation, by a decrease in normal stress σ_n, or by an increase in fluid pressure P_f causing a reduction in effective normal stress across potential slip planes (Sibson 1989, 1990, 2001). Similarly, the growth of hydraulic extension fractures will occur where and when $P_f \geq \sigma_3 + T_s$. That is, where and when fluid pressure is highest, σ_3 is least, or tensile strength is least (Cox et al. 2001).

Combining the Mohr-Coulomb failure theory with laboratory-derived coefficients of friction (e.g. Byerlee 1978) leads to the conclusion that the brittle strength of the crust is of the order of several hundred megapascals under hydrostatic pore pressure conditions, but vanishingly small as pore pressures approach lithostatic (Hubbert & Rubey 1959, Sibson 1974, Zoback et al. 2002). Hence, fluids play an important role in faulting.

4.1.4 The World Stress Map Project

The World Stress Map (WSM) Project, which was initiated in 1986, is a global cooperative effort to compile and interpret data on the orientation and relative magnitudes of the contemporary in situ tectonic stress field and to characterize the stress patterns and understand the stress sources in the Earth's lithosphere (Zoback 1992a). The project is maintained at the Helmholtz Centre Potsdam - GFZ German Research Centre for Geosciences (Heidbach 2009). The project database complements a number of regional data compilations, e.g. the Canadian Crustal Stress Database (Adams 1987, 1995, Cassidy et al. 2008), the European Stress Data Base (Müller et al. 1992), and the Fennoscandian Rock Stress Data Base (Stephansson et al. 1986, 1987). The current WSM digital database release of 2008 comprises 21,750 data points worldwide (Heidbach et al. 2008b, Heidbach 2009).

In the WSM, different types of stress indicators are used to determine the tectonic stress orientation. They are grouped into four categories (Reinecker et al. 2005): earthquake focal mechanisms (fault-plane solutions), well bore breakouts and drilling-induced fractures, in-situ stress measurements (overcoring, hydraulic fracturing, borehole slotter) and young (mostly Quaternary) geologic data (from fault-slip analysis and volcanic vent alignments). Hydraulic fractures, for example, propagate in a direction perpendicular to the minimum principal compressive stress (Hubbert & Willis 1957, Yale 2003). On the other hand, well-bore breakouts are elongated in the direction of the minimum horizontal stress and are considered reliable indicators of horizontal stress directions (Zoback & Zoback 1991). Most of the stress information results from earthquake focal mechanism solutions. A detailed description of the different stress determination methodologies can be found in Zoback et al. (1989), Zoback (1992a) and Amadei & Stephansson (1997). Ljunggren et al. (2003) give an overview of methods of rock stress measurement, with the emphasis on methods applicable to hard rocks and Fennoscandia.

In the WSM, geographic north is chosen as stress reference direction. All data are quality ranked between A and E. The A- and C-quality, for example, indicate that the maximum horizontal stress S_H orientation is accurate to within ±15º and ±25º, respectively. In general, A-C quality data records are considered reliable, e.g. for the use in regional stress analysis and the interpretation of geodynamic processes (Fuchs & Müller 2001).

Stress regimes are defined using the model of Anderson (1905, 1972), based on the relative magnitude of the vertical stress S_V and the maximum and minimum horizontal stresses S_H and S_h, respectively. The three major stress regimes include the normal faulting stress regime (NF) with $Sv>S_H>S_h$, the strike-slip faulting stress regime (SS) with $S_H>S_V>S_h$ and the thrust or reverse faulting stress regime (TF) with $S_H>S_h>S_V$ (Fig. 11; Amadei & Stephansson 1997, Fuchs & Müller 2001). In addition to these three basic regimes, Zoback et al. (1989) and Zoback (1992a) consider transitional stress regimes such as $S_V \approx S_H>S_h$, which produces a combination of normal and strike-slip faulting (NS), and $S_H>S_h \approx S_V$, which produces a combination of thrust and strike-slip faulting (TS; Fig. 11). Importantly, when S_h and S_V are almost equal, a very small change in the horizontal stress field can cause the style of faulting to change from strike-slip to thrust mechanisms or vice versa (Talwani & Rajendran 1991).

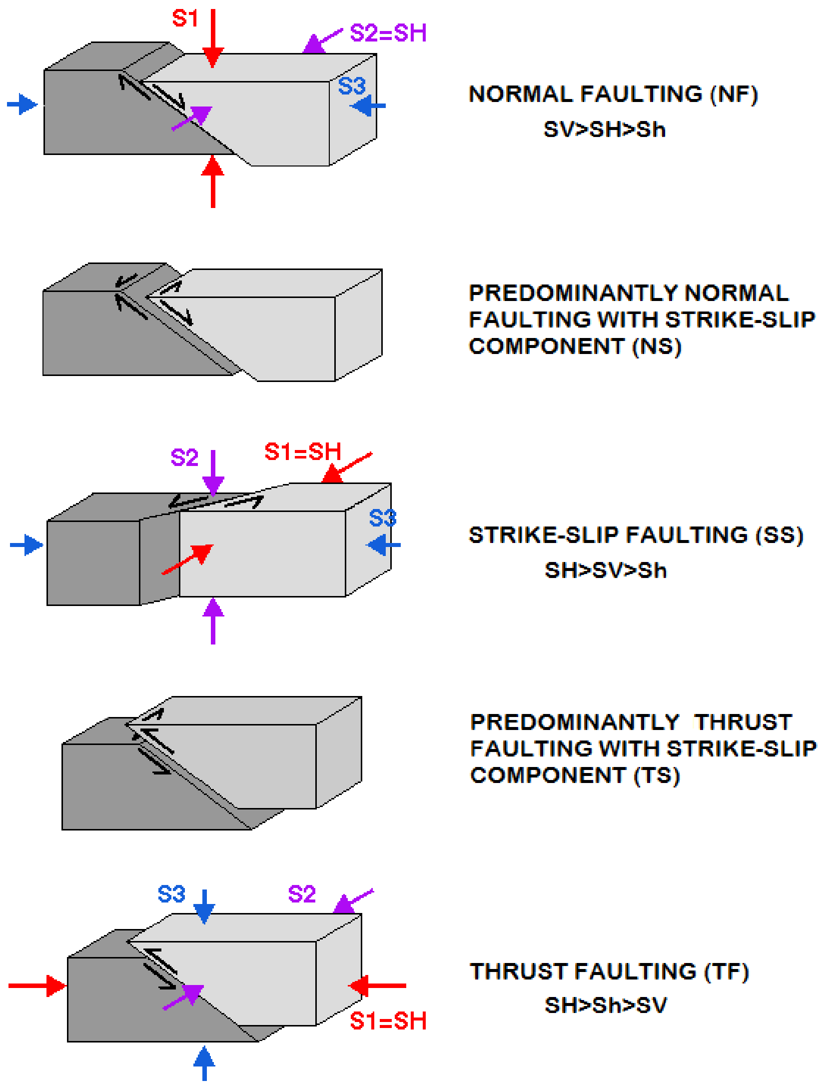

Fig. 11. Different stress regimes and faulting styles. Modified from the WSM project at http://www-wsm.physik.uni-karlsruhe.de/pub/introduction/introduction_frame.html.

The WSM shows clearly the existence of broad-scale tectonic stress provinces and various first-order (broad-scale regional) and second-order (more local) stress patterns in the upper and middle part of the Earth´s crust (Zoback et al. 1989, Zoback 1992a, Stewart & Hancock 1994). Regions of uniform horizontal stress orientation include, for instance, eastern North America, northern South America, and northeastern Asia (Heidbach et al. 2010). Most midplate regions fall within horizontal compressive stress regimes where faulting is characterized by reverse slip. Zoback (1992a) and Nelson et al. (2006) demonstrate that there is a good regional correlation between stress orientations inferred from the near surface and those calculated from earthquake focal mechanisms. Such consistency argues strongly for a tectonic origin for such stresses (Richardson & Solomon 1977, Klein & Barr 1986, Adams & Bell 1991, Zoback & Zoback 1991, Richardson 1992, Engelder 1993, Bezerra 1998, Fuchs & Müller 2001). On the other hand, the observations of Heidbach et al. (2008a, 2010) suggest that local stress sources, such as density contrasts and active fault systems, control the plate tectonic stress pattern in some regions, e.g. in Scandinavia and western Europe. Barton and Zoback (1994) describe stress field discontinuities associated with slip on active faults penetrated by boreholes.

The state of stress in Europe has been well studied. There are numerous fault plane solutions, in situ strain measurements, and a variety of geological indicators of past and recent state of stress (e.g. Reinecker et al. 2005, Heidbach 2009). The directions of the maximum horizontal compressive stress, with few exceptions, trend NW-SE in western Europe and from N-S to NE-SW in eastern Central Europe (Ahorner 1975, Richardsson et al. 1979, Slunga 1979, 1989a, Carlsson & Christiansson 1986, Klein & Barr 1986, Stephansson 1988, Bungum 1989, Stein et al. 1989, Talbot & Slunga 1989, Wahlström 1989, Klasson & Leijon 1990, Gregersen et al. 1991, Ahlbom et al. 1992, Grünthal & Stromeyer 1992, Müller et al. 1992, 1997, Zoback 1992a, Arvidsson & Kulhanek 1994, Arvidsson 1996, Gölke & Coblentz 1996, Brudy et al. 1997, Bungum & Lindholm 1997, Wahlström & Assinovskaya 1998, Fjeldskaar et al. 2000, Hicks et al. 2000a, 2000b, Hakami et al. 2002, Roberts & Myrvang 2004, Pascal et al. 2005).

Müller et al. (1992) note that NW-SE stress orientation in western Europe is sub-parallel to the direction of relative plate motion between Africa and Europe and is rotated 17° clockwise from the direction of absolute plate motion. The present-day maximum horizontal stress direction was established during the Middle and Late Miocene, that is approximately 10 Ma before present (Bergerat 1987, Whittaker et al. 1989, Becker 1993, Hibsch et al. 1995, Muir Wood 1995), and it is unlikely to be changed over the next few hundred thousand years (Saari 2000, La Pointe & Hermanson 2002).

4.1.5 Rock stress in Fennoscandia

The first rock stress measurements in Fennoscandia were done for mining in Sweden in 1951, in Norway in 1958 and in Finland in 1961 (Matikainen et al. 1981, Stephansson et al. 1986, 1987, Muir Wood 1993a). Later civil engineering purposes (e.g. underground storage facilities, repositories for radioactive waste) have gained more interest (Lappalainen 1980, Bjarnason & Stephansson 1986, Klasson & Leijon 1990, Tolppanen & Särkkä 1999a, 1999b). Nils Bernhard Hast gave the first presentations of stresses in Sweden in the late 1950’s (Hast 1969). Ranalli and Chandler (1975) were among the first to compile stress data from Scandinavia. The Fennoscandian Rock Stress Data Base was started in 1986 (Stephansson et al. 1986, 1987). Stephansson et al. (1986) and Muir Wood (1993a) give a review of the history of rock stress measurements in Fennoscandia.

Fennoscandia, as most intraplate regions, is characterized by compressive stress regime. The maximum horizontal stress S_H is orientated broadly in (W)NW-(E)SE direction (Fig. 12; Slunga 1981a, 1991, Bjarnason & Stephansson 1986, Carlsson & Christiansson 1987, Stephansson et al. 1987, Hansen et al. 1989, Wikberg et al. 1991, Ahlbom et al. 1992, Lindholm et al. 1995, Saari & Slunga 1996, Kakkuri 1997, Rhén et al. 1997, Lund & Zoback 1999, Dehls et al. 2000a, Fejerskov & Lindholm 2000, Lindholm & Bungum 2000, Brudy & Kjørholt 2001, Sjöberg et al. 2005, Martin 2007).

Compared to western Europe there are significant variations in the orientation of the maximum horizontal stress in Fennoscandia (e.g. Henderson 1991, Slunga 1991, Gregersen 1992). Clauss et al. (1989) suggest that the stress field is irregular in the cold and stabile shield areas due to irregular block movements. Stress indicators from depths of 300 m or deeper show greater consistency in the NW-SE direction (Stephansson 1988, 1989, Muir Wood 1993a, Lambeck 2005). In Äspö, Sweden, the direction of the maximum horizontal stress is 320º at the depth of 1 km (Wikberg et al. 1991). The corresponding direction in Forsmark, Sweden, is 325° to the depth of 600 m, at least (Martin 2007).

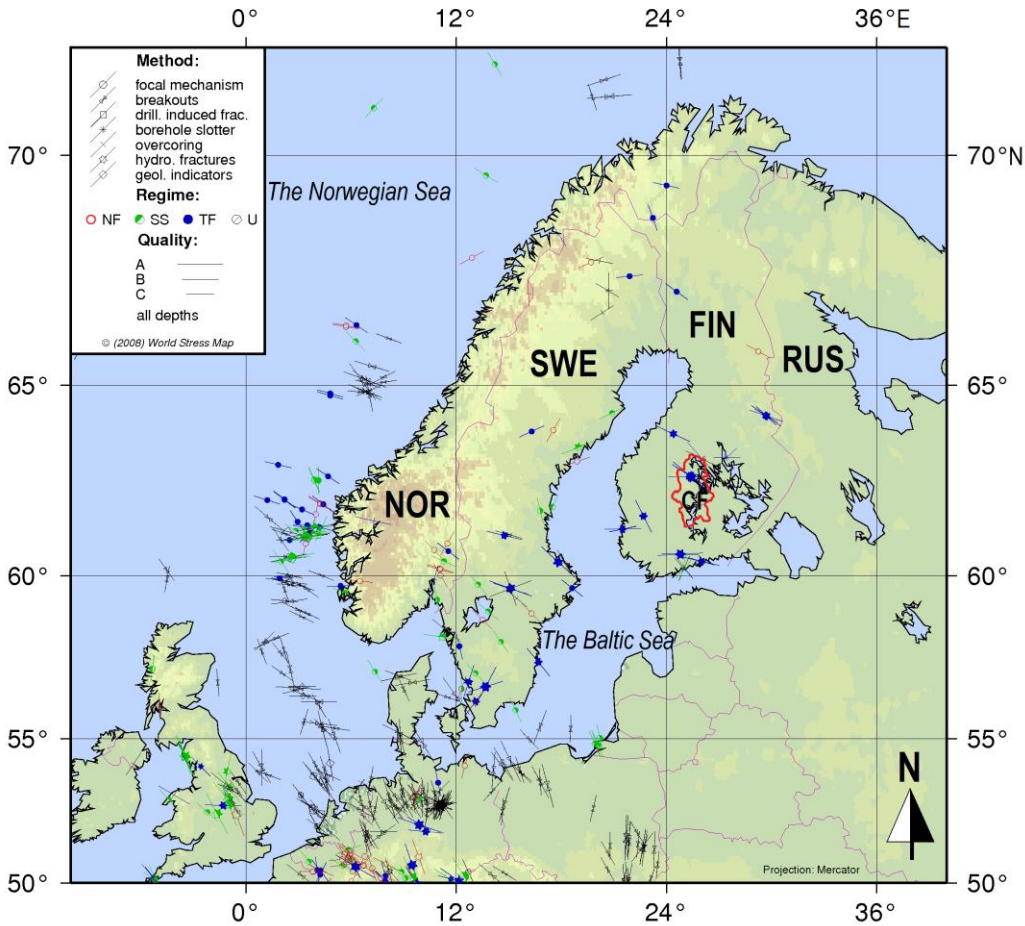

Fig. 12. Maximum horizontal stress (S_H) directions in northern Europe. Stress regimes are indicated as normal (NF), strike-slip (SS), thrust (TF) or undetermined (U). NOR=Norway, SWE=Sweden, FIN=Finland, RUS=Russia. Central Finland (CF) is outlined in red. Modified from the WSM project at http://dc-app3-14.gfz-potsdam.de/pub/casmo/ casmo_frame.html.

The general NW-SE direction of the horizontal stress in Fennoscandia is in good accordance with the direction of ridge push forces from the North Atlantic ridge, which is a spreading type plate boundary locating about 1,000 kilometers offshore Norway (Meyer et al. 1988). However, in the Barents Sea, the maximum horizontal stress direction is almost N-S (Clauss et al. 1989, Stephansson et al. 1991, Gregersen 1992, Gölke & Brudy 1996, Gölke & Coblentz 1996).

Maximum horizontal stress typically varies from five MPa near the ground surface to 50 MPa at a depth of 1,000 m below ground surface (Stephansson et al. 1987, Wikberg et al. 1991, Leijon 1993). In general, the maximum and minimum horizontal stresses exceed the vertical stress to at least 450-500 m depth as estimated from the weight of the overburden (e.g. Carlsson & Christiansson 1986, 1987, Stephansson et al. 1986, 1991, Martin 2007). In Äspö, Sweden, the horizontal minimum stress S_h and the vertical stress S_V are close to each other. However, measurements show that in some boreholes S_h is greater than S_V at a depth of 0,5-1,0 km (Wikberg et al. 1991). For comparison, in the Canadian Shield, the maximum principal stress varies between 60 and 130 MPa at the depth of 2 km (Herget 1986, Adams 1987, Adams & Bell 1991). Both horizontal stresses (S_H, S_h) are higher than the vertical stress S_V till the depth of 2,5 km, at least (Herget 1986, Martin et al. 2003, Mazurek 2004).

The stress pattern in Finland has been verified by the analysis of earthquake fault plane solutions, in situ stress measurements and geodetic information. The current tectonic stress field is rather consistently in the NW-SE direction especially in central Finland (Fig. 12; Kakkuri 1993, 1997, Saari 2000). A relatively strong horizontal stress field has been measured in the upper part of the bedrock, generally being of the order 5-15 MPa at a depth of 0-100 m (Lappalainen 1980, Matikainen et al. 1981). In the four locations chosen as candidates for disposal of high level nuclear fuel waste in Finland the maximum horizontal stress direction S_H varies between 288º and 309º; the estimated magnitudes of the minimum (S_h) and maximum (S_H) horizontal stresses vary at the depths of 500 and 800 m 13-23 and 23-47 MPa, respectively (Klasson & Leijon 1990).

The analysis of the Lappajärvi M_L 3,8-2,6 earthquakes in western Finland in 1979 gave 311° and 316° as directions for the maximum horizontal stress (Slunga 1989a). In Anjalankoski, southeastern Finland, the focal mechanisms of the earthquake swarm corresponded to a 340º orientated horizontal compression (Uski et al. 2004, 2006). In the Kuusamo area, northeastern Finland, the stress field is extensional and deviates significantly from the broad NW-SE stress field (Uski et al. 2003a, Ojala et al. 2004). In Finland, topographically generated stresses are anticipated to be small (Lambeck & Purcell 2003).

According to Barton et al. (1997a), stress orientation is consistent both at a local scale and with regional trends, when relatively strong elastic bedrock is exhumed extending near/to the surface and there are virtually no topographic stress effects. Then the high near-surface compressional stresses are likely the response of the lithosphere to plate-driving forces. This is largely the situation in southern and central Finland.

4.1.6 Rock stress and permeability

Until recently it has invariably been assumed that, in a fractured rock mass, the most effective fluid conduits are extension fractures that parallel the maximum in-situ stress and are perpendicular to the minimum in-situ stress. This is consistent with the notion that extension fractures will experience the lowest normal stress, i.e. the least degree of closure, and have a maximum dilation tendency, and would thus be the most permeable (Larsson 1963, 1971, 1984, Secor 1965, Baweja & Raju 1984, Larsson et al. 1984, Ericsson & Ronge 1986, Lang et al. 1986, Pettersson 1987, Boeckh 1992, Heffer & Lean 1993, Larsson & Tullborg 1993, Anon 1996, Midtbø 1996, Portugal Ferreira & Pacheco 1997, Braathen et al. 1998, 1999, Gaut et al. 1999, Singhal & Gupta 1999, Sirat 1999, Savage & Morin 2001, Hung et al. 2002, Domoney et al. 2003, Fernandes 2003, Sami 2009).

The corollary is that fractures perpendicular to the maximum in-situ stress, i.e. experiencing the highest normal stress, would offer the greatest resistance to fluid flow and therefore have the lowest permeability (Larsson 1959, Carlsson 1979, Carlsson & Olsson 1983, Sundqvist et al. 1988, Faunt 1997, Ferrill et al. 1999, Braathen & Gabrielsen 2000, Coriolano et al. 2000, Lyslo 2000, Lachassagne et al. 2001, Coriolano 2002, Owen et al. 2003, Travaglia & Dainelli 2003, Juhlin & Stephens 2006, Singhal 2008).

Many authors have documented higher fracture permeabilities and well/spring yields along extensional structures parallel to S_H in crystalline rock areas (e.g. Batchelor & Pine 1986, Carlsson & Christiansson 1987, Galanos & Rokos 2006, Neves & Morales 2007a, 2007b). Concurrently, compressive structures perpendicular to S_H have considered lower transmissive (e.g. Larsson 1972, Carlsson & Olsson 1978, 1979, 1980, 1981, Travaglia & Ammar 1998, Bai et al. 1999, Neumann & Sami 2000, Fernandes & Rudolph 2001, Banks

& Robins 2002, Ganerød 2003, Morin & Savage 2003, Ligtenberg 2005, Henriksen & Braathen 2006, Owen et al. 2007, Sami 2009, Chen et al. 2011).

At present, however, it is widely accepted that favorably oriented and critically stressed shear fractures, i.e. those with a high shear to normal stress ratio at or close to frictional failure in the current-day stress field, can also serve as conduits for active fluid flow (e.g. Barton et al. 1995a, 1995b, 1995c, 1997a, 1998, Hickman et al. 1997, Barton & Moos 1998, Chen & Bai 1998, Dholakia et al. 1998, Coriolano et al. 2000, Hillis et al. 2000, Ito & Zoback 2000, Tezuka & Niitsuma 2000, Townend & Zoback 2000, Tröger et al. 2001, Zoback & Townend 2001, Coriolano 2002, Rogers & Evans 2002, Ameen 2003, Chanchani et al. 2003, Ito & Hayashi 2003, Rogers 2003, Min et al. 2004, Geier 2005, Moeck 2005, Moeck et al. 2005, 2007, Henriksen & Braathen 2006).

According to Zoback (2011), critically stressed faults are both mechanically and hydraulically alive whereas not critically stressed faults are both mechanically and hydraulically dead. However, a number of other factors may control the actual permeability of a fault such as the degree of alteration and cementation of the brecciated rock within the fracture and its diagenetic history as well as the current effective normal stress (Zoback 2011).

Along with extension fractures, critically stressed shear fractures have been related to elevated well yields in crystalline basement regions around the world (e.g. Baweja & Raju 1984, Batchelor & Pine 1986, Nascimento da Silva & Jardim de Sá 2000, Talbot & Sirat 2001, Henriksen 2006a, 2006b, Solomon & Ghebreab 2008).

Townend and Zoback (2000) argue that it is the presence of critically stressed faults deep within the brittle crust that keeps the bulk permeability of the crust about four orders of magnitude greater than intact rock samples subjected to appropriate confining pressures. In fractured rock reservoirs, Barton et al. (1997b) have noted that noncritically stressed fractures are not conductive even if they are mode I cracks aligned perpendicular to the least principal stress. In sedimentary rocks of the Monterey Formation, California, United States, Finkbeiner et al. (1997) have found that critically stressed reverse faults and bedding planes orientated subperpendicular to the S_H in a thrust regime provide highly permeable fluid migration paths. According to King et al. (2008), pre-existing, moderately dipping reverse faults striking perpendicular to the S_H have potential for reactivation within the combined thrust and strike-slip regime in the Northern Perth Basin, Australia. In northern Portugal, the direction of the fracture zones with highest spring densities is perpendicular to the regional horizontal compressive stress S_H (Pacheco & Alencoão 2002). According to Zoback (2011), mode I tensile fractures (although relatively ubiquitous in nature) are unlikely to have a significant effect on hydraulic properties, such as bulk permeability, at depth because they are essentially closed at any finite effective stress.

There are also examples of no clear relation between well yield and stress distribution in crystalline rock areas. In the Gothenburg area, southwestern Sweden, Wladis (1995) found no influence of stress on the normalized yield (Q/d) of drilled wells. Banks et al. (1996) drilled boreholes perpendicular to S_H on the Hvaler Island, southeastern Norway, but the borehole orientation with respect to stress field appeared to have at best only a minor impact on borehole yield. They suggested that the impact of in-situ stress anisotropy was overshadowed by other factors, such as orientation, connectivity and mineralization of fractures.

4.2 Earthquakes and seismic activity

4.2.1 Seismic concepts and definitions

An earthquake is a sudden fracturing and release of energy or stress, which has generally continued to build up over a long time at locked parts (asperities) of a stably sliding fault (e.g. Saari 1992, Engelder 1993). A seismic cycle is defined as a repeated earthquake-generating fault slip event with a period of strain build-up that culminates with energy release (e.g. Cloos 2009). The slip rate of a fault describes the sum of discrete displacements, as a result of numerous seismic events, divided by time (Scholz 1990).

The continuous-creep mechanism, i.e. a very slow flow under a constant stress, involves uninterrupted motion along a fault so that strain is relieved continuously and does not accumulate (Sibson 1977, Hatcher 1990). According to Slunga (1990, 1991), there is significant aseismic sliding (about 1 mm/year/100 km) along faults in Sweden and aseismic fault movements are more than 20,000 times more extensive than the seismic fault movements (see also Ahorner 1975).

Active earthquakes worldwide are mainly localized along the plate boundaries, of which the subduction zones contain the most active and serious earthquakes. There also occur earthquakes inside the plates. These are called intraplate earthquakes (e.g. Scholz 1990, Saari 1992). The seismic activity, with respect to the magnitudes and earthquake density, in the Fennoscandian Shield seems to decrease relative to the distance from the North Atlantic Ridge (e.g. Stephansson 1988, Gregersen et al. 1991).

Earthquake magnitude is a measure of the size of an earthquake, and is related to the amplitude of observed ground vibrations at distance from the source due to the generated seismic waves (e.g. Bödvarsson et al. 2006). A modified version of a method originally developed by Charles Richter in 1935 is used worldwide as standard earthquake magnitude. The Richter scale is logarithmic, with an increase of one on the magnitude scale representing a tenfold increase in wave amplitude, and roughly a 30-fold increase in the energy released (Nilsen & Palmström 2000). Magnitude scales have no upper or lower limits. The most common magnitude scales are surface wave magnitude (M_S), body wave magnitude (M_b), moment magnitude (M_W) and local magnitude (M_L; Stewart & Hancock 1994, Tvedt et al. 2002, Bödvarsson et al. 2006).

The seismic energy most commonly is released from a fault below the surface, called the focus or hypocenter. The point on the surface vertically above this point is called the epicenter (e.g. Cloos 2009). The earthquakes are known to originate at depths ranging from about one to nearly 700 kilometers. Most large earthquakes nucleate in the depth range of ~ 5-25 km (e.g. Scholz 1990, Cloos 2009). Commonly earthquakes are due to reactivation of existing faults; the creation of entirely new faults is rare (e.g. McKenzie 1969, Artyushkov 1983, Munier 1993, Sibson 1994).

Faulting may or may not carry through to the surface, and generally is limited to relatively narrow, active faults. However, very few earthquake-generating ruptures propagate to the surface; it is uncommon even for events between M 6...7 (Cloos 2009). Typical displacements when they exist vary from a few centimeters to meters (Nilsen & Palmström 2000). Surface slip events of 10 m are extraordinary and only occur during great earthquakes (Cloos 2009). There are quantitative relationships between earthquake magnitude, the size of rupture areas and the amount of fault slip (Sibson 1986). For instance, the maximum slip for magnitude 7...9 earthquakes is from 2 to 15 meters, whereas for M 2...6 earthquakes the corresponding slip varies from 0,6 mm to 33 cm (Wells & Coppersmith 1994, Cloos 2009).

Seismologists refer to the direction of slip in an earthquake and the orientation of the fault on which it occurs as the fault-plane solution or focal mechanism (Engelder 1993, Stewart & Hancock 1994, http://quake.usgs.gov/recenteqs/). Since S_H directions deduced from focal mechanisms are assumed to be at 45º to the orientation of the fault plane, they may be in error as much as ±45º. The S_H directions derived from thrust and normal fault focal mechanisms are probably much better constrained than S_H directions derived from strike-slip focal mechanisms (Suter 1991, Zoback 2011). There are a number of cases where the same earthquake has even published in two or more papers with radically different results, especially with regard to the azimuths of pressure axis (Ebel & Kafka 1991, Bent 2002, Hyvönen 2008). In addition, earthquake fault-plane solutions are not always compatible with the in situ stress measurements, as the earthquakes sometimes appear to have thrust-faulting mechanisms deeper in the crust despite an indication that the in situ stress measurements should favor strike-slip faulting near the surface (Engelder 1993).

4.2.2 Seismicity and focal mechanisms in Fennoscandia

The seismicity of Fennoscandia is typical for intraplate regions, being characterized by relatively small magnitude earthquakes of infrequent occurrence in zones of weakness (Husebye et al. 1978, Saari 1992). The stress build-up in Fennoscandia is released mainly through microseismicity below the detection limit of the relatively sparse seismic network. In addition, most of the earthquakes happen at a depth of 5 to 20 kilometers, without any surface trace (Slunga et al. 1984, Koskiahde et al. 1985, Scholz 1990, Ojala et al. 2006, Hyvönen 2008). Kaikkonen et al. (2000) have concluded that 80% of the earthquakes in the central Fennoscandian Shield occur at less than 14 km and that a depth of 31 km is the lower limit of the seismogenic crust. In northern Sweden, the focal depths are generally from 3 to 8 km (Slunga 1991).

Seismically active zones are found along the Norwegian coast, in the northern central part of the shield, along the western coast of the Gulf of Bothnia, and around the great lakes in southern Sweden (Fig. 13). Most of the largest earthquakes occur offshore Norway. Lowest seismic levels are reached in the southern half of Finland, where the magnitude of all earthquakes in human records is less than five (Stephansson 1979, Ahjos & Uski 1992, Mäntyniemi et al. 1993, Arvidsson & Kulhánek 1994, Saari & Slunga 1996, Olesen et al. 2000a, Bödvarsson et al. 2006, Ojala et al. 2006, Söderbäck 2008). Skordas et al. (1991) have demonstrated that the seismic events in the North Atlantic Ridge have a clear correlation with the Fennoscandian Shield. The best correlation comes from the events in the ridge between Jan Mayen and Iceland, i.e. from the segment 58-70ºN (see also Olsson 1998).

Talvitie (1978, 1979) suggests that the seismic activity along the Proterozoic Senja and Lapland fracture zone segments indicates that the North Atlantic Ridge push is guided inside the Fennoscandian Shield into/along NW-SE –trending strike-slip faults. Muir Wood (1993a, 1995), however, argues that the rheological differences between the Fennoscandian Shield and its surroundings prevent this. According to Talbot and Slunga (1989), some of the transform faults, the Jan Mayen Transform Fault and the Iceland Transform Fault, may continue into zones of weakness on land in the Fennoscandian Shield. According to Barosh (1986), the northwest-trending fracture zones in eastern United States can be directly or undirectly related to the offshore transform faults in the North Atlantic basin.

The earthquake observations from Fennoscandia have been compiled in a database maintained by the Institute of Seismology, University of Helsinki. The FENCAT database covers all the Nordic countries, the Kola Peninsula and former Soviet Karelia. It is constantly updated and available over the Internet at http://www.seismo.helsinki.fi/fi/bulletiinit/index.html (Ahjos & Uski 1992, Mäntyniemi et al. 2004). The historical data is based predominantly on written documents of macroseismic observations. During instrumental period, earthquakes are recorded mainly by seismic instruments. The first short-period seismograph station was installed in Finland in 1956 (Saari 1992). A review of the seismic information on Fennoscandia is presented in a paper of Husebye et al. (1978).

The oldest known report on seismic activity in Fennoscandia dates back to 1357, when a relatively large earthquake was felt in Gotland, Sweden. The oldest reports for Finland and Norway date back to 1610 and 1612, respectively (Husebye et al. 1978 and references therein). The probably two largest earthquakes reported in Scandinavia took place in Norway in 1819 and 1904 and had magnitudes around 6,0 (Husebye et al. 1978). The largest M_L magnitude earthquake determined in Finland is 4,9 (Bay of Bothnia 65,6ºN 24,5ºE 23/06/1882, http://www.seismo.helsinki.fi/bulletin/list/catalog/Suomi_n.html; Saari & Slunga 1996). On October 25, 1976, an unusually large (M_b 4,5) strike/dip-slip earthquake occurred in the Gulf of Finland near Osmussaar, Estonia (Slunga 1979). According to Wahlström (1980), relatively strong (M_L 3,2) earthquakes can occur also in the uppermost part (< 3 km) of the Fennoscandian Shield.

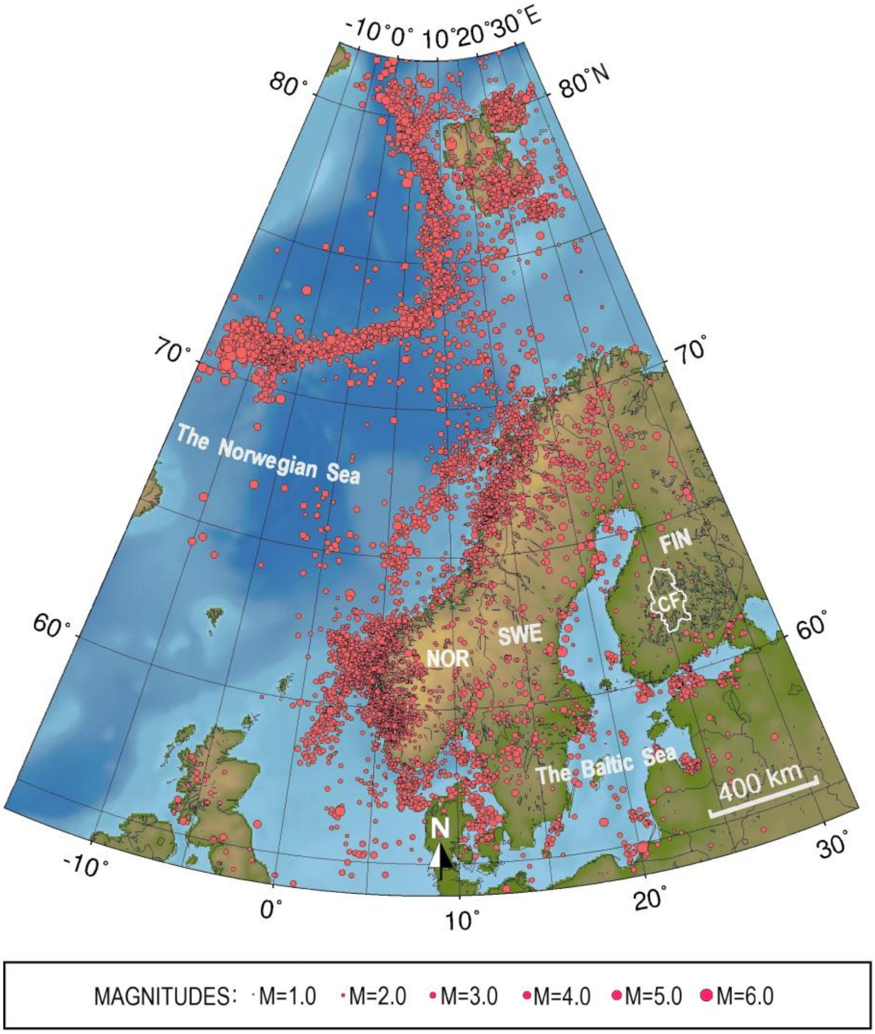

Fig. 13. The seismicity of Fennoscandia and surrounding areas in the period 1985-2005 as recorded by the Norwegian National Seismic Network. Known and probable explosions have not been included in the map. NOR=Norway, SWE=Sweden, FIN=Finland, CF=Central Finland. Modified from Sørensen (2006).

Bödvarsson et al. (2006) estimate that, in Sweden, one should expect at least one magnitude 5 earthquake every century, one magnitude 6 earthquake every one thousand years and one magnitude 7 earthquake every 10,000 years. Saari´s (2000) analysis of future seismicity in Finland strongly suggests that future earthquakes related to the ridge-push will be strike-slip motions preferentially (>90%) along faults striking NW-SE or N-S. These earthquakes may have a maximum M_L = 5,5 (Ahjos et al. 1984, La Pointe & Hermanson 2002).

About 200 fault plane solutions for the Fennoscandian Shield were summarized by Slunga (1989a, 1990, 1991). He concluded that the dominant type of faulting is strike–slip at subvertical fault planes (see also Slunga 1981a). In southern and central Sweden half of the earthquake events are of the strike-slip type, one quarter normal and one quarter reverse. In northern Sweden, reverse faulting is more common than normal faulting (Slunga et al. 1984). According to Gregersen (1992), 50% of the focal mechanisms in Fennoscandia are strike-slip faulting, while respectively 10%, 10% and 30% are thrust, normal and oblique together with indeterminable mechanisms. According to Müller et al. (1997), the large-scale horizontal compression in Fennoscandia is expressed mainly by thrust and strike-slip faulting but also by normal faulting as short-scale variations of the tectonic regimes.

The Fennoscandian reverse faults have typically NE-SW strikes, the normal faults have NW-SE strikes while the strike-slip faults have either N-S or E-W strikes (Slunga 1981a, 1981b, Wahlström 1982, Slunga et al. 1984, Koskiahde et al. 1985). The fault plane solution data are consistent with a N60ºW (300º) direction of the regional principal compression (Slunga & Nordgren 1987, Slunga 1989b, 1989c, 1990, 1991, Gregersen 1993, Gregersen et al. 2007). Focal mechanism solutions for three M 5 earthquakes offshore western Norway in 1986-1989 indicated thrust faulting along N-S to NNE-SSW striking fault plane. For instance, the 1988 M 5,3 earthquake in the central Møre Basin, offshore Norway, occurred at a depth of ca. 25 km and showed reverse faulting with WNW-ESE compressive stress (Hansen et al. 1989). Four M_L 1,9-4,5 earthquakes onshore southwestern Norway between 1983 and 2000 showed thrust faulting with focal depths of 10-18 km and maximum compression in WNW-ESE direction (Hicks & Ottemöller 2001).

In response to the NW-SE regional principal compressive stress in Norway, Karpuz et al. (1991) predict shear movements along conjugate NNW-SSE trending sinistral and WNW-ESE trending dextral faults, and normal movements along NW-SE and reverse movements along NE-SW trending faults. However, as they state, this simplistic pattern would be complicated by stress field variations associated with pre-existing structures as well as the overprinting of isostatic and topographic effects.

In Finland, the low level of seismic activity, the sparse coverage of the national seismic network and the limitations of the instrumentation have restricted the source mechanism inversions to few optimally situated earthquakes. The available fault plane solutions indicate dominantly strike-slip movement along nearly vertical fault planes (Slunga & Ahjos 1986, Uski et al. 2002, 2003a, 2003b, 2006, Ojala et al. 2004). Saari (2000) suggests that NW-SE – and N-S –orientated zones of weakness are advantageous for strike-slip faulting in Finland whereas reverse faulting is more plausible in NE-SW fracture zones perpendicular to the orientation of the main stress field.

The Kolari M_L 2,9 earthquake on the 5th of May 2001 in Finnish Lapland suggested pure thrust-faulting at a depth of 5 km with the nodal plain 035º/30º. The Kuusamo M_L 3,5 earthquake (focal depth 14 km) in the northeastern Finland on the 2nd of May 2001 had a normal-faulting mechanism with the nodal plane 133º/47º (Uski et al. 2002, 2003a). In Anjalankoski, southeastern Finland, the focal mechanisms of the earthquake swarm corresponded to dip-slip motion along a 250º/80º oriented fault plane and 340º orientated horizontal compression (Uski et al. 2004, 2006). The Laitila M_L 1,9 earthquake (focal depth 2,5 km) in the southwestern Finland on the 3rd of January 2007 had a reverse-faulting mechanism with an orientation of compression 313º (Saari 2008, Saari & Lakio 2008). Stress measurements from the deep borehole at Syyry to the north of Central Finland indicate strike-slip conditions between 400-600 m depth but thrust fault conditions at deeper measurement depths (Klasson & Leijon 1990). At Vihanti in central Finland, the stress data show thrust fault conditions for the complete measuring interval (200-600 m) and there are no clear indications for the stress conditions to change (Muir Wood 1993a).

4.2.3 Fault reactivation and permeability

The reactivation of preexisting faults and planes of weakness appears to be an important process in continental deformation (Anderson 1972, Muir Wood & Mallard 1992, Park 2000, Ranalli 2000). Large faults often undergo a long history of movement involving multiple displacements in different directions (e.g. Reading 1980, Geier 2005).

According to Muir-Wood and Mallard (1992), a fault that has moved in the current tectonic regime should be considered active (see also Stewart & Hancock 1994). In Norway, active or potentially active faults are referred to as those that have slipped or are likely to have slipped during the past 10^{4-5} years. The term recently active is normally used only for those faults that may have slipped in the past 10^{2-3} years (Lyslo 2000, Henriksen 2006b). In intraplate regions earthquakes typically reactivate preexisting faults; the recurrence times between brief episodes are long, 10^{4-5} years (Muir-Wood & Mallard 1992, Crone & Machette 1994, Crone et al. 1997).

The reactivation of preexisting faults depends on frictional parameters, depth, pore fluid pressure, stress field, and the orientation of the faults (e.g. Zoback 1992b, Alaniz-Alvarez et al. 1998, Ranalli 2000). Fault reactivation occurs when the shear stress along the fault (τ_s) equals the frictional strength of the fault (τ_f). This may happen due to rising shear stress, decreasing normal stress, or an increase in fluid pressure (Chapter 4.1.3; Sibson 1994). Most of the reactivation occurs on the fault and deformation rapidly decreases with increasing distance from the fault (Bäckblom & Munier 2002). A preexisting fault can be reactivated if the required critical stress difference (σ_1-σ_3) is less than the critical stress difference for the formation of a new fault, and the tectonic stress is sufficiently high (Ranalli 2000). The tectonic stress state in many settings appears to be governed by the critical stress required for reactivation of favorably oriented faults (Townend & Zoback 2000, Zoback 2011). Just as Hast (1969) has stated, bedrock is not in a state of static but dynamic equilibrium.

According to Sibson (1990), faults may be defined as favorably oriented for frictional reactivation, unfavorably oriented, or severely misoriented, depending on their attitude in the prevailing stress field. When the orientation of an existing fault is less favorably oriented for reactivation, the ratio of effective stress required for reactivation increases (Sibson 2001). However, it is not uncommon for faults that are unfavorably oriented or severely misoriented to remain active in compressional tectonic regimes within crystalline rock assemblages, if fluid pressures in rupture nucleation sites are elevated toward lithostatic values and/or the rupture sites contain low friction coefficients (Sibson 1989, 1990, Gowd et al. 1996, Ranalli 2000). In the Canadian Shield there are seismically active thrust faults with dip angles of the order 60º-70º, i.e. they are severely misoriented (Ranalli 2000).

The influence of preexisting primary and secondary structures on faulting is profound. For instance, reverse faulting and strike-slip faulting commonly take advantage of preexisting high-angle weakness in the crust, like ancient faults and shear zones (Hill 1982, Davis 1984, Hatcher & Williams 1986). Sykes (1978) mentions several examples where faults have been reactivated with a different sense of motion from that which first occurred on them. According to Engelder (1993), many faults are reactivated within a stress field, which was not responsible for their rupture. Ito and Zoback (2000) describe a Mesozoic thrust fault being reactivated as a strike-slip fault in the current stress field in the German continental deep drillhole KTB. Chanchani et al. (2003) describe hydrocarbon formation where the original reverse faulting regime has changed to strike-slip regime due to reservoir depletion.

There is geological evidence that many high-angle reverse faults have developed by the reactivation of old normal faults (Jackson et al. 1981, Winslow 1981, Kumarapeli 1987, Sibson et al. 1988, Sibson 1990, 2004). According to Johnston (1989a), virtually all intracontinental rifts in stable continental interiors, though formed under extension, are currently under compression and strike-slip and thrust earthquakes predominate, indicating reversal of faults originally created in a normal faulting sense. In offshore areas of Norway there are many reverse reactivations of old NE-striking normal faults (Lindholm et al. 1995, Wiprut & Zoback 2000, Wiprut 2001, Atakan & Ojeda 2005, Bungum et al. 2005). Also Davis (1984) suggests that one obvious explanation for the origin of some (steep) reverse-slip faults is that they occupy former sites of normal faulting or strike-slip faulting. Where reverse faulting is a result of fault reactivation, the dip of the reverse-slip fault is inherited from a previous event.

It is widely accepted that active (critically stressed close to failure) or recently reactivated fault zones are more permeable and likely to conduct water than inactive ones (Carlsson & Christiansson 1987, Ahlbom & Smellie 1991, Olsson & Brown 1993, Barton et al. 1995a, 1995b, 1995c, 1997a, 1997b, 1997c, Hickman et al. 1995, Koch & Mather 1997, Mayer & Sharp 1998, Yeo et al. 1998, Ferril et al. 1999, Gudmundsson 2000a, 2001, Lyslo 2000, Faybishenko et al. 2001, Gudmundsson et al. 2001, 2003, Lie & Gudmundsson 2002, Henriksen 2006b, 2008, Zoback 2011).

The raise of permeability in reactivated fault zones may happen through several mechanisms, including brecciation, increased surface roughness, and breakdown of seals (Olsson & Brown 1993, Barton et al. 1995a, Hickman et al. 1997, Townend & Zoback 2000, Ito & Hayashi 2003, Rutqvist & Stephansson 2003, Min et al. 2004, Zoback 2011). Fault displacements are concentrated on undulating surfaces, which generate and destroy void space during slip as the result of mismatched surface topography (Bruhn et al. 1994). The void space is elongated parallel to the slip direction and provides tortuous conduits that channel fluid flow (Hickman et al. 1995). On the other hand, continued shearing may reduce conductivity by grinding away contact points and producing gouge (Huntoon 1986, Teufel 1987, Aydin 2001, Ito & Hayashi 2003).

Because a fault zone typically is an interconnected network composed of fractures in a number of different orientations, the fault reactivation most probably increases the fracture interconnection. This in turn may well be a major factor controlling the movements of fluids through the rock masses (e.g. Gale 1983, Gudmundsson 2000a). Furthermore, Hickman et al. (1998, 1999) and Townend and Zoback (2000) suggest that fault zone permeability is high only when individual fractures as well as the overall fault zone are optimally oriented and critically stressed for frictional failure, so that the fault can intermittently slip to counteract the expected permeability reduction due to crack sealing.

The changes in the transmissivities of optimally oriented and critically stressed faults can be much higher than those of other faults (Olsson & Brown 1993, Barton et al. 1995a, Hickman et al. 1997, Ito & Hayashi 2003, Rogers 2003, Rutqvist & Stephansson 2003). Indeed, most fracture zones are highly permeable only for relatively short periods of time following a seismogenic fault slip (Sibson 1994, Gudmundsson 2000a, Lyslo 2000, Lie & Gudmundsson 2002). In New Zealand, for instance, high-permeability fault zones are found along the most recently reactivated structures, older faults having become choked with hydrothermal precipitates (Hickman et al. 1995). According to Hickman et al. (1995), permeability changes can operate at rates that are rapid (100…10,000 years) with respect to the recurrence intervals for large earthquakes.

In the vicinity of active faults that undergo intermittent rupturing, permeability and fluid flux may be tied to the earthquake cycle through a range of mechanisms, leading to complex interactions between stress cycling, pore pressure fluctuating from prefailure lithostatic to postfailure hydrostatic levels, fault gouge forming, mineral precipitation, cementation, mechanical clogging, and crack sealing and healing (Sibson et al. 1988, Sibson 1990, 1994, 2004, Sleep & Blanpied 1992, Bruhn et al. 1994, Hickman et al. 1995, Olsen et al. 1998, Ito & Zoback 2000, Lyslo 2000, Mazurek 2000, Renard et al. 2000, Meyer 2002, Gartrell et al. 2003, Ingebritsen & Manning 2010).

Thus, the hydrodynamic properties of the faults can vary with time and space, and active deformation is necessary to generate and maintain fault permeability (Cox et al. 2001).

4.2.4 Hydrological impacts of earthquakes

Hydrological changes associated with earthquakes have been known for more than 2,000 years. In particular, hydrological effects are related to earthquake focal mechanisms (Muir-Wood & King 1993).

For a normal fault earthquake, elastic rebound following fault rupture is compressional, for a reverse fault earthquake extensional (Muir Wood 1993a, 1994). Hence, large earthquakes that involve normal fault component, expel substantial quantities of water as springs, ponds and increasing river flows, whereas hydrological changes of reverse faulting events of the same magnitude are most notable by their absence or, to the contrary of normal events, by falling wells, drying springs and reduced river flows (Waller 1966, Karpuz et al. 1991, Rojstaczer & Wolf 1992, Muir-Wood 1993a, 1993b, 1994, Muir-Wood & King 1993, 1994, King & Muir Wood 1994, Rojstaczer et al. 1995, Fleeger et al. 1999, Olesen et al. 2000a, 2000c, Babiker & Gudmundsson 2004). For reverse fault displacements, highest fluid flows are found around and beyond the ends of the fault (Muir-Wood 1993b). Strike-slip events typically also expel water locally but not in the quantities associated with normal faulting events (Muir-Wood & King 1993, Tokunaga 1999).

In the interseismic period, water is squeezed out of the crust in compressional tectonic environment but drawn into the crust in extensional regimes. Then, in a region of compressional tectonics, crustal porosity and permeability should decrease between earthquakes, and then suddenly increase after the earthquake (Lagerbäck & Witschard 1983, Lagerbäck 1988, Muir Wood 1993b, 1994, Muir-Wood & King 1993, Olesen et al. 2000a). Hence, in an area of active tectonics, hydrology is dynamic and crustal porosity varies regionally through the earthquake cycle (Muir-Wood 1993b).

The significant hydrological impacts of large earthquakes suggest that even small events should modify hydrological conditions. Bonilla (1988) has shown that surface displacements ranging from a few millimeters to several decimeters have been usually associated with earthquakes having magnitudes between 5 and 6. However, under ideal conditions, magnitudes as small as 3 could result in a few millimeters surface faulting (Bonilla 1988). According to Wells and Coppersmith (1994), earthquakes in the magnitude range from M_L 4 to M_L 6 can produce vertical displacements of the order of 1-10 cm (see also La Pointe et al. 1997). According to Bödvarsson et al. (2006), smaller events with M_L less than 1 typically correspond to lateral movements of 0,001-1 mm, whereas larger events with magnitude around five have slips of 50-500 mm. The Laitila M_L 1,9 earthquake in the southwestern Finland at a depth of 2,5 km indicated a displacement of 1 mm and a source radius of 43 m (Saari 2008, Saari & Lakio 2008).

4.3 Postglacial faulting

During the last few decades, there have been discoveries of late Quaternary faults in the Precambrian bedrock of previously glaciated areas in Europe and North America. Especially in Fennoscandia the deglaciation phase has been characterized by intensive fracturing of the bedrock, neotectonic faulting and seismic activity, which has drastically changed the old concept of general stability of this intraplate continental area (e.g. Gregersen & Basham 1989).

Postglacial (PG) faulting can be defined as tectonically induced movements along distinct fracture surfaces or fault planes near former glacier margins in currently stable shield areas at the end of the last glaciation (e.g. Hasegawa & Basham 1989, Stewart & Hancock 1994, Olesen et al. 2004). Faults have been classified as postglacial based on the fact that they cut younger Quaternary deposits above the fault zones or cut bedrock surfaces polished by glacial ice (e.g. Kuivamäki & Vuorela 2002).

The terms ‘deglacial’ (e.g. Thorson 1996), ‘endglacial’ (e.g. Adams 2005) and ‘lateglacial’ (e.g. Sutinen et al. 2009a) have been used as synonyms for postglacial. Munier and Fenton (2004) have set forth the terms ‘glacio-isostatic faulting’ and ‘glacial rebound faulting’. Lund (2006b), Lund and Näslund (2009) and Poutanen et al. (2010) use the term ‘glacially induced faulting’ (GIF). Because the term postglacial faulting has become established in the published literature (e.g. Kukkonen et al. 2010a), it will be used in this study as a generic name to denote faulting that is related to the glacial loading/unloading.

In northern Fennoscandia there are a number of exceptionally large faults of early postglacial seismic activity (Fig. 14). Such faults were first identified by Kujansuu (1964) in Finland and independently by Lundqvist and Lagerbäck (1976) in Sweden. Since then these faults have been the target of active research, especially in connection with radioactive waste repository studies (Lagerbäck & Henkel 1977, Lagerbäck 1979, 1988, 1989, 1990, 1992, 2009, Lagerbäck & Witschard 1983, Kuivamäki & Vuorela 1985, 2002, Kukkonen & Kuivamäki 1985, 1986, Olesen 1985, 1988, Kuivamäki 1986, Talbot 1986, Henkel 1987, 1989, Vuorela et al. 1987, Wahlström et al. 1987, 1989, Kuivamäki et al. 1988, 1998, Bäckblom & Stanfors 1989, Muir Wood 1989, 1993a, Riad 1990, Olesen et al. 1992a, 1992b, 1995a, 1995b, 2000a, Ericsson & Stanfors 1993, Munier 1993, Stanfors & Ericsson 1993, Donner 1995, Arvidsson 1996, Roberts et al. 1997, Bungum & Lindholm 1997, Anon 1999, 2008b, Dehls et al. 2000a, 2000b, Lundqvist 2000, Gabrielsen et al. 2002, Munier & Fenton 2004, Mörner 2004, Lund 2006b, Paulamäki & Kuivamäki 2006, Sutinen et al. 2007b, 2009a, Lagerbäck & Sundh 2008, Lund & Näslund 2009, Juhlin et al. 2010, Poutanen et al. 2010, Juhlin & Lund 2011a, 2011b).

Largest PG faults are the Pasmajärvi, Ruostejärvi and Suasselkä-Aakenusjärvi faults in Finland, the Pärvie, Lansjärv and Lainio-Suijavaara faults in Sweden and the Stuoragurra fault in Norway (Fig. 14). The faults exist in a rather small area called the Lapland Fault Province (LFP; Muir Wood 1993a, Stephansson 1993, Olesen et al. 2000a). In addition, there are a couple of relatively large PG faults (Röjnoret, Burträsk) in Västerbotten, Sweden. Most of the postglacial faults are NE-SW trending reverse or thrust faults, which run perpendicular to the maximum horizontal stress orientation (Lagerbäck & Henkel 1977, Lagerbäck 1979, 1988, Lagerbäck & Witschard 1983, Talbot 1986, Olesen 1988, Talbot et al. 1989, Olesen et al. 1992a, Arvidsson 1996, Bungum & Lindholm 1997, Kuivamäki et al. 1998, Wahlström & Assinovskaya 1998, Muir-Wood 2000, Munier & Fenton 2004). The Nordmannvikdalen and Vaalajärvi faults are trending perpendicular to the reverse faults (Fig. 14). The former is a normal fault and the latter is a potential normal fault (Olesen et al. 2000c).

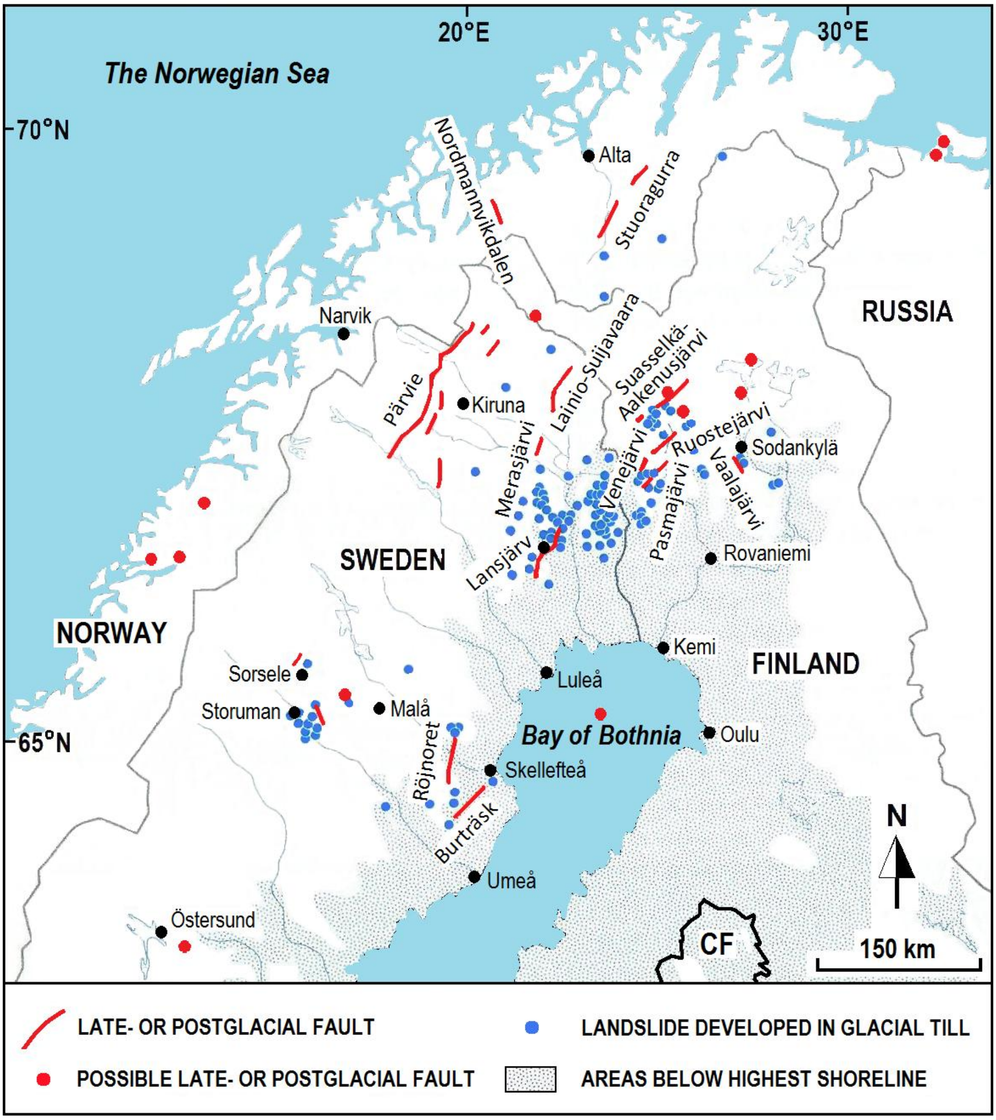

Fig. 14. Location of faults and landslide scars interpreted to be late- or postglacial in age in northern Fennoscandia. CF=Central Finland. Modified from Lagerbäck and Sundh (2008).

The length and scarp height of the postglacial faults may vary 50-150 km and 5-35 m, respectively (Fig. 15). Most often the scarp height is constant over long distances. The aspect ratio (scarp height to fault scarp length) for postglacial faults is often less than 1:10 000, i.e. less than that for most tectonic reverse faults (Scholtz 1990) but still within the range of offsets (Bonilla et al. 1984). The slip also scales well within the scaling laws of large earthquakes (Munier & Fenton 2004). Only vertical/dip-slip displacements have occurred, horizontal component is normally absent (Lagerbäck & Witschard 1983). The spacing between long faults is of the order of about 100 km (Munier & Fenton 2004). According to Muir Wood (1989) and Arvidsson (1996), the dimensions of the largest PG faults indicate that they reach deep (even 30…40 km) into the crust.

(A)

(B)

Fig. 15. A) Oblique aerial photograph of the Pärvie postglacial fault tracing across the tundra at Tjuonajåkk near Gällivare, Sweden. Photo taken from the southwest. B) Ca. 15 m high fault escarpment of the Pärvie fault in the same area. Photo taken from the northeast. Photographs adopted from Karow (2009); original photos Robert Lagerbäck (SGU).

The field evidence shows that the postglacial faults, at least where outcropping, dip between the vertical and some 30° (Kuivamäki et al. 1998). Geophysical and seismotectonic investigations indicate that the dip of some faults is not constant but becomes gentler with increasing depth (Paananen 1989, Olesen et al. 1991, Kuivamäki & Vuorela 2002). That is, some PG faults seem to represent the surface expressions of deeper listric faults (Lagerbäck 1979, Talbot 1986, Henkel 1987, Grant 1993, Kuivamäki et al. 1998, Anon 1999, Paulamäki & Kuivamäki 2006). On the other hand, drilling to the depth of 180 m through the Pasmajärvi fault showed a constant dip angle of 47º (Kuivamäki & Vuorela 2002) and did not indicate any turning of the fault plane to a more horizontal position with depth. The reflection seismic results of Juhlin et al. (2010) and Juhlin and Lund (2011a) suggest that the PG faults dip at a relatively high angle (50…75°) to depths of at least 2-3 km, and possibly deeper (see also Kukkonen et al. 2010a, Lindblom et al. 2011).

Most often the PG faults are the result of reactivation of previously existing fracture zones in a narrow range of azimuth (Lagerbäck & Henkel 1977, Kukkonen & Kuivamäki 1985, Paananen 1987, 1989, Vuorela et al. 1987, Lagerbäck 1988, Vuorela 1990, 1993, Olesen et al. 1991, Muir-Wood 1993c, Veriö et al. 1993, Munier & Fenton 1994, Kuivamäki et al. 1998, Kuivamäki & Vuorela 2002, Bungum et al. 2005, Anon 2008b, Knutsson 2008). Occasionally fracturing has, however, occurred also in relatively unaltered rocks (Henkel et al. 1983, Lagerbäck & Witschard 1983, Olesen 1988, Mörner 1993, Nisca 1995). Often the orientations of the faults are clearly influenced by the structures of the bedrock, e.g. gneissosity and boundaries between different rock units (Lagerbäck & Henkel 1977).

Many of the NE-SW orientated PG faults are situated between ancient NW-SE strike-slip faults suggesting a possible linked system of dislocations and simultaneous activity along these two orthogonal faults (Henkel 1979, 1987, 1989, Henkel et al. 1983, Kukkonen & Kuivamäki 1985, 1986, Kuivamäki 1986, Talbot 1986, Lagerbäck 1988, Talbot et al. 1989, Vuorela 1990, 1993, Kiviniemi et al. 1992, Kuivamäki & Vuorela 1994, Kuivamäki et al. 1998, Park 2000).

Postglacial faults have been interpreted as ruptures of single large (even M 7 - M 8) earthquakes (Muir Wood 1989, Johnston 1993, 1996, Arvidsson 1996, Kuivamäki et al. 1998, Olesen et al. 2000a). The scarps were most probably developed as a one-step event

(Lagerbäck 1989, Anda et al. 2002) or at least during a comparatively short time span (Lagerbäck & Witschard 1983). Anyhow, these deglacial movements have been, from a geological point of view, very rapid and, as a tectonic phenomenon, they have to be regarded as dramatic (Lagerbäck & Witschard 1983). Creep movement over a long period of time is definitively ruled out (Lagerbäck 1990). For example, the trenching of the Stuoragurra Fault in Masi, Norway, has revealed that most of the 7 m high scarp was formed in one seismic event (M 7,4 - M 7,7) during the very last part of the last deglaciation in Finnmark 9,300 years BP or shortly afterwards (Olesen et al. 2004). Kuivamäki et al. (1998) have estimated that the earthquake magnitudes of Finnish PG faults are from 5,5 to 7,0.

According to Muir Wood (1993a), it is likely that the strong seismicity pulse in the deglaciation phase lasted for some centuries, but probably less than 1,000 years. Bungum et al. (2005) keep it possible that the pulse could have lasted one or two millenniums. Saari (2000) proposes that postglacial earthquakes occurred immediately after deglaciation 10,000-8,000 years BP. According to Lagerbäck and Witschard (1983), the Pärvie fault has formed in direct connection to the deglaciation of the area. In some places it has developed before the local deglaciation, and in other places after it. The models of Wu et al. (1999) predict that the onset of postglacial activity in Fennoscandia was around 11-9 ka BP, prior to complete deglaciation, and maximum instability was attained around 10-7 ka BP. Then, from a geological perspective, the PG faulting is recent (Lagerbäck 1990).

In addition to the PG faults, several landslides have been detected in northern Fennoscandia (Fig. 16; Kujansuu 1972, Lagerbäck & Witschard 1983, Lagerbäck 1988, 1990, 1991, 1992, Sutinen 1992, Anon 1999, Olesen et al. 2004, Lagerbäck & Sundh 2008). The geographic and topographic location of these landslides suggests that there is a causal connection between them and postglacial earthquakes and faults (Fig. 14; Kujansuu 1972, Lagerbäck & Witschard 1983, Lagerbäck 1991). According to Sutinen (2005), only a part of landslides were formed in the proximity of retreating ice sheet, but a substantial number of slides occurred during the next 2,000 years after the ice disappearance (see also Vuorela 1990). Therefore, the postglacial faults may have been active enough to trigger landslides for a longer period of time since large-magnitude events soon after deglaciation 9,000-10,000 years ago. Recently, Sutinen et al. (2007a, 2007b, 2008, 2009a, 2009b) have linked sheetflow drainage of former subglacial lakes and subglacial evolution of esker networks and possibly even hummocky moraines to neotectonic fault instability in Finnish Lapland. Sutinen and Hyvönen (2010) have used a 15-cm-resolution LIDAR-data to recognize previously unnoticed paleo-landslides and possibly a new fault scarp near the Suasselkä postglacial fault in Kittilä, Finnish Lapland. In western Norway there are several large rock avalanches occurring in relatively gentle dipping terrain, which have been suggested as indications of large postglacial earthquakes (Olesen et al. 2000a).

Different types of sediment deformation (seismites), interpreted as being seismically induced, have been found in the Lansjärv area in northern Sweden (Lagerbäck 1988, 1991, 1992, 1994, Bäckblom & Stanfors 1989, Hansbo 1993, Lagerbäck & Sundh 2008) and in Balsfjord, northern Norway (Forwick & Vorren 2002). Deformation has proved to be very common in sediments deposited, or already in existence, during the deglaciation phase or shortly after whilst almost no deformation is found in younger deposits. This agrees well with the concept of a short-lived early postglacial co-seismic faulting (Lagerbäck 1991, 1992, 1994).

The PG faults in northern Fennoscandia are still the loci of strain relaxation, but their activity is minor. During the last decades, the improved detection and location capabilities in earthquake recording have been able to show that the epicenters of the earthquakes seem to be located mainly on the SE side of postglacial fault lines and to form parallel zones with fault lines (Kulhánek 1989, Arvidsson 1996, Bungum & Lindholm 1997, Lund et al. 2004, Lagerbäck & Sundh 2008, Juhlin et al. 2010, Lindblom et al. 2011). For instance, small earthquakes including a M_L 3,9 earthquake near the Stuoragurra fault on 21 January 1996, have been recorded within a 30 km wide zone southeast of and parallel to the fault scarp (Tvedt et al. 2002). These observations indicate that postglacial fault lines are surface expressions of old, reactivated, and still active regional fracture zones, which most frequently dip to the SE (Lagerbäck & Henkel 1977, Olesen 1985, 1988, Bungum & Lindholm 1997, Kuivamäki & Vuorela 2002, Ojala et al. 2004).

Fig. 16. Landslide scar at Lompolonvaara, some 20 km south-southwest of Pajala, northern Sweden. Photo adopted from Lagerbäck and Sund (2008), original photo Robert Lagerbäck.

Despite the (micro)seismic activity, there is no surficial evidence for any of postglacial faults having been reactivated by later seismic events (Talbot 1986, Lagerbäck 1989, Grant 1993, Kuivamäki & Vuorela 2002, Lagerbäck & Sundh 2008). To the contrary, the conditions at the fault scarps seem to be consolidated (Lagerbäck 1979). Geodetic measurements made in the Pasmajärvi and Nuottavaara postglacial faults in Finnish Lapland have either not yielded any significant motions of bedrock (Paananen 1987, Vuorela et al. 1987, Kiviniemi et al. 1992, Poutanen & Ollikainen 1995, Takalo et al. 2004).

According to Ahlbom et al. (1991b), it is quite unlikely that Lapland is the only region in Fennoscandia that has been subject to large deglaciation-related earthquakes. Also Bungum et al. (2005) keep it likely that similar reactivation of faults during deglaciation have occurred in other parts of Sweden and Finland. Tanner (1930) proposed that there ought to be postglacial faults everywhere in the Baltic Shield. Some authors, however, think that the Lapland Fault Province with its large PG faults is anomalous to the rest of Fennoscandia (e.g. Stephansson 1993). According to Lagerbäck and Sundh (2008), no conclusive evidence for late or postglacial faulting has been found south of Västerbotten County, i.e. below the 64°N latitude, in Sweden. If present, however, late- or postglacial faulting is thought to have been of minor magnitude as compared to the northern Sweden (Lagerbäck & Sundh 2008). The situation appears to be similar in Norway (Olesen et al. 2004, 2010) and in Finland (Kuivamäki et al. 1998), where distinct postglacial faults so far identified are concentrated to the northern parts of the countries. Anda et al. (2002) and Blikra et al. (2002) describe the postglacial Berill Fault from southern Norway and Helle et al. (2007) report on a possible NE-SW trending neotectonic fault zone between Bergen and Voss in Norway. These faults have, however, later discarded from the catalogue of Norwegian postglacial faults (Olesen et al. 2010). The postglacial fault candidates in Russian Karelia (Lukashov 1995) have neither been confirmed so far (Vuorela & Kuivamäki 1997).

Instead of large PG-faults, small faults cutting polished and washed bedrock outcrops have been found at several places in southern Finland (Askola, Dragsfjärd, Hiittinen, Ilomantsi, Kustavi, Oravi). The vertical dimensions of those scarps vary from some millimeters to one meter, and the length from some meters to tens of meters (Fig. 17; Edelman 1949, Tynni 1966, Nenonen & Huhta 1993, Sorjonen-Ward 1993, Kuivamäki & Vuorela 1994, Kuivamäki et al. 1998). According to Hudson and Cosgrove (2006), in Olkiluoto, southwestern Finland, geological evidence exists that small-scale reactivation of some fractures has occurred during the most recent glaciation. To the present author's best knowledge, postglacial faults of any kinds have so far not been detected in Central Finland.

Fig. 17. Part of the PG fault in Ilomantsi, southeastern Finland. A swarm of small postglacial fault scarps are cutting an ice-polished bedrock surface. For a detailed report see Nenonen and Huhta (1993) and Sorjonen-Ward (1993). Photo adopted from Kuivamäki et al. (1998); original photo Aimo Kuivamäki.

In the Stockholm region, Sweden, Mörner et al. (1989) and Mörner and Tröften (1993) report on numerous short NE-SW oriented open (reverse) faults and fractures, which have been reactivated in postglacial time. Also Grant (1993) was convinced that some bedrock structures around Stockholm were postglacial faults. Tröften (1997) describes a few NE-SW –trending faults around Stockholm area as possible postglacial faults; no large-scale primary faults were detected. Nisca (1995) sets forth a list of research articles from the period 1916-1948 concerning possible postglacial faulting in southern Fennoscandia. Nisca states that postglacial faulting has not happened only in northern Fennoscandian Shield but has happened and also will happen in the whole shield area. In Sweden, Sjöberg (1994) has studied bedrock caves and fractured rock surfaces from the point of view of postglacial faulting. According to Systra and Spungin (2007), microseismic investigations undertaken in the Republic of Karelia, Russia, display an extensive range of postglacial fault motion and seismo-gravitational dislocation.

The fact that indisputable postglacial faults do not seem to occur, at least so frequently, in southern Fennoscandia may be due to following factors (Lagerbäck 1979): denser afforestation, a frequent rough relief, human activity, abrasion and sedimentation below the highest shore level. Most postglacial faults in northern Fennoscandia seem to occur in the central part of the former ice sheet, where frozen-bed conditions prevailed and escaped erosion during the late Weichselian (Kleman et al. 1997 and their Fig. 10). Also the fact that the deglaciation and isostatic rebound were most rapid and ice load greatest and of longer duration in the north possibly explains why PG structures seem to be restricted to northern Fennoscandia (Lagerbäck 1990, Grant 1993, Muir Wood 1993a, Saari 2000, Anon 2006). Moreover, Talbot (1999) has proposed that the pulse of seismotectonic activity in the Lapland Fault Province may have coincided with the time when the Late Weichselian ice sheet in that region was of a size that coincided with the critical wavelength of the underlying lithosphere. Lund et al. (2009) and Lund and Schmidt (2011) argue that the deglacial tectonic (thrust) stress field could be the key to understanding why large postglacial faulting only occurred in northern Fennoscandia.

On the other hand, the absence of discovered PG faults in central and southern Fennoscandia is not a guarantee that such structures do not exist there (Lagerbäck 1979, Olesen et al. 2000a). According to Björck and Svensson (1992), small amplitude movements (1-2 m) will be almost impossible to find without using fieldwork. One possible explanation is that the faults are hidden, i.e. they either originate at depth and do not reach the surface or their movements are dissipated by ductile layers (Bungum et al. 2005).

Mörner and his co-workers in Sweden (Mörner & Tröften 1993, Tröften & Mörner 1997, Mörner et al. 2000, Mörner 2009) have set forth widespread contemporaneous soft-sediment deformation in varved sequences that appear to have been triggered by strong seismic shaking during the late- or postglacial period. Mörner (2004, 2005) lists indirect observations (shoreline bends, uplift irregularities), secondary structures (bedrock fracturing, bedrock caves, rock falls) and secondary effects (liquefaction, tsunamis, earth slides, turbidites) in Sweden, which, according to him, provide evidence of high amplitude palaeoseismic events in postglacial time and, especially, at the time of deglaciation (see also Björkman & Trägårdh 1982, Anundsen 1989, Lagerbäck 1990, Nisca 1995, Tröften 1997, 2000, Mörner 1996, 2001). According to Mörner (1997), the varve chronology may provide information about the exact age, the areal size of deformation and the recurrence time of multiple seismic events.

According to Lagerbäck (1990), the occurrence of contorted structures described by Aartolahti (1987) in glaciofluvial deposits in southern Finland may indicate that strong earthquakes, similar to the northern parts of the country, struck southern Finland at the time of deglaciation (see also Mohindra & Bagati 1996). Further evidence of palaeoseismic activity as soft-sediment deformation comes from offshore southwestern Finland (Kotilainen & Hutri 2004, Hutri 2007, Hutri & Kotilainen 2007, Hutri et al. 2007, Virtasalo et al. 2007, 2010). At Hunnenberg, southern Sweden, Björk and Digerfeldt (1982) describe a Late Weichselian shore displacement, which may be explained by postglacial faulting. On the other hand, Lagerbäck et al. (2005, 2006) did not found any fault dislocations or sediment distortions of tectonic origin in the Forsmark and Oskarshamn regions in Sweden. To the contrary, minor sheers were generally the result of glacial plucking, governed by vertical joints and sheeting, and sediment disturbances were interpreted as caused by sliding.

In the rest of Europe there are a few observations on PG features. Along the southwestern coast of the Baltic Sea, NE Germany, Hoffmann and Reicherter (2011) have found soft-sediment deformation (liquefaction, slumping, faulting) most probably caused by earthquake-induced shaking immediately after deglaciation. In western Scotland, late to postglacial reactivation has resulted in movement on the Kinloch Hourn Fault (Firth & Stewart 2000, Stewart et al. 2001) and in palaeo-liquefaction and slumping in postglacial sediments at Glen Roy and Loch Broom (Sissons & Cornish 1982, Ringrose 1989a, 1989b, Stoker & Bradwell 2009). In Northern Ireland, Knight (1999) suggests that the postglacial reactivation of old lineaments has generated normal faulting with meter-scale displacement in the southern Sperrin Mountains. Mohr (1986) has detected possible late-glacial displacements in Iar Connacht, Ireland. Increased postglacial tectonic activity has also been reported from the European Alps (e.g. Becker et al. 2005).

Postglacial faults in the Lapland Fault Province appear to have much larger displacements, and much greater lengths, than their North American equivalents (Adams 1989a, Munier & Fenton 2004). An inventory of postglacial faults in eastern and Arctic Canada (Fenton 1994a) reveals only two candidate structures that might possibly approach the scale of the Lapland faults: the Aspy Fault and the Peel Sound fault zone (Grant 1990, Dyke et al. 1991, Adams & Clague 1993). Be that as it may, Hunt and Malin (1998) suggest that similar postglacial earthquakes had occurred in northern Canada. In southeastern Canada, postglacial faults with throws >1 m have not been found so far (Mazurek 2004). In Washington, USA, the postglacial reactivation of the Saddle Mountain West fault has generated reverse faulting with a meter-scale displacement (Witter & Givler 2007). According to Adams (2006), the Canadian craton might behave differently from the Scandinavian craton due to the scale of its glaciation, or due to all the deglacial scarps forming under the ice and thus rapidly destroyed, as might have been the case in southern Fennoscandia, too.

Adams (1981) gives a discussion of possible postglacial faulting and an annotated bibliography of 54 references to postglacial faulting in eastern Canada and adjacent parts of the United States. These faults have minute throws (a few tens of mm or less), are of the reverse (thrust) type and occur on bedding planes, cleavages, joints or other high-angle, preexisting planes of weakness roughly parallel to ice margins. These faults seem to suggest north-south compression, although they show a considerable scatter (Hasegawa & Adams 1981).

However, a number of these features were later understood to have resulted from other mechanisms. Such non-tectonic mechanisms are, for instance, gravity-induced faulting, frost heaving, glaciotectonics (e.g. ice push features), erosional scarps, overburden draping of underlying bedrock features and stress release features (Rast et al. 1979, Thompson 1981, Basham & Adams 1984, Barosh 1986, Kumarapeli 1987, Gorrell 1988, Adams 1989a, 1994, 1996, Broster & Burke 1990, Schroeder et al. 1990, Adams & Bell 1991, Adams & Clague 1993, Adams et al. 1993, Dredge 1993, Fenton 1994a, 2002, 2003, Firth & Stewart 2000, Stewart et al. 2000, Godin et al. 2002, Munier & Fenton 2004, Stewart & Jarman 2005).

Indeed, one should critically examine the claims for postglacial faulting in order to distinguish seismogenic deformation from deformation arising from non-tectonic mechanisms associated with glaciation. Criteria for identification of postglacial faulting have been presented by Mohr (1986), Adams et al. (1993), Muir-Wood (1993a), Fenton (1994a, 1994b), and Munier and Fenton (2004).

Muir Wood (1993a) criticizes Fennoscandian palaeoseismical studies, which have assumed a whole range of phenomena to be palaeoseismic indicators without any obvious connection with earthquakes. According to Muir Wood (1993a), bouldery tills, fragmented bedrock, boulder caves, varve sedimentation rates, position shift of eskers etc. should not be considered as seismic ground-motion indicators. Instead, they may have a number of other possible explanations (groundwater pressure changes, frost-shattering, rockslides). Some of the disrupted rock masses may, according to Muir Wood (1993a), reflect shallow stress relief phenomena. In Norway, field checking of neotectonic reports and claims have shown that the majority of them can be attributed to effects other than tectonic faulting (Olesen et al. 2003a, 2003b, 2010, 2011a, 2011b). Gravitational sliding and glacial erosion (plucking) have been the most important agents to form scarps that have been misinterpreted as postglacial faults (Olesen et al. 2000a). The influence of winter ice as well as water freezing in bedrock joints may have been underestimated, too (Lindberg 2007).

Recent stress-relief structures, such as axial fractures or borehole offsets, are relatively common along rock facies in Fennoscandia and central and northern North America (Block et al. 1979, Schäfer 1979, Bell 1985, Roberts 1991, 2000, Wallach et al. 1998, Roberts & Myrvang 2004, Pascal et al. 2005, 2006, 2007, 2010b). Axial fractures are vertical tension fractures produced by gas overpressure inside the drillhole when the blast occurs; their strike reflects the orientation of the ambient maximum horizontal stress axis. The reverse-slip direction associated with the borehole offsets is found to be strongly controlled by the maximum horizontal compressional stress acting on the rock mass. Typically the boreholes are offset by a few centimeters in a few years time (Block et al. 1979, Schäfer 1979, Bell 1985, Roberts 1991, 2000, Roberts & Myrvang 2004, Pascal et al. 2005, 2006). The displacements are thought to record release of tectonic stresses because the displacements are congruent with the present-day regional stress field (Bungum et al. 2010, Pascal et al. 2010b).

In eastern North America, to the contrary of Fennoscandia, a special kind of neotectonic shallow stress relief/release phenomena, i.e. pop-ups, quarry-floor buckles and folds, indicate ongoing crustal instability (Sbar & Sykes 1973, Saull & Williams 1974, Adams 1982, Williams et al. 1985, Wallach & Chagnon 1990, Martini & Bowlby 1991, Wallach 1992b, Engelder 1993, Karrow 1993, McFall 1993, Rutty & Cruden 1993, Stewart & Hancock 1994, Wallach et al. 1998, Muir-Wood 2000, Karrow & White 2002, Jacobi et al. 2007). These features are small, generally from 10 to 3,000 m long and less than 2 m high, surface folds in bedrock (Adams 1989a). They develop in response and perpendicular to maximum horizontal compressive stress and are present both in open fields and on quarry floors. They have sometimes misinterpreted as postglacial or tectonic faults (Fenton 1994a). The presence of favorably oriented preexisting horizontal planes of weakness is a prerequisite for pop-ups to nucleate and develop (e.g. Saull & Williams 1974, Wallach 1992a). All the pop-ups occur within the Paleozoic sedimentary cover rocks; none has been found on the Precambrian shield itself (Hasegawa & Adams 1981, Jacobi et al. 2007).

Although much effort has been spent on investigating the PG faults with both geological and geophysical methods, key questions concerning the formation and current status of the faults are still largely unresolved. These include, for example, fault geometry at depth, fault strength and current deformation rates as well as their hydrogeology (Lund 2006b, Lorenz & the SDDP Working Group 2009, Anon 2010, Kukkonen et al. 2010a, 2010b, 2010c, 2011a, 2011b, Postglacial Fault Drilling Project http://www.sddp.se/PFDP).

4.4 Glacial isostatic adjustment (postglacial rebound)

Glacial isostatic adjustment (GIA) or postglacial rebound is the response of the solid Earth to the changing surface load brought about by the waxing and waning of large-scale ice sheets and glaciers (e.g. Fjeldskaar et al. 2000, Watts 2001, Sella et al. 2007). The GIA is commonly accepted as an isostatic readjustment of the crust and the upper mantle towards equilibrium (e.g. Cai & Grafarend 2007).

The postglacial rebound of Fennoscandia has been extensively studied during the last two hundred years; see Ekman (1991) for a historical review. The uplift pattern has been constructed on the basis of sea-level records, lake-level records, gravity data and repeated high-precision levellings (Suutarinen 1983, Ekman 1985, 1993, 1996, Kakkuri & Vermeer 1985, Kakkuri 1987, Björck & Svensson 1992, Ekman & Mäkinen 1996, Mäkinen & Saaranen 1998, Mäkinen 2000, Danielsen 2001, Saaranen & Mäkinen 2001, 2002, Mäkinen et al. 2002, Lambeck & Purcell 2003, Vestøl 2006, Steffen & Wu 2011).

Many space geodetic techniques have the sensitivity to recover not only vertical but also horizontal deformation components due to postglacial rebound (James & Lambert 1993, Mitrovica et al. 1993, Scherneck et al. 1996, 2001, 2009, Peltier 1998, Argus et al. 1999, Campbell & Nothnagel 2000, Haas et al. 2000, 2002a, 2002b, 2003, Koivula & Poutanen 2001, Johansson et al. 2002, Koivula et al. 2004, Milne et al. 2004, Marotta et al. 2004, Argus & Peltier 2005, Lidberg et al. 2010, Poutanen & Ivins 2010, Poutanen et al. 2010). Initiated in 1993, the BIFROST Project (Baseline Inferences for Fennoscandian Rebound Observations, Sea level, and Tectonics; Johansson et al. 2002) constructed a network of permanent global positioning system (GPS) receivers throughout Finland and Sweden to observe the regional deformation field (Ollikainen et al. 2003, Poutanen et al. 2003a, 2003b, Cai & Grafarend 2007).

In determining the rate of recent Fennoscandian uplift—based on geodetic observations—the uplift of the crust relative to mean sea-level is called the apparent land uplift, which is at its highest around 9 mm a^{-1} in the northern part of the Gulf of Bothnia near the center of the former ice sheet (Fig. 18; Kakkuri 1987, Ekman & Mäkinen 1996, Gudmundsson 1999, Fjeldskaar et al. 2000, Danielsen 2001, Milne et al. 2001). At present, Fennoscandia shows a dome-like uplift with the ellipsoidal form of the uplift isolines having the longer axis in the SW-NE direction (e.g. Nikonov 1980, Vestøl 2006). There are local disturbances in the uplift pattern (Kakkuri & Chen 1992, Veriö et al. 1993, Lehmuskoski 1996, Fjeldskaar et al. 2000).

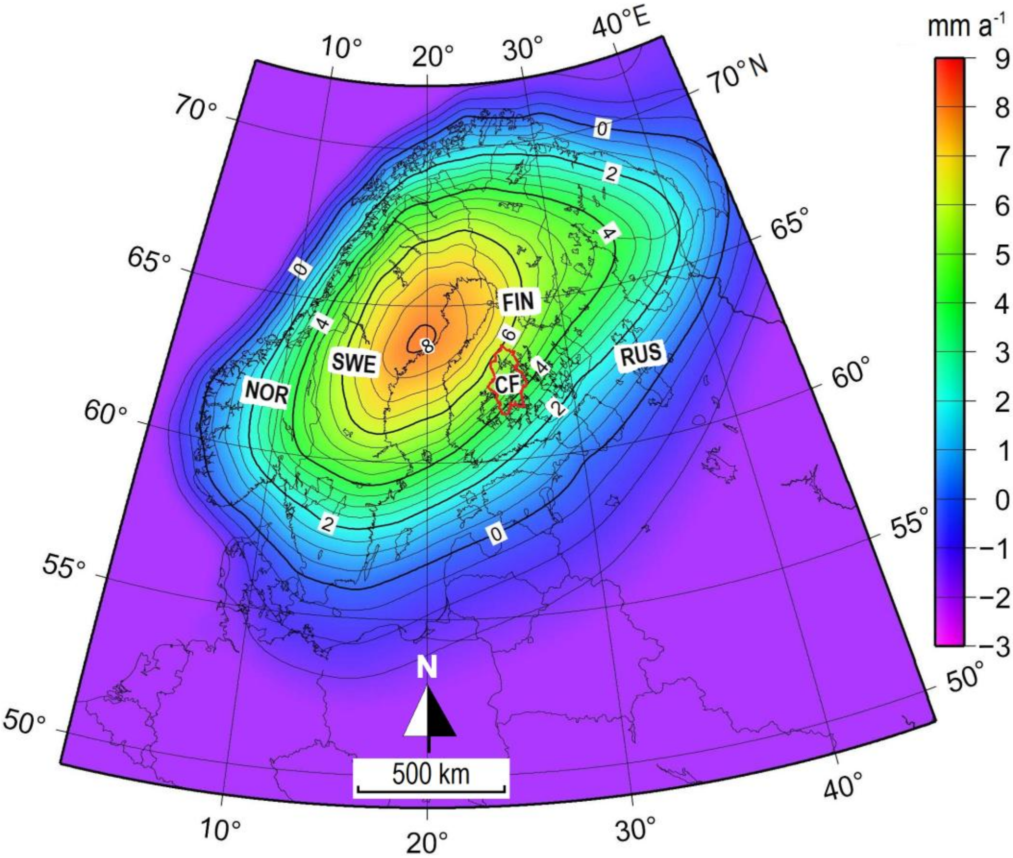

Fig. 18. Apparent uplift (mm a^{-1}) for the final land uplift model NKG2005LU (RH 2000 LU) of Fennoscandia. NOR=Norway, SWE=Sweden, FIN=Finland, RUS=Russia. Central Finland (CF) is outlined in red. Modified from Ågren and Svensson (2007).

During the Quaternary, ice sheets have covered Fennoscandia several times during the last 2-3 million years. The peak of the Weichselian glaciation occurred around 20,000-18,000 years BP (Donner 1995). The ice sheet caused a considerable depression of the Fennoscandian lithosphere. The maximum thickness of the ice sheet was of the order 2,5-3 km, with a crustal downwarping around 1/3 of this (Saari 1992).

At 16,000 years BP the most depressed area was situated on the Scandinavian Highlands about 200 km to the north of Oslo (Påsse 2001). Weichselian deglaciation began about 12,000 years BP in the Scandinavian mainland. At the same time began a fast successive change of the position of the most depressed area from the Highlands to Bothnian Bay (Påsse 2001). At 8,500 years BP most of Fennoscandia was ice-free (Donner 1995, Lundqvist & Saarnisto 1995). The associated glacio-isostatic recovery caused the Fennoscandian land area to rise, which is continuing.

The centre of uplift started to rise at about 13,000 years BP, i.e. long before the local deglaciation (Mörner 1980b, Påsse 1997). The absolute rates of uplift peaked at around the deglaciation and/or the end of the Younger Dryas Stadial and amounted 50-500 mm a^{-1} (Mörner 1977, 1978, 1979b, 1980b, 1981, 2004, Saarnisto 1981, Glückert 1989, 1994, Björck & Svensson 1992, Glückert et al. 1993, Donner 1995). This can be compared with the uplift values of greater than 10 m/100 yr (i.e. >100 mm a^{-1}), which have been derived from the postglacial delevelling curves at sites within Hudson Bay, Canada (Andrews 1991). The rate of land uplift decreased significantly ca. 8,500–8,000 years BP (9,500–8,800 calibrated years BP; Ristaniemi et al. 1997). The total absolute uplift is 830 m in the Kvarken region, "throat" of the Gulf of Bothnia (Mörner 1979a, 1980b). It is generally accepted that there is a good correlation between the total postglacial uplift and the present uplift rates in Fennoscandia (e.g. Mörner 1980a).

In the center of the rebound area vertical (radial) velocity is at its highest and horizontal (tangential) velocity is close to zero. Vertical velocities decrease radially outwards from the center of rebound whereas horizontal velocities increase and peak near the margins of former Fennoscandian ice sheet (Mitrovica et al. 1994, Scherneck et al. 1998, Haas et al. 2002b, Marotta et al. 2004, Milne et al. 2004, Koivula et al. 2006, Steffen et al. 2006). Peak horizontal velocities due to glacial rebound are observed to be ≤ 2 mm a^{-1} (James & Lambert 1993, Mitrovica et al. 1994, Argus et al. 1999, Haas et al. 2002b). Areas of subsidence within the forebulge, the height of which is not expected to have exceeded 150 m, surround the zone of rebound. Here the motion is downward at rates of some millimeters per year (Fig. 18; Ahlbom et al. 1991b, Scherneck et al. 1996).

Published maps of vertical and horizontal crustal rates in Fennoscandia (e.g. Ekman 1996, Milne et al. 2001) may indicate that land uplift has taken place under the last 10,000 years domically and without major irregularities. However, it was Härme (1966) among the first who suggested that the postglacial uplift does not take place as plastic warping, but as block movements, and mainly these movements adhere to the old fault lines. Different blocks are separated from each other by deep vertical to subvertical faults, some of which are reaching through the entire crust. Also nearly horizontal faults have been discovered at various depths of the crust (e.g. Anttila 1988, Saari 1992). The presence of fault blocks in a region undergoing postglacial rebound raises the possibility of differential block uplift and tilting (Koshechkin et al. 1975, Talvitie 1977a, Lagerbäck 1979, Hasegawa & Basham 1989).

The present vertical bedrock movements of Finland have been studied in several areas with leveling profiles crossing fault zones of different size categories (Veriö et al. 1993). Of the 53 profiles leveled, 28 recorded statistically significant local changes in elevation, while 7 showed variations that deviate substantially from predictions based on uniform plastic uplift (Veriö et al. 1993). Risberg et al. (2005) consider that the isolation of nine lake basins in southeastern Sweden has been the result of small-scale irregular uplift. On the other hand, no evidence for vertical displacement between bedrock blocks has been found in the Finnish shoreline displacement data (Ristaniemi 1987, Eronen & Ristaniemi 1992, Eronen 1994, Ristaniemi et al. 1997).

At present, it is widely accepted that land uplift in Fennoscandia is taking place on a regional scale plastically, but on a local scale as differential block movements preferentially within zones of pre-existing weakness (Talvitie 1977a, Björkman & Trägårdh 1982, Koskiahde et al. 1985, Mörner 1985, 2004, Niini 1987, Anundsen 1989, Tirén & Beckholm 1989, Saari 1992, Muir Wood 1993a, Veriö et al. 1993, Kuivamäki & Vuorela 1994, 2002, Lehmuskoski 1996, Kuivamäki et al. 1998, Ojala et al. 2004, 2006, Olesen et al. 2004, Paulamäki & Kuivamäki 2006, Miettinen et al. 2007, Kollo & Vermeer 2009).

The isostatic recovery will continue yet for several thousands of years, even though there are uncertainties in the calculated rate of residual uplift (Eronen et al. 2001). The remaining uplift is estimated to be from 30 m to 200 m (Bjerhammar 1977, 1980, Balling 1980, Ekman 1989, Stephansson 1989, Björck & Svensson 1992), though the most recent calculations suggest an amount of ca. 90 m for residual uplift (Ekman & Mäkinen 1996, Påsse 1996, 2001, Kaufmann et al. 2000).

The project DynaQlim aims to integrate existing data and models on GIA processes, including both geological and geodetic observations (Poutanen et al. 2007, Poutanen & The DynaQlim Group 2008, Poutanen 2010).

4.5 Plate tectonics or postglacial rebound causing earthquakes

The relative importance of plate tectonics and postglacial rebound in the generation of postglacial and current earthquakes in Fennoscandia and Laurentia has been and still is debated and the problems related to this subject are far from being solved (Båth 1978, 1984, Stein et al. 1979, Quinlan 1984, Ekman 1985, Hasegawa et al. 1985, Meyer & Ahjos 1985, Klein & Barr 1986, Basham & Gregersen 1989, Davenport et al. 1989, Hasegawa & Basham 1989, Slunga 1989c, Skordas 1992, Muir Wood 1993a, Vuorela 1993, Wahlström 1993, Wu & Hasegawa 1996a, Kakkuri 1997, Klemann & Wolf 1998, Plag et al. 1998, Stewart et al. 2000, Hicks & Ottemöller 2001, Ojala et al. 2004, Mazzotti & Adams 2005, Poutanen & Ivins 2010, Poutanen et al. 2010, Carafa et al. 2011, Steffen et al. 2011).

There are geological and geophysical evidence that support tectonic stress as the dominant cause of current earthquakes in northern Europe and North America. First of all, the spatial distribution of recent earthquakes shows little correlation with the center of postglacial rebound in both areas (Talvitie 1977a, Adams & Basham 1989, Bungum 1989, Slunga 1989a, 1991, Wahlström 1989, Gregersen et al. 1991, Skordas et al. 1991, Ahjos & Uski 1992, Skordas & Kulhánek 1992, Zoback 1992a, 1992b, Wu & Hasegawa 1996b, Wu et al. 1999, Byrkjeland et al. 2000, Olesen et al. 2000a, Gregersen 2002, 2006, 2008, Uski et al. 2003a). Also the (micro)seismic epicenters clustered parallel to the postglacial faults in northern Fennoscandia (Lund et al. 2004, Lagerbäck & Sundh 2008, Juhlin & Lund 2011a) suggest that the cause of present-day earthquakes is plate tectonics.

Secondly, the orientation of the contemporary stress field does not appear to be dominated by the effects of past glaciation, i.e. rebound stress has little influence on contemporary stress orientation (Talvitie 1977a, Bungum & Fyen 1980, Slunga 1981a, 1981b, Slunga et al. 1984, Koskiahde et al. 1985, Slunga & Ahjos 1986, Marrow & Walker 1988, Adams 1989b, Clauss et al. 1989, Gregersen & Basham 1989, Stephansson 1989, Adams & Bell 1991, Bungum et al. 1991, Gregersen et al. 1991, Gregersen 1992, 1993, 2002, 2006, 2008, Müller et al. 1992, Zoback 1992a, 1992b, Johnston 1993, Assameur & Mareschal 1995, Wu et al. 1999, Lindholm & Bungum 2000, Munier & Fenton 2004, Pascal et al. 2005, 2006, Bödvarsson et al. 2006, Smelror et al. 2007, Gregersen & Voss 2010).

Yet quite often both tectonic forces and rebound stress have been needed to explain the distribution and style of contemporary earthquakes in both continents (Arvidsson et al. 1992, Wahlström 1993, Arvidsson & Kulhánek 1994, Ericsson et al. 1996, Stewart et al. 2000, Saari & Uski 2002). Also other mechanisms have been suggested (e.g. flexural stresses from sediment loading, differences in lateral lithospheric density and elevation; Bungum et al. 1991, Byrkjeland et al. 2000, Fejerskov & Lindholm 2000, Fjeldskaar et al. 2000, Atakan & Ojeda 2005, Pascal & Cloetingh 2009). According to Påsse (2001), instead of the amount of crustal uplift or subsidence, it is the tilting direction in and between the rock blocks that is assumed to be the most important factor for creating glacio-isostatic induced seismic activity.

Early investigations (Walcott 1970, Stein et al. 1979) found that postglacial rebound stresses alone could be responsible for the mode of earthquake failure. According to Sykes (1978) and Hasegawa (1988), rebound stress is probably not large enough to initiate rock fracture in intact rocks, but can reactivate preexisting faults. Quinlan (1984) has argued that postglacial rebound is capable of triggering earthquakes in prestressed regions but rarely capable of dictating the focal mechanism of these earthquakes. According to Stephansson (1988), the stresses in the Fennoscandian Shield are mainly due to three mechanisms: 1) residual stress left from the isostatic rebound after deglaciation, 2) concentration of the stress due to creep, and 3) stress caused by the spreading of the North Atlantic Ridge. Ojala et al. (2004) state that in Fennoscandia, for the current level of seismicity though low and heterogeneously distributed, the glacial isostatic adjustment has a very important role, because the brittle crust is near the point of failure, and, consequently, small changes, like glacial rebound, can nucleate earthquakes within optimally oriented pre-existing weaknesses. According to Lambeck (2005), tectonic stresses now dominate the largely relaxed glacial-load stresses. Latest analyses indicate that strong ridge-push forces characterize the Baltic and Canadian shields and that present-day stress relief, in Fennoscandia and northeastern America, is mostly triggered by plate-scale ridge-push forces and not by residual glacial loading stresses (Bungum et al. 2010, Pascal et al. 2010a, 2010b, 2010d).

By means of GPS measurements several authors have concluded that the Fennoscandian and Laurentian deglaciation centers are subjected to present-day extension with relatively small tangential deformation rates and encircled by a region in which the tangential motion is larger and radiated outwards (e.g. Scherneck et al. 2001, Johansson et al. 2002, Latychev et al. 2005, Calais et al. 2006, Koivula et al. 2006, Sella et al. 2007, Klemann et al. 2008). However, as Bungum et al. (2010) and Pascal et al. (2010b) state, these horizontal uplift motion patterns should not be interpreted as diagnostics for extensional stresses, although some authors have suggested such stresses due to the postglacial uplift (e.g. Stein et al. 1989, Arvidsson & Kulhanek 1994, Gudmundsson 1999). From the stress point of view, the 'outward' horizontal velocities are an apparent phenomenon driven in large part by the bending of the elastic lithosphere. That is, due to the curvature of the Earth, even purely radial motion (i.e., no tangential components) will increase the lengths of the vectors between the stations used for distance measurements (Mitrovica et al. 2001, Mäkinen et al. 2002). Other reasons might be, for example, non-symmetric surface mass loads and in the Fennoscandian Shield long-wavelength deformation from the deglaciation of Laurentian ice sheet (Milne et al. 2001, Whitehouse 2009).

Instead of tangential extension, a consequence of glacio-isostatic uplift is horizontal compression in combination with the ambient tectonic stresses causing shortening, which is likely to be relieved by strike-slip and thrust (reverse) faulting on existing structures (e.g. Adams & Clague 1993). Furthermore, the rebound results in a gradual decay of compressional glacial stresses (i.e. a slow return to the pristine preglacial loading conditions), in agreement with viscous properties of the lithosphere and ice histories (Gregersen et al.

1991, Wu & Hasegawa 1996b, Wu et al. 1999, Turpeinen et al. 2008, Hampel et al. 2010a, Pascal et al. 2010b).

The timing of faulting and paleoseismicity in Fennoscandia correlates well with the end of deglaciation, which may indicate that glacial stresses may have played a more important role in earthquake generation during early postglacial time (e.g. Muir Wood 1989, 2000). The transitions between glacial and interglacial periods have been associated with considerable changes in the mass distributed on Earth´s surface in the form of ice sheets, glaciers and lakes and owing to the rise and fall of the sea level (Hampel & Hetzel 2006, Hetzel & Hampel 2006, Karow & Hampel 2009, Poutanen & Ivins 2010). Such climate-controlled mass fluctuations have affected the lithosphere, which has responded by flexure and rebound to the addition and removal of mass on its surface, respectively, and by coincident stress perturbations (Walcott 1970, Cohen 1993, Thorson 1996, 2000, Wu et al. 1999, Muir Wood 2000, Sauber et al. 2000, Stewart et al. 2000, Wu & Johnston 2000, Watts 2001).

The resulting changes in the ambient plate tectonic (background) stress state have considered to be caused by the process of glacial loading and unloading itself, which primarily affects the vertical stress (Johnston 1987, 1989a, 1989b, Wu & Hasegawa 1996a, 1996b, Sauber & Molnia 2004), and by the glacially induced flexure and rebound, which mainly alter the horizontal stress (Watts 2001, Hetzel & Hampel 2005, Hampel & Hetzel 2006, Turpeinen et al. 2008, Hampel et al. 2009, 2010a, 2010b, Karow 2009, Lund et al. 2009). Pascal et al. (2010c) have modeled the late glacial stress situation (i.e. 10 ka BP) in the centre of the Fennoscandian Shield, where surface topography had been previously depressed up to ca. 1 km by glacial loading. Their modeling predicts that ridge push stresses were up to three times larger than today.

Fault stability during a glacial cycle has been modeled with varying degrees of model complexity during the last decades. Early models (Walcott 1970, Stein et al. 1979, Quinlan 1984), which did not consider a viscous mantle and, therefore, did not capture stress relaxation effects (Lund 2006b), generally agreed on increased fault stability during the loading phase of the continental ice sheet and then increased fault instability during deglaciation.

Sykes (1978) and Johnston (1987, 1989a, 1989b, 1993) pointed out that large (M>5) earthquakes are virtually absent beneath modern ice sheets in Antarctica and Greenland, and that background seismicity is lower than expected relative to comparable, but unglaciated, intraplate settings (e.g. Gregersen 1989, Mäntyniemi et al. 2004, Voss et al. 2007). Johnston (1987) proposed that the growth of continental ice sheets, by increasing the vertical minimum principal stress σ_3 and hence reducing the differential stress between the maximum horizontal principal stress σ_1 and σ_3 (Chapter 4.1.3), keeps these regions stabilized, with the result that faulting is suppressed. Johnston (1987) also predicted that accumulated stress from glacial periods would be released during deglaciation. Although being as visionary as they are, Johnston´s analyses omitted, however, the effect of flexural stresses.

In order to evaluate the stress state resulting from plate tectonics and both from glacial loading and unloading as well as from lithosphere flexure and rebound, various numerical models have later been developed, which also account for the behavior of the viscoelastic crust and sublithosphere mantle (Wu & Hasegawa 1996a, 1996b, Wu 1997, Johnston et al. 1998, Klemann & Wolf 1998, Wu et al. 1999, Sauber et al. 2000, Wu & Johnston 2000, Ivins et al. 2003, Lambeck & Purcell 2003, Lund & Zoback 2003, 2007, Sauber & Molnia 2004, Hetzel & Hampel 2005, Lund 2005a, 2005b, 2006a, 2006b, Hampel & Hetzel 2006, Hampel et al. 2007, 2009, 2010a, 2010b, Turpeinen et al. 2008, Karow 2009, Karow & Hampel 2009, Lund & Näslund 2009, Lund et al. 2009, 2011, Lund & Schmidt 2011). These models use the difference in the fault stability margin, dFSM, or the difference in the Coulomb failure stress, dCFS, as a stability criterion based on the Mohr-Coulomb theory (Chapter 4.1.3).

Generally these models confirm that earthquake activity is suppressed by the emplacement of the large ice sheets but greatly enhanced at the end of deglaciation. In addition, the models show that the onset and location of fault instability varies with the ice sheet dimension and temporal evolution, lithospheric and mantle structure and the initial state of stress. However, depending on the selection of time for isostatic equilibrium, either before the onset of glaciations or at the glacial maximum, quite contradictory modes of faulting have been modeled (Bungum et al. 2010, Kukkonen et al. 2010a).

In addition to the induced horizontal and vertical stresses, the ice sheet will increase the water pressure in the rock below (e.g. Muir Wood 1993b). Pore pressure is an important parameter in the assessment of fault stability as the fluid pressure acts to decrease the effective normal stress on the faults (Chapter 4.1.3). In a similar way, as the height of the gas column increases in a hydrocarbon reservoir, at some point the pore pressure will be sufficient to induce fault slip, providing a mechanism to increase fault permeability and allow leakage from the reservoir (Wiprut & Zoback 2000, Wiprut 2001).

The development of fluid overpressures probably has been a contributory factor in triggering the postglacial faulting (Muir Wood 1989, 1993a, Lagerbäck & Sundh 2008, Vidstrand et al. 2008, Lund & Näslund 2009). The ice sheet is expected to increase pore pressure in the crust below, both through the increase in mean stress by the load itself and through high water pressures at the base of the ice sheet (Näslund 2006). The basal conditions of the ice sheet, the ice sheet thickness and the characteristics of the hydrological system of the ice sheet determine how the fluid pressure at the base of the ice develops (Näslund 2006). The simulations made by Lund et al. (2009) and Lund and Schmidt (2011) show a very strong dependence of fault stability on the glacially induced excess pore pressures. It is probable, too, that high glacial pore pressures would be transmitted down to seismogenic depths during the time of a glaciation (Lund et al. 2009, Lund & Schmidt 2011).

The last Weichselian ice sheet apparently was cold-based throughout most of its existence in northern Fennoscandia (Kleman et al. 1997, Kleman & Hättestrand 1999). It was especially beneath this cold-based ice sheet resting on a deeply frozen crystalline bedrock, where fluid pressures would have raised close to lithostatic because of no chance to be released through the overlying permafrost even at the edge of the ice sheet (Boulton & Caban 1995, Boulton et al. 1995, 1996, Milnes 2002, Lönnqvist & Hökmark 2010). Fault stability margins would have significantly been reduced under these raised fluid pressures, which could tentatively explain the particular circumstances of northern Fennoscandia for the generation of the postglacial episode of reactivation of pre-existing faults. Overpressured fluids are likely to play a role especially in the reactivation of steep reverse faults and other structures that are unfavorably oriented for reactivation (Sibson 1974, 1989, 1990, 2000, Chester et al. 1993, Ge & Garven 1994, Cox 1995, Ryberg & Fuis 1998, Sibson & Scott 1998, Eberhart-Phillips & Reyners 1999).

Hence, in combination with the tectonic background and glacially induced stresses, increased pore pressure may have caused the rock to fail, thereby causing disturbances ranging from increased fracture permeability to large earthquakes, such as with the postglacial faults of northern Fennoscandia (Lund et al. 2009, Lund & Schmidt 2011).

4.6 Hydrotectonic models

Hydrotectonic models describe the interaction between state of stress, fault kinematics and fluid flow along faults and fractures (Moeck et al. 2005). If any trends will emerge whether particular lineaments are more productive than others, this in turn will provide the basis for a groundwater hydrotectonic model for that region.

In his fundamental hydrotectonic model, Larsson (1959, 1963) stated that tensile fractures parallel to the dikes in the Precambrian basement areas of Sweden would be more transmissive and open than shear fractures, because the latter are held closed by a component of normal stress. Later Larsson (1967) pointed out that pre-existent anisotropies might cause local deviations from his idealized integrated ruptural model (see also Larsson 1968, 1971, 1972, 1977). Larsson´s data appeared to support his conclusions and the model has been widely used elsewhere (e.g. Reddy et al. 1993, Gustafsson 1994).

However, Larsson´s model has some weak arguments (Greenbaum 1990a, Banks et al. 1996). For example, the model assumes that shear or tension fractures generated in Precambrian times will retain their shear or tensional nature to the present. Tensional fractures may also be filled in and sealed with later mineralization that dramatically reduces the permeability. Later Larsson (1987) has added that the determining factor is the recent orientation of the stress field affecting the openness of the fractures.

A tectonic-hydrogeological model, originally developed in Scandinavia (Ericsson & Ronge 1986) and used in Botswana (Jonasson et al. 1986), assumes that rhomb-shaped lineament intersections are the result of two episodes of strike-slip faulting orientated at 90º to one another (Greenbaum 1990b). It further assumes that steep tension gashes parallel to the maximum principal stress will have formed and will possess the best potential for groundwater (Greenbaum 1990b).

Rohr-Torp (1987, 2003) examined the relationships between well yields and fracture zone directions in crystalline rocks of southern Norway. He classified fracture zones into shear fractures, tensional fractures and extensional fractures based on the maximum stress direction in the late Precambrian folding phase. Tensional fractures were developed parallel to the maximum stress direction N30ºE, while shear fractures were developed at 30º angles to this direction. Extensional (relief) fractures were developed N120ºE, i.e. parallel to the fold axes and at right angle to the maximum stress direction. The study showed that the highest yields were obtained in the tensional and extensional directions N30ºE and N120ºE, respectively. The shear fracture directions (N0ºE and N60ºE) gave the second highest yields, while other directions had pronounced inferior yields.

Dietvorst et al. (1991) state that the occurrence of groundwater in southeastern Botswana is not primarily controlled by large-scale fracture systems and lithologies, but by intersecting folds forming a “chess-board” pattern of dome and basin structures in the Precambrian basement. The most productive wells are situated in the tectonic basins, whereas the low-yield wells are located on the tectonic domes.

According to Greenbaum et al. (1993), the existence of sub-horizontal fracturing may be as important for borehole siting as steeply dipping lineaments. For this reason, they have proposed a conceptual model that takes account of both the sub-vertical and sub-horizontal fractures to explain the existence of successful well sites.

Banks et al. (1996) tested the hypothesis that fractures parallel to S_H would be expected to be more hydraulically open, and thus a borehole drilled perpendicular to S_H would cross these fractures and have the highest transmissivity and yield. However, the drilling results were inconclusive and borehole orientation with respect to stress field appeared to have at best only a minor impact on borehole yield.

Lyslo (2000) presented analytical and numerical models of stress fields associated with fracture lineament sets and groundwater transport in fractured rocks of Iceland and Norway. Those lineament sets, which were considered to have a potential for reactivation, would increase their permeability because of higher degree of fracture and pore interconnection along seismogenic rupture planes than outside of them.

Taylor and Howard (2000) have proposed a ‘tectono-geomorphic’ model of the hydrodynamic properties of crystalline rocks from Uganda. Their model demonstrates the interrelationships among the geomorphology, hydrology, and hydrogeology of weathered crystalline rock on a regional scale highlighting the practical importance of the high storage capacity of the thick weathered mantle above the fractured bedrock. Thus, these two units form an integrated aquifer system, albeit of highly variable and relatively low transmissivity.

Fernandes and Rudolph (2001) have investigated the role of the most recent brittle tectonics on opening or closing of both young and old (active or reactivated) fracture systems in the São Paulo region in Brazil. They used a method called 'homogeneous tectonic domain' (HTD), which was designed to assess the influence of orientation and presumed formation mode of the fractures to identify possible correlations with well productivity. The basic idea was that the most recent stress fields and tectonic movements probably control the current fracture aperture size and, consequently, the circulation of groundwater in the fractured media (see also Fernandes 2003). The HTD method considers aperture as the most important factor in controlling hydraulic conductivity of the rock. By means of a lineament analysis and knowledge of the tectonic evolution of the study area, the direction of the permeable extension fractures can be estimated and thus recognized within the lineament cluster.

Lachassagne et al. (2001, 2008, 2009, 2011) have developed a methodology for delineating favorable prospecting zones for the purpose of siting high-yield water wells in the Massif Central, France. The methodology adopts a functional approach to hard rock aquifers using a conceptual model of the aquifer and its structural compartments: weathered and decayed surficial rock, the underlying weathered-fissured zone and the fractured bedrock. The methodology considers lithology and hydrogeological properties, nature and thickness of the weathered rock and fissured zone, depth to the water table, slope, fracture networks and present-day tectonic stress.

Moeck (2005) and Moeck et al. (2005, 2007) investigated the relationship between stress orientation and groundwater pathways in the karstic Algarve Basin, South Portugal, by fault plane analysis. The hydraulically conductive faults were derived from electromagnetic measurements. The hydrotectonic modeling results indicated that the prefential groundwater conduits were highly critically stressed faults with respect to the current stress field.

Solomon and Ghebreab (2008) have examined how the fracture systems are related to the regional stress field and to well data in order to understand permeability and groundwater flow in hard rock aquifers of Eritrea. Frequency of groundwater occurrence (producing wells close to fracture lineaments), well yield and distance or locations of well data from the fractures were the main parameters considered in the analysis using geographic information systems (GIS). The relationship between well yield and distance to fracture lineaments indicated that the shear fractures are hydrotectonically more significant than the tensile fractures. The data summarized in a conceptual hydrotectonic model could be used as a working reference in selecting potential sites for future groundwater exploration in the region.

Surrette et al. (2008) have introduced a hydrostructural domain approach, which they tested and validated in fractured sedimentary rocks of the Gulf Islands, British Columbia, Canada. Their model-derived permeability and transmissivity values showed a good spatial agreement with the corresponding values obtained from pumping test. Moreover, the K-values increased with proximity to fault lines indicating structurally controlled groundwater flow and showed promise for use in regional characterization of fractured bedrock aquifers.

According to Greenbaum (1990a), the drilling successes and failures should be integrated with the tectonic data to develop a groundwater model that can be used to guide further exploration. However, Greenbaum recommends that no generalized tectonic model should be used as a basis for borehole siting unless it can first be demonstrated to have local validity. According to Greenbaum (1990b), it is possible that in many situations some empirical model that roughly accounts for local relationships is as much as can be expected.

According to Gustafsson (1994), the use of hydrotectonic models, based on lineament data, to identify certain fracture lineaments as more promising for groundwater exploration has to be considered as very difficult without extensive and precise exploration drilling and test pumping. According to Sander (1996), hydrotectonic models, as guides to promising lineaments, are not feasible, unless a large number of boreholes with high positional accuracy or current stress measurements are available to substantiate the existence of higher yielding zones. Likewise, there is sometimes an exaggerated faith in hydrotectonic models and impact of original or present stress regimes when analyzing lineaments (Sander 2007).

5 STUDY AREA

5.1 Location and general features

Central Finland region forms the NE part of the province of Western Finland. It is situated in between the geographic coordinates 61º27' - 63º37' N and 24º01' - 26º47' E (Fig. 19).

The total area of Central Finland is 19,950 km^2, of which the proportion of land is 16,706 km^2 and that of lakes 3,244 km^2. Forests cover some 80% of the land area. Lake Päijänne is the largest and lowest (+79 m a.s.l.) water body in the area. The land surface is at its highest 269 m a.s.l. on Kiiskilänmäki in the municipality of Multia. Central Finland has 273,000 inhabitants living in 17 municipalities and 6 towns. Jyväskylä, with its population of 130,000, is the capital of the region (www.keskisuomi.info).

Central Finland is located within the humid boreal climatic region with cold winters and warm summers. Due to the influence of the Gulf Stream and the Baltic Sea, the climate in the area is in many respects more favorable than in most other regions located between the 60th and 65th latitudes.

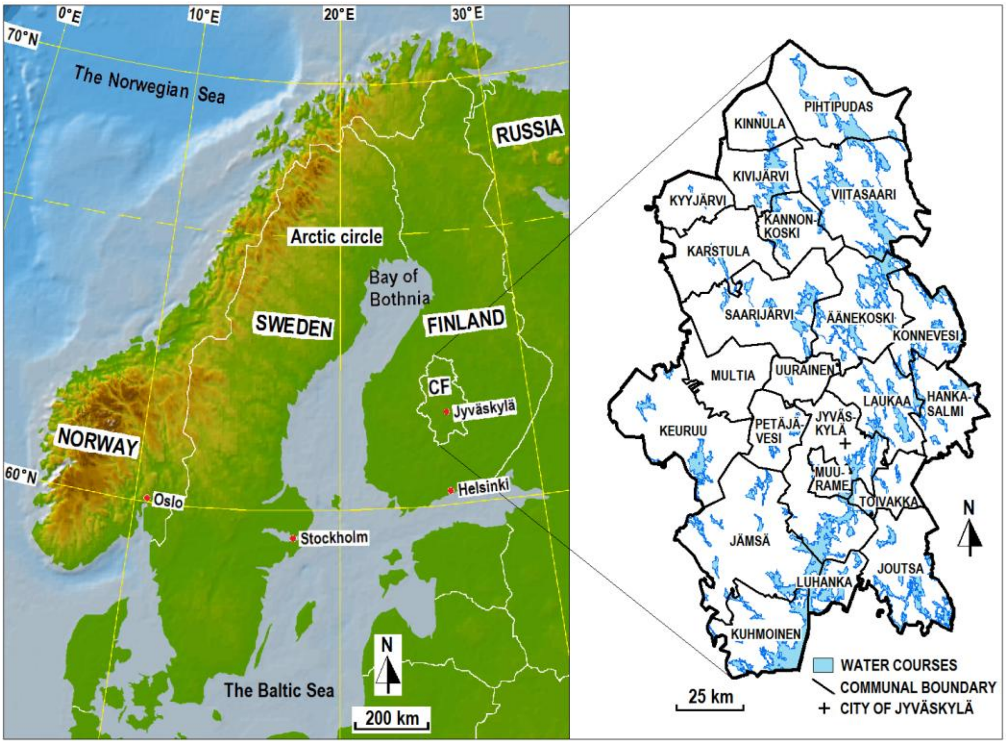

Fig. 19. Location and municipalities of Central Finland (CF). The map of Fennoscandia modified from http://www.freeworldmaps.net/printable/scandinavia/relief.jpg.

In Jyväskylä, the average annual air temperature is +3,0 ºC (Fig. 20). Monthly air temperatures are lowest in February with a mean of –8,7 ºC and peak in July with a mean of +16,0 ºC. The total annual precipitation is moderate varying between 600 and 650 mm, with some 30...40% in the form of snowfall (Fig. 20). The monthly precipitation is relatively uniform, with highest averages recorded in August (88 mm) and lowest in February (31 mm). Annual precipitation is twice that of evapotranspiration. The duration of snow cover is from 4 to 6 months. Ground frost generally develops in late November and lasts until April when the spring snowmelt begins (http://en.ilmatieteenlaitos.fi/normal-period-1971-2000).

Values for groundwater recharge and groundwater runoff from hard-rock environments in Fennoscandia are few, and restricted to local catchments. Values are in the range of 0,15…3,0 Ls^{-1} per km^2, that is 1…15% of annual rainfall 650 mm (Niini 1977, Gustafson 1988, Olofsson 1994, Olofsson et al. 2001, Henriksen 2003a, Bockgård 2004, Rodhe & Bockgård 2006).

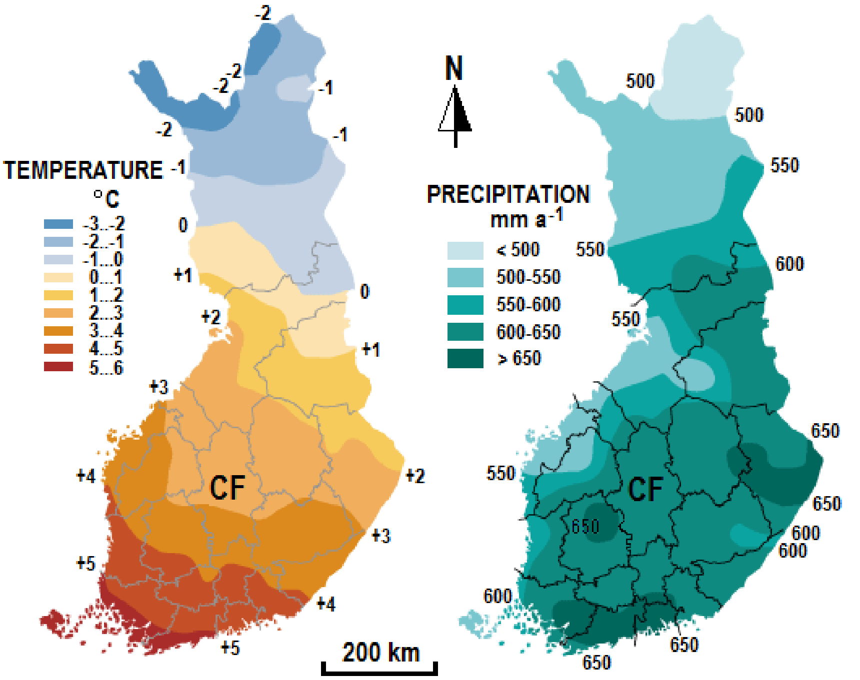

Fig. 20. The mean annual air temperature (°C) and the average annual precipitation (mm a^{-1}) in Finland during the years 1971-2000. CF = Central Finland. Modified from the Finnish Meteorological Institute at http://en.ilmatieteenlaitos.fi/normal-period-1971-2000.

In Central Finland, about 80% of inhabitants have joined a piped community water system maintained by a communal or private cooperative organization of water supply. In Central Finland, there exist around 150 water intakes with total water delivery of 52,000 m^3d^{-1}. The average specific water consumption is 240 $Lday^{-1}$ per person. Some 85% of all consumed water is groundwater or artificial groundwater extracted from glaciofluvial aquifers. The city of Jyväskylä uses also purified surface water in winter time. Bedrock groundwater is used as a common water supply in some 30 villages and small communities. Around 60% of rural inhabitants use their own wells, a major part of which are drilled bedrock wells (www.ymparisto.fi).

5.2 Precambrian bedrock

5.2.1 Svecofennian domain

The bedrock of Northern Europe is dominated by a large area of Precambrian crystalline rocks – the Fennoscandian Shield – in Finland, most of Sweden, south and southwestern Norway as well as Karelia and the Kola Peninsula in Russia (Fig. 21).

The Fennoscandian Shield, which is surrounded by the Caledonides in the west and flanked by young Phanerozoic sediments in the south, can be divided into three domains: the Archean, the Svecofennian and the Scandinavian (Gaál & Gorbatschev 1987). Central Finland region lies in the Svecofennian domain (Fig. 21). The Svecofennian domain can be divided into three tectonic units: primitive arc complex of central Finland, the accretionary complex of western Finland and the southern Finland accretionary complex; the last one does not reach Central Finland (Korsman et al. 1997, Nironen 1997, 2003, 2005).

The Svecofennian domain was formed by accretional and collisional type orogenies, by subduction and deformation and by high-grade metamorphism and extensive crustal melting during the period of 2,0-1,55 Ga. It is composed solely of Palaeoproterozoic crust (Gaál 1982, 1986, 1990, Park 1985, Huhma 1986, Gáal & Gorbatschev 1987, Hölttä 1988, Korja et al. 1993, Kärki et al. 1993, Lahtinen 1994, Korja 1995, Vaasjoki 1996, Nironen 1997, Nironen et al. 2000, Rämö et al. 2001, Korja & Heikkinen 2005, Vaasjoki et al. 2005).

Various plate tectonic models have been applied to the Svecofennian domain but they have narrowed to arrive at a model whose essential features were stated by Hietanen (1975). The tectonic models differ mainly in the number and direction of subduction events (Gaál & Gorbatschev 1987, Gaál 1990, Park 1991, Ekdahl 1993, Lahtinen 1994, Korja 1995, Ruotoistenmäki 1996, Lahtinen & Huhma 1997, Nironen 1997, Lahtinen at al. 2006). The Svecofennian tectonic system has been presumed to resemble the modern Indonesian archipelago between the Australian and Eurasian continents (Nironen 1997).

In addition to the plate driving forces, also gravitational (extensional) forces were important during Palaeoproterozoic crustal and lithospheric growth (Korja 1995, Korja & Heikkinen 1995, Korja et al. 2004). The gravitational collapse is displayed in the deep seismic reflection profiles of FIRE as shallow, upper crustal post-collisional extensional structures (3-8 km) cross-cutting the 20-25 km thick collisional ones (Kosunen et al. 2006). Nironen et al. (2002c), Korja et al. (2004) and Lahtinen et al. (2005) suggest that the Svecofennian domain comprises of at least five orogens each having compressional and extensional stages.

By 2,0 Ga the arcs of different evolutionary stages started to form microcontinents and at 1,89 Ga they accreted to the Karelian continent. Oblique collision caused some of the shear zones at or close to the terrane boundaries. The microcontinental collision stage resulted in thickened lithosphere in the core area (Nironen 1997). At the end of the microcontinent accretion stage a larger but unstable continental Fennoscandian plate had developed. 1,80 Ga localized several continent-continent collisions around the Fennoscandian nucleus. This period was characterized by rapid uplift, voluminous granitoid magmatism and pegmatite intrusions around the nucleus (Nironen 1997).

After the collision stage Fennoscandia underwent a major stabilization period at 1,79-1,6 Ga with late tectono-magmatic episodes and formed a stable cratonic platform (Lahtinen et al. 2006). The Svecofennian orogeny was followed by a long period of erosion. Erosion removed the top 15 km of the bedrock before the beginning of the Cambrian about 650 Ma ago (Vaasjoki et al. 2005). During the last 600 Ma years, the ground surface has lowered by erosion some tens or hundreds of meters to its present level (Vuorela & Niini 1982, Saari 1992).

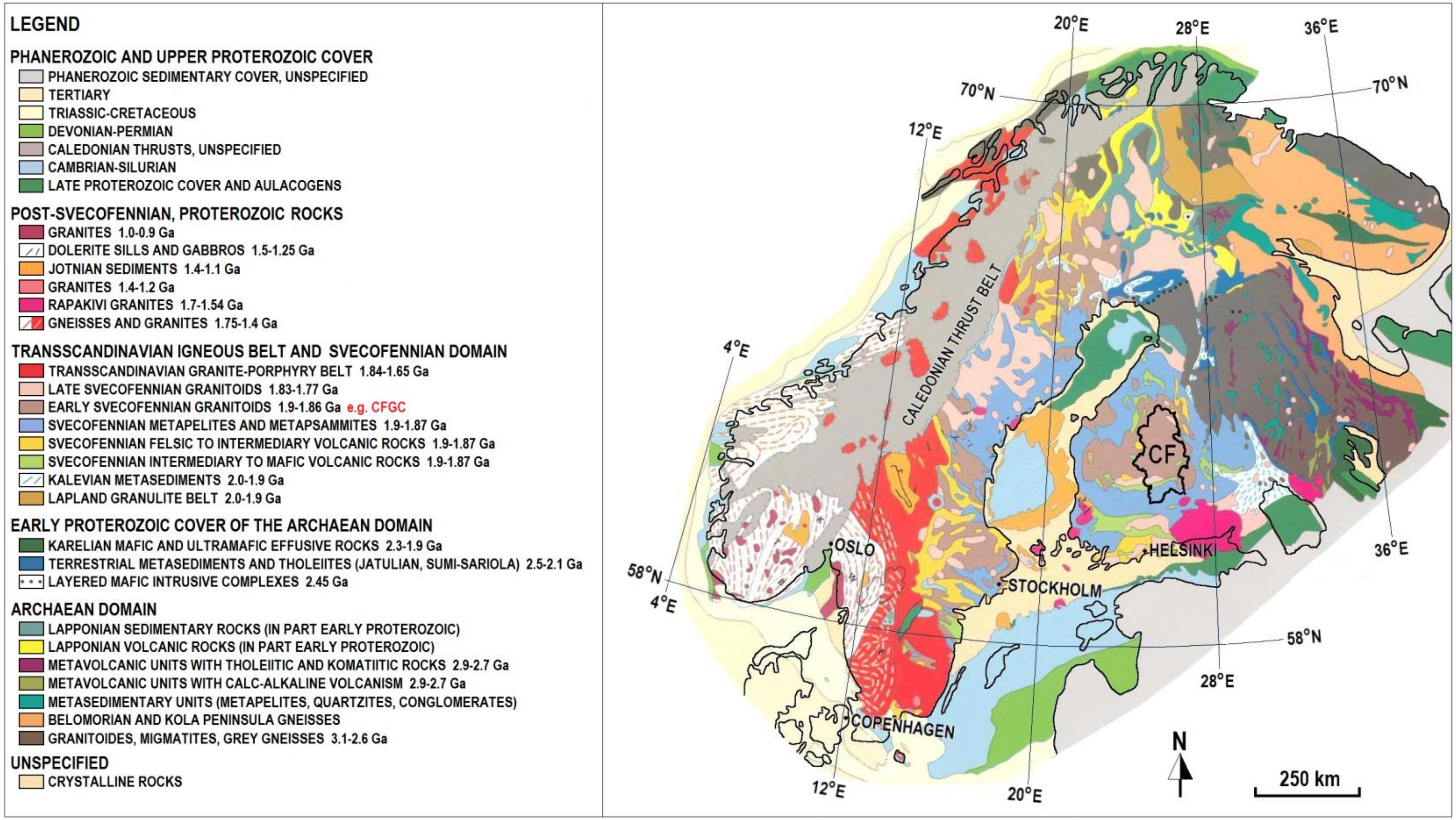

Fig. 21. Generalized lithostratigraphic map of the Fennoscandian Shield and adjacent sea areas. CF=Central Finland. Modified from Larsson and Tullborg (1993).

5.2.2 Central Finland Granitoid Complex

The Central Finland Granitoid Complex (CFGC), covering an area of about 44,000 km^2, constitutes the core of the Svecofennian arc complex of Finland (Fig. 21; Korsman et al. 1997, Nironen et al. 2002a, Nironen 2003). The complex is bordered in the south by the Tampere and Pirkanmaa Belts with well-preserved volcanic and sedimentary rocks, which continue in the east to the interior of the CFGC (Kähkönen 1989, 2005, Nironen 1989). In the west the CFGC is bordered by the Bothnian Belt (Nironen 1997, Kähkönen 2005). In the east and north the complex is bordered by the Savo Belt with the Ladoga–Bothnian Bay zone (LBB) as a boundary between the Palaeoproterozoic Svecofennian and the Archean Karelian areas (Hietanen 1975, Korsman et al. 1984, Gaál & Gorbatschev 1987, Vaasjoki & Sakko 1988, Kärki et al. 1993, Kähkönen 2005, Moisio 2005). The NW-SE orientated LBB was formed as a consequence of an oblique collision between the Archean continent and the Svecofennian island arc system (Korja & Heikkinen 2005). The Svecofennian crust is exceptionally thick in Central Finland, up to 60 km (Korja et al. 1993, Vaasjoki et al. 2005).

The CFGC comprises mainly calc-alkaline I-type granitoids (1,89-1,88 Ga) with minor amounts of mafic plutonic rocks as well as remnants of deformed sedimentary and volcanic rocks (Front & Nurmi 1987, Korsman et al. 1997, Nironen et al. 2002a). Felsic plutonic rocks are predominant rock types in the CFGC area (Fig. 22). Even to medium-grained granodiorites are the most typical plutonic rocks. There is a whole gradation from even-grained granodiorites to porphyritic granodiorites (Nironen 2003). In places the granodiorites grade into tonalites. Also even-grained and porphyritic granites are common. Pyroxene-bearing granite is found as a marginal variety of some postkinematic type 3 granites (Nironen et al. 2002a, Nironen 2003, 2005).

The intermediate plutonic rocks occur as complexes, in which the composition grades from quartz monzodiorite to monzodiorite and quartz diorite. The coarse-grained or coarse-porphyritic quartz monzonites occur as marginal phases or constitute the whole pluton (Nironen et al. 2000). Small gabbro-diorite bodies are found throughout the CFGC area (Fig. 22; Nironen et al. 2002a, Nironen 2003, Peltonen 2005). A few ultramafic bodies are associated with the gabbros, but due to their smallness they are not seen in the bedrock map of Central Finland (Nironen 2003). Subvolcanic rocks, which are typically porphyritic, grade in places into plutonic rocks.

The plutonic rocks can be divided into synkinematic and postkinematic with respect to the prominent deformation in the CFGC area. The postkinematic rocks display many of the characteristics of the rapakivi granites in southern Finland (Elliott et al. 1988, Elliott 2001, 2003, Rämö et al. 2001, Nironen 2003, 2005).

The most typical supracrustal rocks in the CFGC area are fine-grained volcanic tuffites and tuffs of intermediate composition (Nironen 2003). Mafic volcanic rocks are mainly located in the southern part of the complex. Also felsic volcanic rocks are found in a few places. More common felsic rocks are quartz-feldspar schists and gneisses that are found among the other supracrustal rocks throughout the CFGC area (Nironen et al. 2002a, Nironen 2003). Mica schists are found within the Tampere belt and its extension, the Luhanka-Leivonmäki area, in southern CFGC. The mica gneisses are migmatitic biotite or biotite-hornblende gneisses. Peak metamorphism took place 1,885-1,880 Ga ago when the synkinematic tonalites and granodiorites were emplaced (Nironen 2003).

The oldest Svecofennian rocks are 1,92-1,91 Ga gneissic tonalites in the northeastern part of the CFGC. They are the remnants of the primitive arc complex of central Finland (Korsman et al. 1984, Vaasjoki & Sakko 1988) and indicate plutonism that predated the 1,89-1,88 Ga synkinematic period of the main part of the CFGC area (Aho 1979, Huhma 1986, Kallio 1986, Pääjärvi 1991a, Lahtinen & Huhma 1997, Rämö et al. 2001). The tonalites, granodiorites and gabbrodiorites appear to be the oldest synkinematic rocks in the CFGC area. The subvolcanic rocks are part of the synkinematic magmatic event (Nironen 2003). The synkinematic plutonic rocks crosscut the adjacent supracrustal rocks. In Central Finland the postkinematic rocks are almost coeval with the synkinematic rocks.

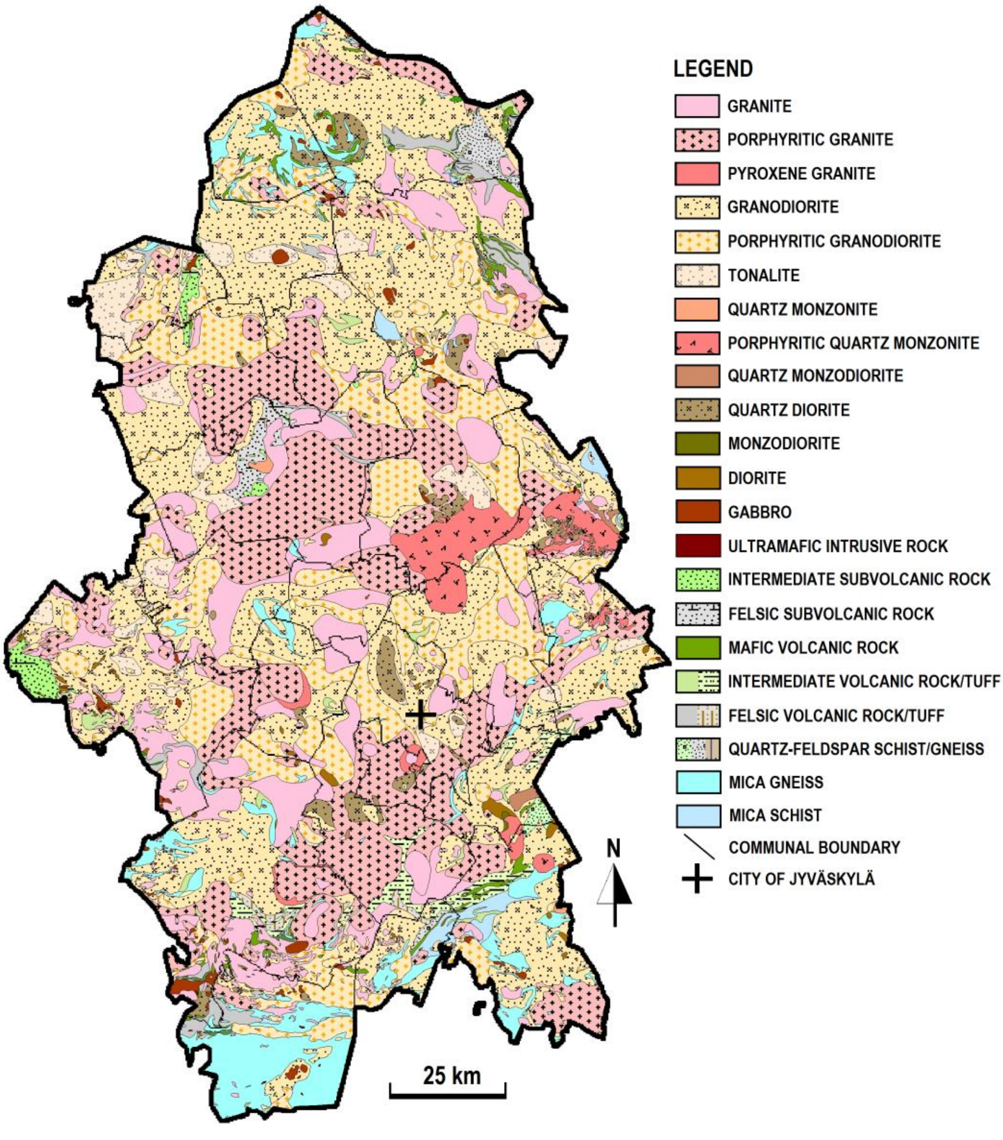

Fig. 22. Bedrock map of Central Finland. Modified from GTK (2010).

There are at least eleven asteroid impact craters in Finland, of which Karikkoselkä and Keurusselkä are situated in Central Finland (Pesonen et al. 1999, 2005, Kinnunen & Hietala 2009).

For a more thorough review of the Svecofennian bedrock of Central Finland, the reader is referred to maps and reports by Sederholm (1894, 1911), Frosterus (1900, 1902), Wilkman (1929, 1931, 1935, 1936, 1938), Saksela (1934), Laitakari A. (1942), Laitakari I. (1971, 1973), Simonen (1960, 1980), Nykänen (1962, 1963), Marmo (1963a, 1963b), Salli (1963, 1967, 1969, 1971, 1983), Pipping (1966, 1972, 1976), Marmo and Laiti (1970), Laiti (1976), Kallio (1982, 1986, 1987, 1988), Sjöblom (1984, 1990), Pääjärvi (1985, 1991a, 1991b, 2000a, 2000b), Anon (1989b), Marttila (1992, 1993), Pipping and Vaarma (1993), Vaarma and Pipping (2003), Kilpeläinen (2007) and Kilpeläinen et al. (2008).

5.2.3 Tectonic development and lineaments

According to Gaál (1982), by about 1,85 Ga ago the crust had sufficiently consolidated to accommodate a regional uniform stress field in which the main principal stress acted subhorizontally in an ESE direction and the least principal stress subhorizontally in a NNE direction. The crust reacted to stress in a semi-plastic manner. Large-scale fracture zones or megashears in the CFGC area are associated with intense deformation around 1,8 Ga ago (Talvitie 1979, Berthelsen & Marker 1986).

According to Talvitie (1971, 1975), the most prominent and continuous wrench faults in the Kuopio region, to the northeast of the study area, have a strike of 320º-330º (NW-NNW) and a sub-vertical to vertical dip. However, the glacial abrasion has possibly emphasized the prominence of the direction. On the other hand, the 25º-30º (NNE) striking faults have a dip between 40º and 90º, which, according to Talvitie (1975), indicates that they are not wrenches but shears of a different kind. Fracture zones have been remobilized repeatedly during geological time (Talvitie 1979). Kärki (1995) has investigated the structural geology of Kainuu region to the north of the CFGC. According to him, the NW-SE orientated dextral shear zones form a conjugate system with the NE-SW striking sinistral shear zones.

According to Nironen et al. (2002a) and Nironen (2003), the trend directions of the major lineaments in the CFGC area may be grouped grossly into 1) 20º-40º (NNE-NE), 2) 120º-135º (WNW-NW) and 3) 0º (N). The group 2 (wrench) faults have deformed the group 1 faults and contained originally a dextral horizontal shear displacement component as the result of convergence towards the present north (Gaál 1986, Ekdahl 1993, Nironen 2003). Later they appear to have controlled the emplacement of many postkinematic intrusions in extensional or transtensional stage during 1,88-1,87 Ga (Lahtinen & Huhma 1997, Nironen et al. 2000, 2002a, 2006, Nironen 2003). Nironen (2003) has interpreted the group 3 faults as the youngest major faults in the CFGC area. The faults have been active in many different stages, especially during the Rapakivi magmatism (1,650-1,510 Ma; Nironen 1997) but also during the Sveconorwegian (1,100-900 Ma) and the Caledonian (450-350 Ma) orogenies (Kohonen & Rämö 2002, 2005).

In the southern and southeastern part of the CFGC area there exist WNW-ESE trending diabase dikes, which belong to the Häme diabase dike swarm (1665 Ma; Kallio 1987, Laitakari 1987, Vaasjoki et al. 1991, Vaasjoki 1996, Rämö & Haapala 2005, Nironen et al. 2006). These diabase dikes, which represent the youngest rocks in Central Finland, intruded along faults reactivated in an extensional phase about 200 Ma after the postkinematic magmatism of the CFGC area and some 20 Ma prior the start of the Rapakivi magmatism in southern Finland (Vaasjoki et al. 1991, Nironen 2003).

The FIRE profiles reveal a three-layered structure of the crust in the CFGC area (Kosunen et al. 2006). Shallow SE dipping listric structures, which flatten out near the middle crust, characterize the upper crust down to 10 km. These structures probably represent thrust faults developed during the collisional stages of the Svecofennian orogen (Heikkinen et al. 2004, Kukkonen et al. 2006, Sorjonen-Ward 2006). The surface intersects of some of these structures broadly correlate with the NE-trending changes in magnetic anomaly levels. These structures are transected by several relatively steep dipping lineaments (Lahtinen et al. 2005, Kosunen et al. 2006). The middle crust from 10 to 35 km is segmented by strong reflections dipping gently SE. The lower crust down to almost 60 km is characterized by diffuse, nearly horizontal reflections (Kosunen et al. 2006).

The author has made lineament interpretation for Central Finland in the late 1990´s (Mäkelä 1997). The interpretation was updated in 2007. Lineaments were identified from topographic maps (scale 1:20 000 with contour lines at 2,5…5 meter intervals and depth contours shown for 3, 6, 10, 15, 20, etc. meters). Because especially long lineaments were not of straight-line form, lineaments were also segmented, i.e. curved lines were split into straight lines (e.g. Tam et al. 2004). No attempts were made to interpolate or extrapolate lineaments. Lineaments of less than 0,4 km in length were discarded. The digitized lineament map (1:200 000) classifies as a true fracture lineament map where non-geologic and doubtful features are absent (Mäkelä 1997).

Lineament azimuths were plotted on an azimuth-frequency plot (a rose diagram or "rosegram"; Brown 1994). Lineament frequency distributions are often normalized with lineament length to allow the short lineaments to count less than long lineaments, which perhaps provides a better way to describe the data set from a hydrogeological point of view (Wladis 1995, Lie & Gudmundsson 2002). Because the digitized lineaments in this study were segmented when they deviated from a straight line, there was no need for normalization.

The multiphase deformation tectonic history of the study area has resulted in a complex distribution of numerous lineaments, which produce a mosaic structure in the crust of Central Finland (Fig. 23). There are about 5,400 lineament segments in the lineament map. Their total length is approximately 5,840 km. The length of non-segmented single lineaments varies from 0,4 to 100 kilometers. The longest lineaments trend in the NW-SE direction. Some of them traverse Central Finland and continue outside the study area. Most lineament segments (41%) trend in a NNW-SSE direction. Another frequency maxima, though much lower, can be detected in a NNE-SSW direction in the azimuth rose diagram (Fig. 24).

The lineament interpretations match for most parts with those by Salmi et al. (1985) and Veriö et al. (1993). They have identified lineaments based on satellite and aerial photos and geophysical and topographic maps for the block mosaic model of bedrock used in determining the most suitable sites for the final disposal of spent nuclear fuel in Finland. Because of the different purpose of the work, the author´s interpretation excludes many lineaments, which are included in Salmi´s and Veriö´s maps. Indeed, as Limaye (2004) has stated, hard rock hydrogeologists are divided into two main groups: those interested in obtaining groundwater for water supply by exploring fractured and permeable zones and those interested in locating impermeable or very low permeability zones for storage of hazardous waste.

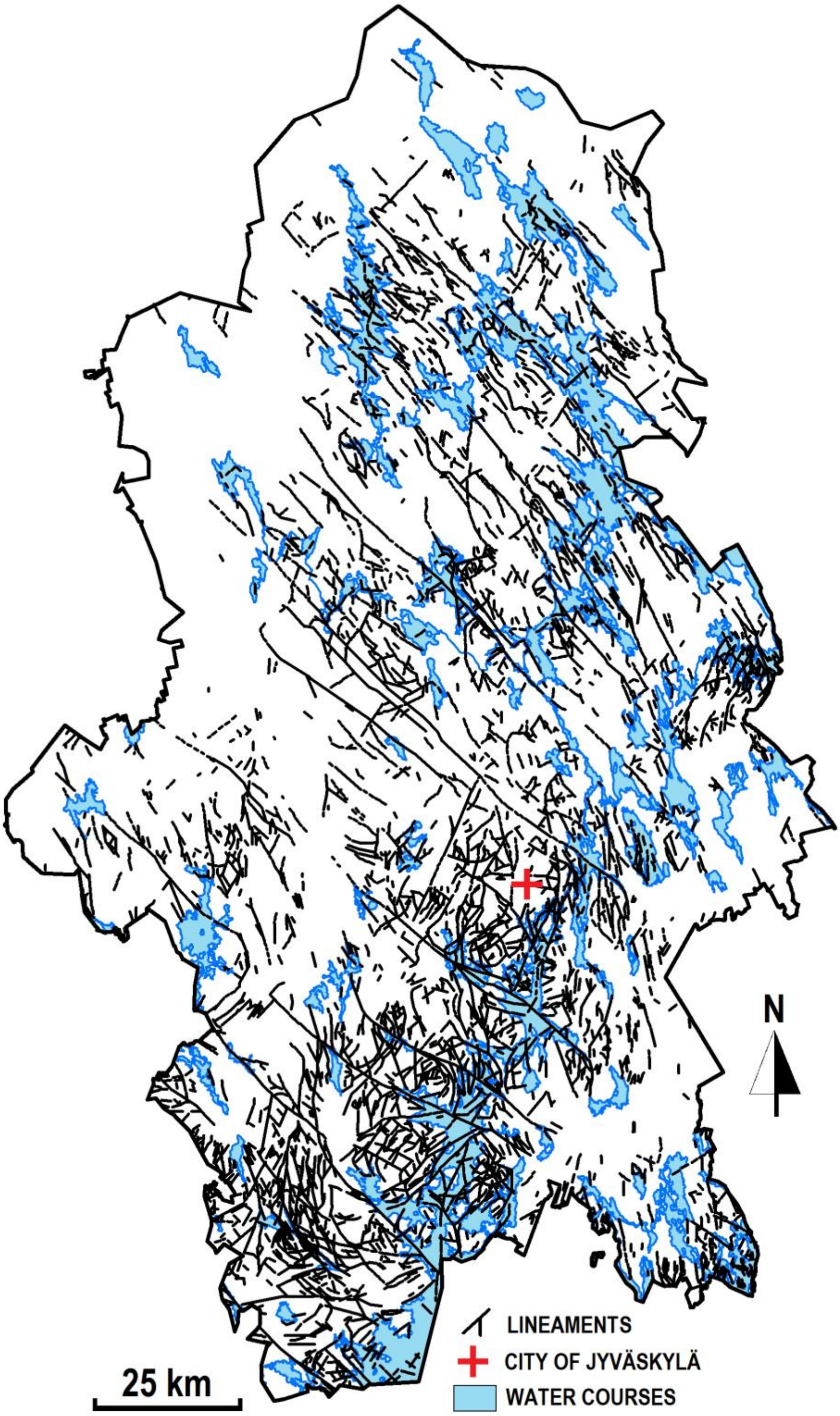

Fig. 23. Topographic fracture lineament interpretation for Central Finland. Modified from Mäkelä (1997).

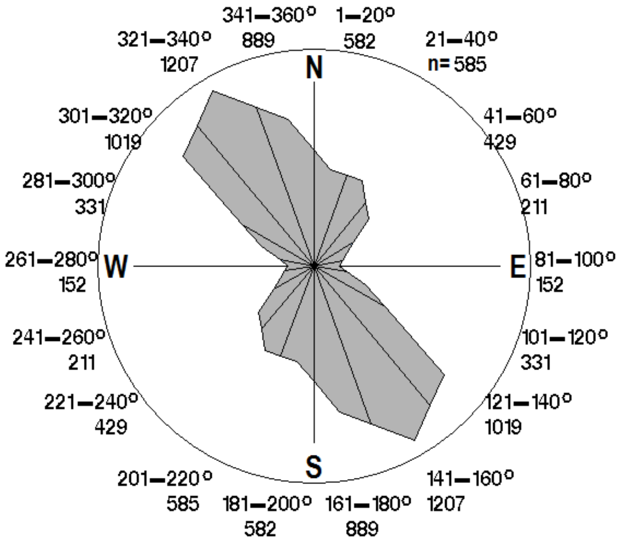

Fig. 24. Azimuth-frequency plot for the lineament segments in Central Finland (n=5405). The sector width is 20 degrees. The radius of the circle is n=1300.

5.3 Quaternary deposits and landforms

The surficial geology of Fennoscandia is characterized by young, unconsolidated Quaternary deposits of glacial origin on glacial-eroded, unweathered, Precambrian bedrock (Möller 1978, Englund et al. 1988, Donner 1995, Knutsson 1997, Seppälä 2005). There are widespread areas of bare rocks in the mountains and along the coasts, especially in Norway. Preglacial weathering surfaces formed before the glaciation have survived in a few places only, most notably in Lapland, northern Fennoscandia (Vaasjoki et al. 2005).

In Central Finland, as in Fennoscandia as a whole, the dominating Quaternary deposit is the hard packed, sandy or silty till, which is a few meters thick on an average, but can have a thickness of 50 m or more at some localities (Fig. 25). The tills often constitute morphologically prominent features such as hummocky or Rogen moraines and drumlins (Glückert 1973, Mäkelä 1988, 1995, 1996, Mäkelä & Illmer 1992).

The most striking Fennoscandian features from the Quaternary period are the elongated ridges of glaciofluvial gravel, sand and silt deposits – the eskers, deltas, outwash plains, sandurs, valley deposits, and ice-marginal landforms - e.g. the Salpausselkäs in Finland, the Middle Swedish ice marginal ridges and the Raernes in southern Norway (Rainio 1991, 1996, Seppälä 2005, Knutsson 2008). These ice-marginal deposits can be followed around the entire Fennoscandia.

The SW-NE orientated Central Finland ice-marginal landform (CFI) lies in the middle of the study area (Fig. 25). Glaciofluvial deposits consisted of sand and gravel form largest aquifers and are of great importance to the public water supply of Central Finland (e.g. Sara 2001). There are often fossil dunes and dune fields on the esker flanks (Mäkelä & Illmer 1990, Nevalainen 2011). Peat is found over large areas but clay is almost missing in Central Finland, which is away from coastal areas. The soil profiles formed after glacial time are mostly podsolic.

For a detailed account of the Quaternary geology of the study area the reader is referred to maps and reports by Frosterus (1903, 1913), Sederholm (1906), Sauramo (1923, 1924, 1926, 1929a, 1929b), Leiviskä (1928, 1951), Saksela (1930), Brander (1934a, 1934b), Brenner (1945), Mölder (1953), Mölder and Salmi (1954), Virkkala (1954, 1963), Mölder et al. (1957), Hyyppä (1963), Repo (1964), Alhonen (1971), Repo and Tynni (1971), Aartolahti (1972), Rainio (1972, 1991, 1996), Glückert (1974, 1977), Kukkonen (1976), Niemelä (1979), Punkari (1979), Lindroos (1988), Mäkelä (1988, 1995), Haavisto-Hyvärinen and Rainio (1992) and Kielosto et al. (1992).

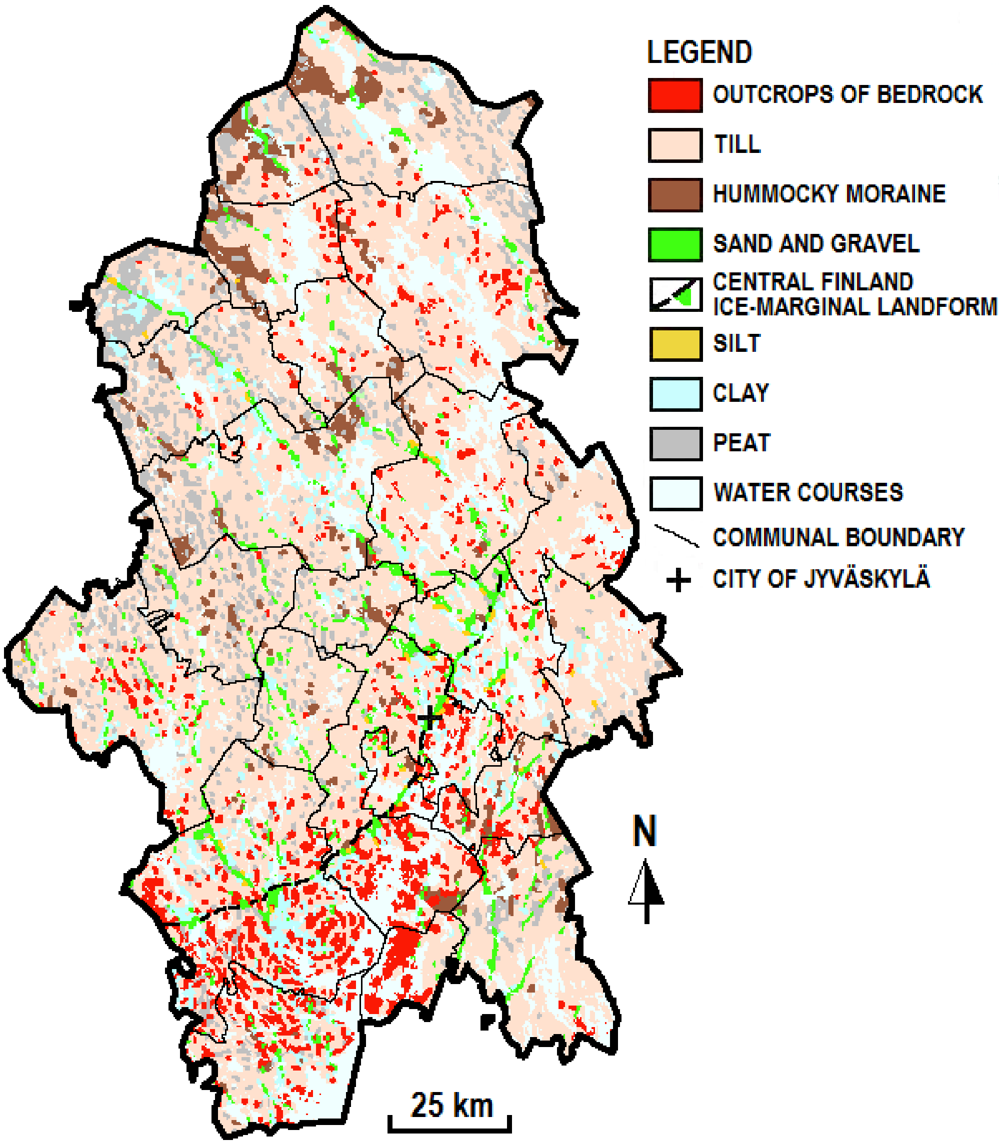

Fig. 25. Map of Quaternary deposits and landforms in Central Finland. Modified from GTK (2009).

5.4 Deglaciation and shore-level displacement

The front of the Weichselian ice sheet retreated from the Second Salpausselkä to Central Finland in the Yoldia stage shortly after 10,500 years BP (Fig. 26). During the glacial retreat there were two fan-shaped ice lobes that extended to Central Finland (Punkari 1979, 1997). The western ice lobe deposited the Central Finland ice-marginal landform (CFI) when the melting ice front readvanced for at least 50 km from the present Keuruu region and halted at the Jämsä-Jyväskylä-Laukaa region (Fig. 25; Virkkala 1963, Aartolahti 1972, Rainio et al. 1986). According to Saarinen (1994), the CFI took shape at about 10,250-10,150 years BP, i.e. about 11,000 calendar years ago (Ojala et al. 2005). After the deposition of the CFI the ice front began to retreat and the glacial ice finally melted in the northern Central Finland shortly after 9,400 years BP (Glückert 1974, Ignatius et al. 1980).

Glacial striae and drumlins indicate that the predominant ice-flow direction in Central Finland has been south-southeastward (Glückert 1973, Mäkelä 1988, 1995). The glacial ice cleaned out and eroded in particular those fracture lines, valleys and lake basins parallel with the glacial flow.

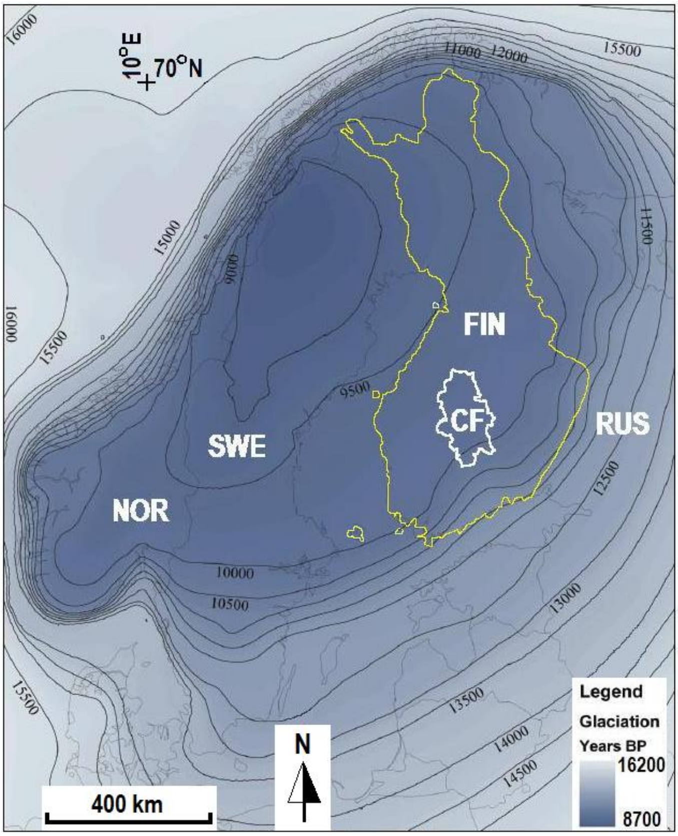

Fig. 26. The successive ice-marginal lines for the Weichselian deglaciation in Fennoscandia between ca. 16,000 and 8,500 years BP. NOR=Norway, SWE=Sweden, FIN=Finland, RUS=Russia, CF=Central Finland. Modified from Munier and Fenton (2004).

Ristaniemi (1985, 1987) has studied the shore-level displacement in Central Finland. His observations prove the existence of three easily distinguishable, fully developed ancient shore levels: the highest shore level of the Baltic Sea (HSL), the Ancylus limit and the level of Ancient Lake Päijänne. The HSL was metachronously formed during the Yoldia stage except in northern Central Finland, where it took shape in the beginning of the Ancylus stage (Ristaniemi 1987, Mäkelä 1988).

As consequences of land uplift the shore levels have tilted southeastwards, the HSL as being the oldest having the steepest gradient. For the same reason the HSL is much higher in the NW part of the area in comparison to the SE part (Fig. 27:A). The HSL is at its lowest 130 m a.s.l. in Joutsa, southeastern Central Finland, and at its highest about 200 m a.s.l. in Pihtipudas, northwestern Central Finland. The altitude of the Ancylus limit is 110 m a.s.l. in Kuhmoinen, 130 m a.s.l. in Jyväskylä and approximately 165 m a.s.l. in Viitasaari. The Ancient Lake Päijänne level at 6,100 years BP is ca. 95 m a.s.l. in Kuhmoinen, ca. 105 m a.s.l. in Laukaa and ca. 120 m a.s.l. in Pihtipudas. No hinge lines are to be found in the shore levels (Ristaniemi 1987). A considerable part of Central Finland lies above the highest shore level (Fig. 27:B). The Litorina Sea did not reach Central Finland (Glückert 1989, 1993, Tikkanen & Oksanen 2002).

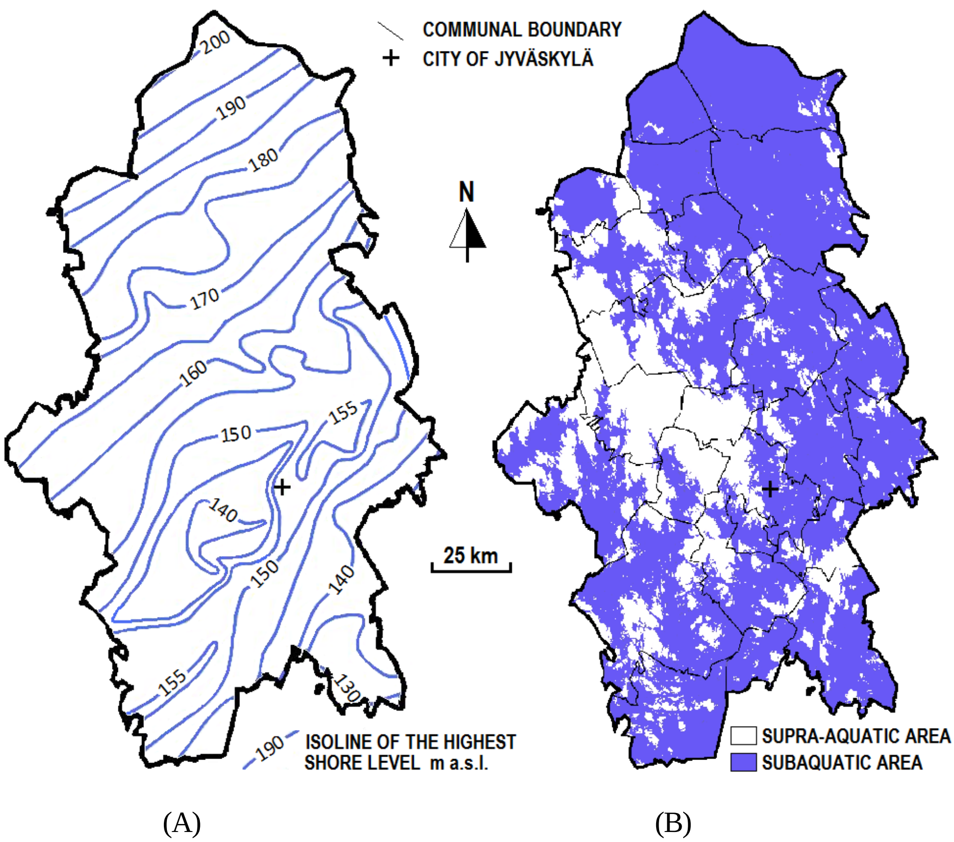

Fig. 27. A) The isolines of the highest shore level (m a.s.l.) of the Baltic Sea in Central Finland interpolated manually by the author from the shore level data of Ristaniemi (1987). B) The distribution of supra-aquatic and subaquatic areas in Central Finland reconstructed by subtracting the HSL isoline surface from the database of present topography (e.g. Leverington et al. 2002). Digital elevation database © National Land Survey of Finland; map reconstruction with GIS by Juha Romula/CETECF.

According to Ristaniemi (1987), the isolines of the highest shore level of the Baltic Sea (Fig. 27:A) reflect the position of the latest Weichselian ice margin and the progress of deglaciation in Central Finland. During the ice lobe stage there existed a calving bay between Jämsä and Konnevesi. This can be seen as higher HSL values (+155) of this area. Interestingly, also the Keuruu readvance can be detected from the HSL as lower shore levels to the west of Jyväskylä (+140…+150; Fig. 27:A).

5.5 Seismotectonics

To the author´s knowledge, only a few rock stress measurements have been carried out in Central Finland. In the city of Jyväskylä the measurements have been done near the ground surface (depth 5-19 m). The results of the two measurements strongly deviate from each other: 268° (N92°W) and 351° (N9°W) (Matikainen et al. 1981). On the other hand, at the Kivetty site of Äänekoski, the stress measurements have been performed in a deep borehole for the need of final disposal studies of high-level nuclear waste in Finland (Klasson & Leijon 1990). These measurements have been done with the hydrofracturing method on both the 500 m level and the 800 m level below the ground surface. The maximum horizontal stress magnitude ranged in 11 tests between 20-42 MPa. The orientation data, which were comprised in 13 tests, ranged between 295-328° with an average S_H direction of 309° (N51°W) (Figs. 12 and 28; Klasson & Leijon 1990).

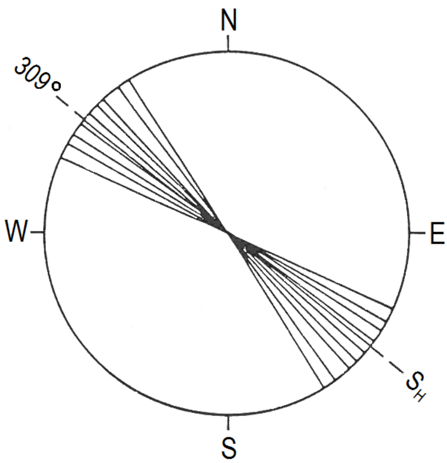

Fig. 28. Measured orientations of the maximum horizontal stress S_H in the Kivetty borehole KI·KR1 in the municipality of Äänekoski, Central Finland. Modified from Klasson and Leijon (1990).

In Central Finland there occur extremely few earthquakes. According to the FENCAT database of the Institute of Seismology, University of Helsinki, only 10 earthquakes have occurred in the study area in five different sites since the year 2000. Their magnitudes have ranged between M_L 0,3-2,4 and the focal depths between 0,2-6 km. The Toivakka M_L 2,4 earthquake on the 11th of May 2000 indicated thrust or reverse faulting at a depth of 5 km with a nodal plain trending 358°/42° (Uski et al. 2003a). The Jämsä M_L 1,5 earthquake on the 20th of July 2002 was steep dipping normal or gently dipping reverse type earthquake (Uski et al. 2003b). The Pyhäjärvi M_L 2,1 earthquake just outside Central Finland on December 12th 2007, initiated by prior mining in the vicinity of it, was steep dipping normal type striking 45º (Uski et al. 2008). The largest historical earthquake in Central Finland happened on the 16th of November 1931 in the municipality of Laukaa (M_M 4,5; Mäntyniemi 2004).

In Central Finland, the present uplift rate relative to mean sea level from three precise levellings varies from 3,6 to 6,5 mm a^{-1} (Fig. 29; Mäkinen & Saaranen 1998, Mäkinen 2000, Saaranen & Mäkinen 2001, 2002, Mäkinen et al. 2003, Saaranen 2005). The reliability of the latest uplift models is around 0,2 mm a^{-1} in Central Finland (Vestøl 2006).

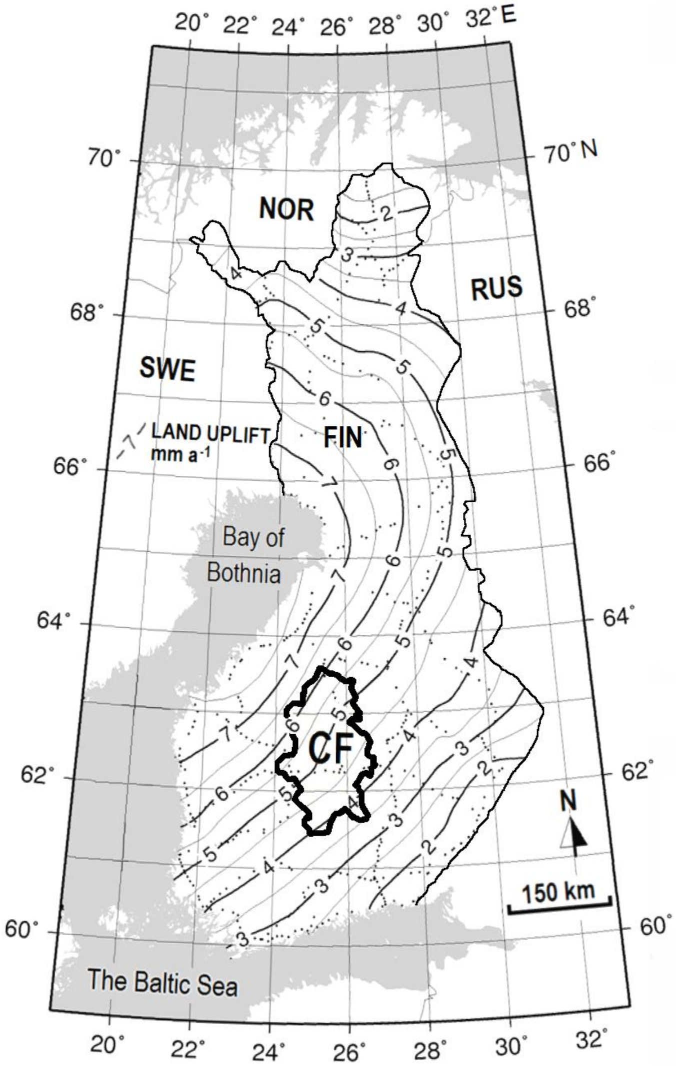

Fig. 29. Land uplift (mm a^{-1}) relative to mean sea level from three precise levellings in Finland. Black dots indicate calculation network, starting value at Hanko 2,63 mm a^{-1}. SWE=Sweden, NOR=Norway, FIN=Finland, RUS=Russia, CF=Central Finland. Modified from Saaranen (2005).

6 MATERIAL AND METHODS

6.1 Central Finland drilled well database

6.1.1 Background

Most countries operate some form of well and borehole archive, containing all the details of wells reported by drillers, owners and operators (Misstear et al. 2006). The archive may be in the form of paper records or it may be computerized. It may be lodged with a geological survey, or with a water or environment agency or ministry. There may be legislation requiring the driller or owner of any new well or borehole to submit details to the archive (Misstear et al. 2006).

There is often a limited amount of borehole data especially in countries, which lack a national policy for borehole registration. If inadequate archiving has been made, a major and unnecessary loss of investment will have occurred, since the cost of obtaining the equivalent data by drilling will be very high (Foster 1984). The lack of hydrogeological database is also a major obstacle to groundwater resources evaluation (e.g. Ajayi & Abegunrin 1994, Robins et al. 2002).

In Finland, there is no statutory well database at the moment, albeit in 1982 an initiative was made to establish such a database by the Finnish Association of Engineering Geologists (Lindblad & Rönkä 1982, Rönkä & Lindblad 1982). The Geological Survey of Finland (GTK) has a nationwide database of a few thousand drilled wells with comprehensive water quality information but, unfortunately, sparse well construction and yield data (e.g. Lampén et al. 1992). The Finnish Radiation and Nuclear Safety Authority (STUK) has a nationwide database of 9,200 drilled wells with natural radionuclide information, for example radon determinations (Salonen et al. 2003).

In 1976, the Finnish Environment Institute (FEI, former National Board of Waters) started a project with the aim of acquiring more basic information on drilled wells in Finland (Rönkä & Turtiainen 1980, Rönkä 1983, 1988). During the project, data from about 700 drilled wells in southern and central Finland were collected.

In 1984, the present author started to gather information on drilled wells in Central Finland. In the beginning of the year 2009 the drilled well database contained information on 2,509 drilled wells; the database is regularly updated. The database is maintained at the Centre for Economic Development, Transport and the Environment for Central Finland (CETECF) in Jyväskylä, Central Finland.

Other Fennoscandian countries have their own national databases. The Geological Survey of Sweden (SGU) started to gather well information from various counties into a database in 1966 (Modig 1977). In 1976 a law made it compulsory for the well drillers to report information regarding all new drilled wells (Fagerlind & Modig 1977). There were 102,562 drilled wells in the SGU´s database at the end of March 1989 (Fagerlind 1989). The Well Archive (in Swedish ‘Brunnsarkivet’) has from then been growing with about 5,000-30,000 wells a year and contains now data from over 310,000 wells (Fagerlind 1987, Wallroth & Rosenbaum 1996, Törnros 2007, www.sgu.se 15/03/2011). Since 1992 well drillers in Norway have been legislatively obliged to provide information on drilled wells to the Geological Survey of Norway (NGU). At present, there are 51,400 drilled wells in the database of NGU (Gundersen & de Beer 2007; www.ngu.no 15/03/2011).

In Denmark, where groundwater has for many decades been extremely important in water supply because of the scarcity of surface waters, registration of wells has been compulsory since 1926; in the late 1970’s there were about 135,000 boreholes and wells in the database of the Geological Survey of Denmark (DGU; Möller 1978). At present, there are

251,500 boreholes and wells in the database; 125,000 of them are used for water supply (Martin Hansen, written communication, 05/02/2009).

The USGS National Water Information System (NWIS) contains extensive water data for the United States. Public access to many of these data is provided via NWISWeb (http://waterdata.usgs.gov/nwis/gw). The groundwater database consists of more than 850,000 records of wells, springs, test holes, tunnels, drains, and excavations in the United States. Available site descriptive information includes well location information such as latitude and longitude, well depth, and aquifer.

The TWDB Groundwater Database application is one of several Internet-based applications created and maintained by the Water Infrastructure Integration and Dissemination System (WIID), which is a section of the Texas Water Development Board (TWDB) in the United States. Of the nearly 1,000,000 water wells drilled in Texas in this century, approximately 125,000 are registered in the TWDB groundwater database. The WIID facilitates public access to reported water well information for all of those wells registered with the TWDB through a map-based interface. (http://wiid.twdb.state.tx.us/help/waterWellMapper.htm)

Since the mid-1970's water well drillers have been required by legislation to submit drilling reports to Alberta Environment, Canada. The database contains approximately 500,000 records with about 5,000 new drilling reports received annually. Information about individual water well drilling reports, chemical analysis reports up to the end of 1986, springs, flowing shot holes, test holes and pump tests conducted on the wells are available. Through this system anyone can view the information from the original report received by the Groundwater Information Centre. One will also be able to print the individual records plus print the overlaying map(s) that show the wells in relation to other wells and roads and so forth. (http://www.telusgeomatics.com/tgpub /ag_water/)

According to Southern African Development Community (SACD), the sub-Saharan countries operating some sort of organized water related databases are Botswana, Lesotho, Malawi, Mauritius, Mozambique, Namibia, South Africa, Swaziland and Zimbabwe (http://www.sadcwscu.org.ls/programme/groundwater/prog_groundprog_list3.htm). The National Groundwater Data Bank (NGDB) of South Africa contains information from approximately 200,000 boreholes. In Namibia and Zimbabwe the number of wells in the databases is 40,000 and 30,000, respectively. Rather unorganized databases exist in Angola, Ghana, Kenya, Tanzania, Uganda and Zambia (see also Kellgren & Sander 1997, Blecher 2000). A well-stored and fairly accurate database of more than 16,000 water wells has been managed in Burkina Faso with a good well distribution throughout the country (Courtois et al. 2010).

In Brazil, the Novo SIAGAS database, which covers the entire country, contains over 70,000 wells and more are being added continuously from existing records and new wells. It is accessible to all well drillers in Brazil via the Internet. Drillers will be able to enter all relevant information about new wells and submit it electronically to the central SIAGAS databank at Serviço Geológico do Brasil (CPRM). The database is accessible to any person equipped with a computer connected to the Internet, through CPRM's web site. From there, the user will be able to perform queries on the database, visualize lithological profiles and well construction details graphically, and retrieve data for specific wells in text form. (http://proasne.net/Databaseproject.htm)

The groundwater database of the Department of Sustainability and Environment of Victoria, Australia, contains information on approximately 135,000 boreholes throughout Victoria. (http://www.ourwater.vic.gov.au/monitoring/groundwater/groundwater-database)

6.1.2 Data sources and characteristics

The collection of drilled well information in Central Finland began in 1984 by compiling all available data into the Central Finland drilled well database (the CF database, Mäkelä 1994b, Fig. 30). The source data included, for example, groundwater investigation reports from consulting companies and local water works (e.g. Hokkanen et al. 1982), grant and subsidized loan application forms for improving the household water supplies in sparsely populated areas, and various documents built up with the guidance and supervision of groundwater resources and the Water Act.

During the following years, the CETECF executed systematic house-by-house surveys of existing wells in the field in connection with hydrogeological mapping and groundwater investigations. Indeed, as Sander (1999) states, the acquisition of borehole information from archives, reports and field surveys is an integral part of all groundwater projects.

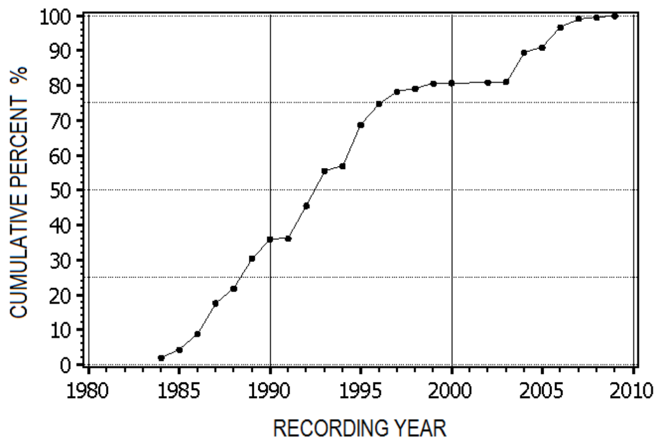

Fig. 30. Cumulative percent frequency of drilled wells in the CF database in one-year periods of the data recording. In the beginning of 2009 the total number of wells in the database was 2,509.

The vast majority of well information has been obtained from drilling companies (n=556) and by talking to the well owners on the phone and during water well inventories (n=1018). For example, the well completion data from a private drilling company (Vesto Oy, n=105), which consist of cable-tool wells drilled mainly in 1950's (Laakso 1966a, 1966b), are included in the well archive. Finnish Environment Institute has donated information from 219 wells drilled mainly by the same contractor in the 1970's (Rönkä 1983). Geological Survey of Finland has delivered information from 150 drilled wells (e.g. Idman 1997; Table 1).

During the years 1985-1989 the CETECF participated in a project, which aimed at developing common water supply in sparsely populated rural areas of central and eastern Finland (Mälkki 1983a, 1983b, 1989, 1990, Anon 1990). In this project a special attention was focused on siting and construction of drilled wells. Nearly 30 test wells, of which 2/3 were located in Central Finland, were drilled (Ihalainen & Öhberg 1988, Rinne 1988a, 1988b, Öhberg & Ihalainen 1988, Mäkelä 1990b, 1990c, 1990d).

In 1990-2008 the CETECF has continued the project by drilling 55 test wells in Central Finland. All well information collected in the course of these projects have been carefully stored, treated and finally added to the archive. The author has supervised around 50 well drillings in the field.

Table 1. Drilled well information sources in the CF database (total number of drilled wells n=2509).

Code	Data source	Frequency	Percent	Cumul. frequency	Cumul. percent
1	Well driller	556	22,2	556	22,2
2	Supervised drillings	56	2,2	612	24,4
3	Finnish Environment Institute	219	8,7	831	33,1
4	Owner of the well	1018	40,6	1849	73,7
5	Well loan application document	88	3,5	1937	77,2
6	Geological Survey of Finland	150	6,0	2087	83,2
7	Miscellaneous inventories and projects	355	14,1	2442	97,3
8	Well owner´s neighbor	62	2,5	2504	99,8
9	New owner of the well	5	0,2	2509	100,0

In test well studies, a thorough desk study and reconnaissance survey has been implemented and appropriate target sites for geophysical measurements have been identified based on favorable geomorphic and hydrogeological indications. Detailed fieldwork has then involved geophysical measurements, borehole drilling and pumping tests. The drilling sites have been chosen depending on the geophysical analysis and accessibility of the drilling team. Preferred user location, existing supply infrastructures and potential sources of groundwater contamination (current land use, groundwater protection) have also been important criterions during site selection.

The well information from each well has been arranged into a well form as a paper record attached with water quality analyses and a topographic map of the well region in a scale of 1:20 000 combined with a possible plan of the well site with greater accuracy. Each well has been identified with a six numbers long well identity number (city code + well number, e.g. 172092). The database has been computerized in the late 1980’s.

Most of the wells in the archive are private household wells drilled for domestic (n=1128) or farmhouse (n=922) use or for leisure purposes (n=87; Table 2). Around 110 wells have been drilled for the needs of industry and private enterprises. Nearly 250 drilled wells including 73 test wells have been constructed for public water supply (e.g. schools, villages, municipalities); in the beginning of 2009 roughly half of them delivered water to consumers. There are no energy wells in the archive.

Table 2. The purpose of construction of the drilled wells in the CF database (n=2509).

Code	Purpose of construction	Frequency	Percent	Cumul. frequency	Cumul. percent
0	Unknown	11	0,4	11	0,4
1	Domestic household	1128	45,0	1139	45,4
2	Farmhouse and livestock	922	36,7	2061	82,1
3	Summer house	87	3,5	2148	85,6
4	Industry, private enterprise	113	4,5	2261	90,1
5	Public institute	150	6,0	2411	96,1
6	Water works	25	1,0	2436	97,1
7	Test well	73	2,9	2509	100,0

Around 10% of the wells in the database have been abandoned by the owners for different reasons, for example, when the yield of a well has proven to be inadequate (5% of wells) or because of the poor water quality. Many of these well owners have later joined a piped community water system, e.g. a private cooperative organization of water supply.

6.1.3 Data file structure and description

The CF digital database consists of two ASCII data files: technical and water quality data files. The technical data file contains several types of information: well location data, well construction and performance data, geologic and topographic data, lineament data and well catchment area data. The water quality data file contains water quality data plus sampling and laboratory information. There may be several analyzed samples for a single well in the database. About 20% of the wells have missing water quality data. The data files can be linked with each other via the well identity number. The CF database consists of 132 fields of entries. The technical and water quality variables with the total number of entries for each variable in the database are listed and described in Tables 3 and 4, respectively.

As for optimal data processing, complete matrices of data are displayed in the database, where rows represent observations and columns represent variables (Table 5). However, the individual borehole records vary with respect to the completeness of data entries. As usual, every item of information is not available for every well and the number of analyzed data differs for studied parameters. This is a common finding in corresponding databases around the world (e.g. Daniel 1987, 1989, Morland 1997). The effect of these incomplete records will be seen in the statistical analyses that follow, especially for computations that are based on more than one variable. Then the final computations are based on no more than and commonly fewer observations than the smallest number of variable entries.

The datasets in the database are all "desktop datasets" (Sander 1997) that can be retrieved, interpreted and processed even without visiting the particular area being studied. The only prerequisite is that the analyst is an experienced user of similar data in a similar terrain.

The creation of the well database has been a time-consuming and laborious exercise, as a vast number of field, laboratory and desktop work including the data format and entry has been carried out with a high precision and accuracy. However, the accomplished and workable database allows easy and quick analysis and interpretation of different features, which aid, for example, at selection of promising sites for water wells.

Table 3. Technical variables and their total number in the CF drilled well database as 1/1/2009 (continued in the following page). *) Measured from the ground surface.

Variable group and abbreviation	Explanation and dimension	Precision	Number of data entries	Percent of the total data entries
LOCATION				
ID	Well identity number (city code + well number)	-	2509	100,0
MAP	Number of topographic map sheet (scale 1:20 000)	-	2497	99,5
NCOORD	North coordinate (national grid coordinate system) [m]	10	2479	98,8
ECOORD	East coordinate (national grid coordinate system) [m]	10	2479	98,8
CONSTRUCTION				
CY	Construction year	1	1672	66,6
DR	Drilling contractor	-	1548	61,7
ASL	Height of the well site in meters above sea level [m a.s.l.]	1	2479	98,8
DROV	Drilled well not penetrating bedrock [0/1]	-	2033	81,0
CAS	Length of casing [m] *)	0,5	858	34,2
GWT	Depth to static groundwater table [m] *)	0,5	788	31,4
GWM	Month when the GWT was measured [1-12]	1	377	15,0
STR	Depth to first (main) water strike [m] *)	0,5	813	32,4

FRAC	Fracturing degree of the bedrock at well site [1-3]	-	580	23,1
DEPTH	Total depth of well [m] *)	0,5	2033	81,0
DIA	Diameter of well [mm]	1	1404	56,0
PUMP	Installation depth of well pump [m] *)	0,5	882	35,2
USE	Use of well [0/1/2]	-	2020	80,5
WDEM	Water demand of well owner [m^3d^{-1}]	0,1	754	30,1
OWN	Well owner´s surname/well´s name [character variable]	-	2033	81,0
PURP	Construction purpose of well	-	2498	99,6
SOUR	Well information source	-	2509	100,0
REC	Recorder of well information	-	2509	100,0
DATE	Year of well information recording	1	2509	100,0
PERFORMANCE				
YIELD	(Short-term) yield [Lhr^{-1}]	5	1320	52,6
YIELM	Month when yield was measured [1-12]	1	927	37,0
SCS	Short-term specific capacity [$Lhr^{-1}m^{-1}$]	5	64	2,6
SCL	Long-term specific capacity [$Lhr^{-1}m^{-1}$]	5	32	1,3
DEVE	Well development [0/1/2]	-	582	23,2
DEVY	Yield after well development [Lhr^{-1}]	5	76	3,0
DSCS	SCS after well development [$Lhr^{-1}m^{-1}$]	5	12	<0,01
SUF	Water sufficiency [0/1]	-	1337	53,3
GEOLOGY				
SOIL	Soil type at well site [1-5]	-	2479	98,8
OVER	Thickness of overburden at well site [m] *)	0,5	1565	62,4
ROCK	Rock type at well site [1-22]	-	2479	98,8
ROCM	Main rock type in the catchment area [1-22]	-	2479	98,8
PKIN	Post-kinematic rock type at well site [0-3]	-	2479	98,8
HSL	Highest shore level of the Baltic Sea at well site [m a.s.l.]	1	2479	98,8
UPLIFT	Annual land uplift rate at well site [mm a^{-1}]	0,1	2479	98,8
TOPOGRAPHY				
TOPO	Topographic setting of well [1-4]	-	2479	98,8
ASLH	Highest point in the catchment area [m a.s.l.]	1	2479	98,8
ASLL	Lowest point in the catchment area [m a.s.l.]	1	2479	98,8
HDIS	Distance of well to the nearest (bedrock) hilltop [m]	5	986	39,3
HASL	Height of the (bedrock) hilltop [m a.s.l.]	1	986	39,3
LINEAMENT				
LDIS	Distance of well to the nearest lineament [m]	1	752	30,0
LPRO	Prominence of the nearest lineament [1-3]	-	752	30,0
LFRE	Lineament frequency in the catchment area	-	752	30,0
LLEN	Total length of the lineaments in the catchment area [m]	10	752	30,0
LINF	Lineament intersection frequency in the catchment area	-	752	30,0
LIND	Distance of well to the nearest lineament intersection [m]	5	202	8,1
LASL	Height of the bottom of the nearest lineament [m a.s.l.]	1	752	30,0
LWAT	Water course in the nearest lineament [0/1]	-	752	30,0
LAZI	Azimuth of the nearest lineament [º]	5	752	30,0
LOSI	Position of well with respect to the nearest lineament [º]	5	752	30,0
CATCHMENT				
WDIS	Distance of well to the nearest water course [m]	5	1651	65,8
WASL	Height of the nearest water course [m a.s.l.]	1	1651	65,8
RDIS	Distance of well to the nearest public road [m]	5	1672	66,6
RMAI	Maintenance class of the nearest public road [1-5]	-	1672	66,6
RASL	Height of the nearest public road [m a.s.l.]	1	1672	66,6
BED	Bedrock outcrops in the catchment area [%]	4	2479	98,8
FOR	Forest land area in the catchment area [%]	4	2479	98,8
PEA	Peat/paludified forest land in the catchment area [%]	4	2479	98,8
CUL	Cultivated land in the catchment area [%]	4	2479	98,8
SET	Settled area in the catchment area [%]	4	2479	98,8
WAT	Water courses in the catchment area [%]	4	2479	98,8
Σ 64			Σ 100659	

Table 4. Water quality variables and their total number in the CF drilled well database as 1/1/2009 (continued in the following page).

Variable abbreviation	Explanation and dimension	Precision	Number of data entries	Percent of sampled wells
ID	Well identity number (city code+well number)	-	3651	81,0
DMY	Date of water sampling [day-month-year]	-	3560	79,1
TEMP	Temperature [º C]	0,1	1152	30,2
COLO	Colour [mgl^{-1} Pt]	5	1584	42,1
TURB	Turbidity [FNU]	0,01	1518	37,4
O2	Dissolved oxygen [mgl^{-1}]	0,1	727	16,9
CO2	Dissolved carbon dioxide [mgl^{-1}]	0,1	679	17,0
PH	pH	0,1	2090	53,2
COND	Electrical conductivity [25º C mSm^{-1}]	0,1	2011	50,9
ALK	Alkalinity [$mmoll^{-1}$]	0,01	1205	32,1
HCO3	Bicarbonate [mgl^{-1}]	0,1	1205	32,1
HARD	Total hardness [$mmoll^{-1}$]	0,01	1475	38,6
COD	Oxidisability (COD_{Mn}-O_2) [mgl^{-1}]	0,1	2593	66,5
KMNO4	$KMnO_4$ consumption [mgl^{-1}]	0,1	2593	66,5
TOC	Total organic carbon [mgl^{-1}]	0,1	323	7,8
NO3	Nitrate [mgl^{-1}]	0,001	2055	54,2
NO2	Nitrite [mgl^{-1}]	0,001	1755	46,4
NH4	Ammonium [mgl^{-1}]	0,001	1593	41,3
TOTP	Total phosphorus [mgl^{-1}]	0,001	810	19,5
PO4P	Phosphate phosphorus [mgl^{-1}]	0,001	128	4,7
SO4	Sulphate [mgl^{-1}]	0,1	1137	32,0
S	Sulphur [mgl^{-1}]	0,01	6	0,2
CL	Chloride [mgl^{-1}]	0,1	1627	41,4
BR	Bromide [mgl^{-1}]	0,001	86	3,4
F	Fluoride [mgl^{-1}]	0,01	1224	35,4
SIO2	Silica [mgl^{-1}]	0,1	719	22,0
CA	Calcium [mgl^{-1}]	0,1	948	27,7
MG	Magnesium [mgl^{-1}]	0,1	927	27,5
NA	Sodium [mgl^{-1}]	0,1	969	25,3
K	Potassium [mgl^{-1}]	0,1	882	24,6
AG	Silver [μgl^{-1}]	0,01	87	3,5
AL	Aluminium [μgl^{-1}]	1	497	13,0
AS	Arsenic [μgl^{-1}]	1	450	11,7
B	Boron [μgl^{-1}]	1	156	5,0
BA	Barium [μgl^{-1}]	1	97	3,8
BE	Beryllium [μgl^{-1}]	0,1	87	3,5
BI	Bismuth [μgl^{-1}]	0,01	87	3,5
CD	Cadmium [μgl^{-1}]	0,1	563	15,6
CO	Cobalt [μgl^{-1}]	1	141	5,5
CR	Chromium [μgl^{-1}]	1	468	12,5
CU	Copper [μgl^{-1}]	1	560	15,6
FE	Iron [μgl^{-1}]	1	2828	69,3
HG	Mercury [μgl^{-1}]	0,1	59	1,2
LI	Lithium [μgl^{-1}]	0,1	36	1,4
MN	Manganese [μgl^{-1}]	1	2592	66,1
I	Iodine [μgl^{-1}]	0,01	6	0,2
MO	Molybdenum [μgl^{-1}]	1	93	3,7
NI	Nickel [μgl^{-1}]	1	562	15,5
PB	Lead [μgl^{-1}]	1	563	15,5
RB	Rubidium [μgl^{-1}]	1	93	3,7
SB	Antimony [μgl^{-1}]	1	152	4,9

SE	Selenium [µgl^{-1}]	1	152	4,9
SN	Tin [µgl^{-1}]	0,1	6	0,2
SR	Strontium [µgl^{-1}]	1	97	3,8
TH	Thorium [µgl^{-1}]	0,01	87	3,5
TL	Thallium [µgl^{-1}]	0,01	87	3,5
U	Uranium [µgl^{-1}]	1	546	15,8
V	Vanadium [µgl^{-1}]	1	93	3,7
ZN	Zinc [µgl^{-1}]	1	563	15,6
RA	Radium [Bql^{-1}]	0,01	118	4,5
RN	Radon [Bql^{-1}]	1	969	28,0
TA	Total alpha activity [Bql^{-1}]	0,01	163	6,2
BACT	Bacteria [0=none/1=observed]	-	1249	30,5
LAB	Analyzing laboratory [1-11]	-	3651	81,0
RALAB	Analyzing laboratory for radionuclides [1-14]	-	1126	32,7
REP	Representativeness of the sample [1-6]	-	3651	81,0
CLA	Sample classification index [1-10]	-	3651	81,0
STA	Statistical index [1-5]	-	3651	81,0
Σ 68			Σ 71269	

Table 5. Example of data lines (n=37) in the technical data file of the Central Finland drilled well database. The exact well coordinates and the well owner´s name are not shown. The explanations of different abbreviations are given in Table 3.

```
     W           N     E                                                                   U
     E           C     C                    D                      YY                      P
     L           O     O       D          F E      P    W PS  D    II   D D D S O R RP     LT  A  A  H  H   LLL   LL   L  LL  L  L  W  W R R  R
     L     M     O     O      AR C G G   SR P  D   UU   DOUOR A    EE SSE E SSO V O OK  H  IO  S  S  D  A   DPF   LI   I  AW  A  O  D  A D M  A B F P C S W
     I     A     R     R C D  SO A W W   TA T  I   MS   EWRUE T    LL CCV V CUI E C CI  S  FP  L  L  I  S   IRR   EN   N  SA  Z  S  I  S I A  S E O E U E A
     D     P     D     D Y R  LV S T M   RC H  A   PE   MNPRC E    DM SLE Y SFL R K MN  L  TO  H  L  S  L   SOE   NF   D  LT  I  I  S  L S I  L D R A L T T
17929032110868XXXX34XXXX78 2105. . . .   .. 90110  .1  0.9X25195    . ....  . ..2 . 4 4.1444.52123 79560123   ...    ..  .   ..  .  .350 79   ..  . 440 04012 4
17929132110268XXXX34XXXX81 3103. . . .   ..116   .  .1  4.0X242854000 ....  . .12 . 2 2.1524.64155 823801251401214001440 850105195  .   .4504115 456 432 4 0
17929232110268XXXX34XXXX . 3105. . . .   ..    .   .  .1  4.0X24285    . ....  . .12 . 2 2.1524.62155 833201251551214001510 850105195  .   .4004115 044 444 8 0
17929332110268XXXX34XXXX62 .122. . . .   .. 51100 201  3.0X24296    . ...0  . .12 1 2 2.1524.61148 85 501253401216600  .1071145 55230107 604122 032 44016 8
17929422341068XXXX25XXXX82 3178.12 . .   ..120   .  .1  1.6X25284 400 1...  . .12 1 5 5.1384.8121314211018521022196013801630150240  .   . 9041771256 420 8 0
17929522331268XXXX25XXXX61 .190. . 2 .   .. 30   .  .1  0.9X25284 700 ....  . .12 21010.1434.73203178360203   ...    ..  .   ..  .  .210178   ..  . 056 436 4 0
17929622330968XXXX25XXXX6114200. . . .   .. 80   .  .1  0.6X25284 100 ....  . .12 2 4 4.1424.712031471002033303212000  .1750 90180   .   .   ..  . 076 416 4 0
17929722341068XXXX25XXXX61 . 98. . . .   .. 67   .  .1  1.0X25284 300 ....  . .12 2 2 2.1404.83114 91   .   .  ...    ..  .   ..  .  .130 91   ..  .1240 024 816
17929832110868XXXX34XXXX73 3 90. . 4 . 60. 98   .  .1  3.0X24489    . ...0  . .11 2 2 2.1434.44188 79   .   . 802217801200 970165 75190 791454 791640 032 4 8
17929932110568XXXX34XXXX89 3 81.10 . .   .. 73110 501   . X14290    . ...0  . .12 11818.1484.54150 7923010513031 8400   . 791  5275 30 7956031454020 0 0 436
17930032110468XXXX34XXXX90 3 85.10 612 701103155 701   . X24290200011..0  . ..1 11818.1464.44138 7925011012031 8000   . 810125215 60 79   ..  . 836 032 816
17930122331168XXXX25XXXX . .126. . . .   ..100   .  ..   . .261921500 ....  . ..2 . 2 2.1514.62146 98100137   ...    ..  .   ..  .  .310 98 304125 052 036 4 8
17930222341169XXXX25XXXX . .173. . . .   .. 70   .  ..   . .26192    . ....  . ..2 4 4 4.1424.8220014330019240011 9000  .1430 25295280143 6041721644 02412 4
17930322341068XXXX25XXXX70 .200. . . .   4. 21100 181  1.5.241933500 ...0  . ..2 0 5 5.1394.812101582402071803214000  .1840 60330   .   .   ..  .1256 424 4 0
17930422341169XXXX25XXXX93 2148. 3 1 . 101125100 501   . X21394 500 6..0  . .12 3 4 4.1434.8219313021019321023210013801480160250510130   ..  .1280 4 4 0 0
17930532110668XXXX34XXXX93 3 95.10 5 .   ..157150 801   . X34194120012..0  . .12 7 2 2.1474.53115 79   .  .1801220600   . 790 45315 80 79   ..  . 456 0 4 432
17930632110568XXXX34XXXX94 3 82.14 . . 10. 84150 501  1.0X141951000 4..0  . ..2101818.1494.54115 79   .   . 803220000   . 791 50140 25 793103 811216 016 848
17930722341068XXXX25XXXX88 .135. . . .   ..234   .  .1  2.0X15195    . ....  . ..2 . 5 5.1404.82171 93310171   ...    ..  .   ..  .   .   .  . 3031383656 0 4 4 0
17930822341068XXXX25XXXX85 .140. . . .   .. 70   .  ..   . X15195    . ....  . ..1 . 5 5.1404.82171 91390158   ...    ..  .   ..  .   .   .  .10031383248 4 412 0
17930922331168XXXX25XXXX89 3 93. 4 6 . 75.105150 801  2.5X241953000 ...0  . .14 1 2 2.1514.63110 80   .  .2602215801430 850175 85280 80 704 92 020 86012 0
17931032110268XXXX34XXXX94 3 95.10 . .   ..133163  .1  1.2X11595 900 5...  . .11 0 2 2.1514.63108 79300105   ...    ..  .   ..  .  .180 792903 983620 0121220
17931132110268XXXX34XXXX . . 88. . . . 70. 85   .  ..   . .231951500 ....  . ..2 . 2 2.1534.63146 79   .   .  ...    ..  .   ..  .  .180 79 404 881216 0401616
17931222331168XXXX25XXXX64 .103. . . . 39. 39   .  .0   . X241961000 ...0  . .14 . 2 2.1454.63136 845301231902220601540 970  5 95   .  .2000 89 844 03216 0
17931322331168XXXX25XXXX . .100. . . .   .. 52   .  .0   . X23196 900 ....  . .14 . 2 2.1454.64139 835401231002221201490 960  5 95   .  .2600 89 448 03612 0
17931422331168XXXX25XXXX .13 97. . . .   .. 50   .  .0   . X24196    . ...0  . .14 2 2 2.1464.63140 82   .  .2402322601430 930  5 95   .  . 900 89 432 056 8 0
17931522341068XXXX25XXXX96 3148.10 . .1201145160  .1   . X21196 120 1...  . .12 3 1 1.1424.81158104 50151   ...    ..  .   ..  .   .   .  .1104146 472 020 4 0
17931622341169XXXX25XXXX78 3145. . 5 .   .. 24150 221   . X14196    . ...0  . .12 2 5 5.1414.841761192701581301216000  .1370 35125290119 7041382468 0 4 4 0
17931732110268XXXX34XXXX86 .125. .20 . 20. 50   .  .1   . X242961000 ...0  . .12 1 2 2.1524.61148 84 301253801215600  . 8801102002801071204123 424 05212 8
17931822331168XXXX25XXXX96 3122. 2 . . 94. 97160 501   . X14296    . 1..0  . .12 2 2 2.1524.61180 87 551252601215600  .1071115 25140107 304122 428 4361216
17931922341169XXXX25XXXX96 3124.12 . .   .. 97160 701  0.5X21196 500 3...  . .12 . 5 5.1414.82156112205141   ...    ..  .   ..  .  .370116 554124125216 412 4
17932022341068XXXX25XXXX96 2 95.13 . .100.1331401001  1.0X11196    . 6..0  . .12 9 1 1.1414.84156 91450156 953211001190 921 45135 40 921403100 848 4 8 824
17932122331168XXXX25XXXX96 3103.24 . .   .. 91160  .1   . X111961100 8..0  . .12 . 2 2.1454.64137 823001201402321201400 930  0270370 843600 87 048 03616 0
17932232110268XXXX34XXXX96 3 86.10 7 .120.138162 901  1.0X11197 40010..0  . .12 1 2 2.1514.62119 793201071601111000  . 791170260 30 79   ..  .2816 0 0 452
17932332110468XXXX34XXXX90 3 98. . . .   .. 45160 411   . X14197    . 7...  . .1214 1 1.1454.42158 84300148   ...    ..  .   ..  .   .   .   .  ..  . 864 420 4 0
17932432110468XXXX34XXXX97 2 83.19 4 .1001121140  .1 50.0X72197  50 7..125012.2101818.1464.44163 79   .   . 251111300   . 830135 45170 79   ..  . 056 02016 8
17932532110568XXXX34XXXX96 3 84.10 2 . 60.157160 651   . X111971000 7..0  . .12 4 1 131474.54138 79   .  .1401220001370 791  0 90 50 79   ..  . 436 020 832
17932632110368XXXX34XXXX97 3 92.10 0 7 74. 74160  .1  0.3X111972000 7..0  . .12 2 2 2.1514.64154 82   .   . 252218001180 900130 40   .   .   ..  . 856 028 8 0
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6.1.4 Reliability of questionnaire data

In the early 1990's, the author statistically tested the reliability of well construction data donated by well owners by comparing well owners' data with the data delivered by well drillers from the same wells. The drillers´ data were assumed correct. Besides testing the overall reliability of the questionnaire data, it was aimed at finding out whether the age of wells has any influence on well information obtained from well owners.

In order to run the reliability test, 50 drilled wells in all were chosen from the CF database by random sampling and a questionnaire form for each well was sent to well owners in 1993. The well data that were questioned included construction year of well, well drilling contractor, overburden thickness at well site, total well depth, diameter of well, and yield of well. From the 50 questionnaires mailed 42 (84%) were returned bearing information about the wells (Appendix II).

The normality of the questionnaire data was tested with the Shapiro–Wilk test (Shapiro & Wilk 1965). Its null hypothesis assumes that sample data come from a normally distributed population. Because the normality assumption for the test data was not satisfied or the sample sizes were too small and because the observations in the two data sets were related, the reliability of the questionnaire data was tested with the non-parametric paired sample Wilcoxon sign rank test (e.g. SAS 1990c). This method is used to test whether the means of two dependent datasets are equal (null hypothesis). Paired sample Wilcoxon sign rank test uses magnitude and sign of the paired difference ranks.

The test results showed that the distributions of the two data sets were not statistically different from each other at $\alpha \leq 0{,}05$ risk level (Table 6). Considering the well age influence only the total well depth distributions from the 1970's were statistically different at $\alpha \leq 0{,}05$ risk level. All well owners knew their drilling contractor. Based on the test results it was concluded that well information obtained from well owners could be considered statistically reliable regardless of its age.

6.1.5 Data quality filtering

A good well database is, naturally, a prerequisite for any statistical analysis of well characteristics. According to Banks et al. (2010), such a database should ideally be:

(1) Large: in order to provide the validity and a good degree of confidence in data screening with different statistical methods (e.g. Courtois et al. 2010);

(2) Representative: for example, there may be a clear danger that "failed" (i.e. poorly yielding) wells will tend to be under-reported to national databases, relative to successful wells that are eventually taken into production (e.g. Graham et al. 2009);

(3) Reliable: well data, often of variable quality, should be quality-filtered to secure reliable statistical analysis (e.g. Morland 1997, Henriksen 2008).

It is important to understand the possible sources of error connected to any large data set collected over an extended period (e.g. Morland 1997). As Sander (2007) states, quite commonly the well data exhibit poor spatial completeness, heterogeneous coverage of key parameters, difficult control of uncertainty and incompatible coordinate systems.

An essential requirement is to be able to record the locations of the wells accurately on the maps (Wladis 1995, Sander 1996, 1999, Martin & van de Giesen 2005, Kenny et al. 2006, Misstear et al. 2006). In the CF database, wells with initially uncertain locations have been located as accurately as possible in connection of fieldwork and by making phone calls to the well owners. Most well locations in the dataset are accurate to ±10…20 m, at least. The locations of 30 wells (1,2%) are not available in the database; when required these wells were omitted from statistical analyses.

Table 6. Results of the reliability test of the questionnaire data. The explanations of different abbreviations are given in Table 3. *) All questionnaire entries correct.

CY	Variable	n	Shapiro–Wilk test	p	Wilcoxon sign rank test	p
All wells	DR	42	..	..	..	*)
	CY	38	0,335628	<0,0001	4	0,6719
	OVER	35	0,848964	<0,0001	0	1,0000
	DEPTH	40	0,726070	<0,0001	-44,5	0,0519
	DIA	34	0,621603	<0,0001	27,5	0,1995
	YIELD	36	0,629625	<0,0001	2	0,9299
1950's	CY	3	..	..	..	*)
	OVER	2	1,000000	1,0000	-0,5	1,0000
	DEPTH	6	0,908053	0,4067	-2	0,6250
	DIA	3	0,978067	0,7161	-0,5	1,0000
	YIELD	5	0,697419	0,0105	-5	0,1250
1960's	CY	3	0,749986	<0,0001	-0,5	1,0000
	OVER	3	0,923059	0,4632	1,5	0,5000
	DEPTH	3	0,749986	<0,0001	-0,5	1,0000
	DIA	3	0,749986	<0,0001	-0,5	1,0000
	YIELD	3	0,998666	0,9302	2	0,5000
1970's	CY	9	0,844513	0,0643	1	1,0000
	OVER	9	0,837072	0,0534	3,5	0,3750
	DEPTH	8	0,954912	0,7616	-13	**0,0313**
	DIA	7	0,851910	0,1312	4	0,3750
	YIELD	7	0,778538	0,0246	3,5	0,5625
1980's	CY	9	0,432051	<0,0001	1,5	0,5000
	OVER	8	0,416854	<0,0001	-0,5	1,0000
	DEPTH	9	0,467162	<0,0001	0	1,0000
	DIA	9	0,645631	0,0004	7,5	0,0625
	YIELD	7	0,456937	<0,0001	-0,5	1,0000
1990's	CY	14	0,293001	<0,0001	-0,5	1,0000
	OVER	13	0,901028	0,1346	-5	0,4531
	DEPTH	14	0,514748	<0,0001	3	0,2500
	DIA	12	0,627625	<0,0001	-1	0,8750
	YIELD	14	0,462278	<0,0001	0,5	1,0000

One characteristic with the CF database is that the wells have been drilled over a large span of time (about 60 years). This may cause problems when comparing different wells. Other data bias may arise because of various technical practices between well drillers (e.g. cable-tool and DTH wells, different methods in yield determination). Selecting homogeneous subsets when analyzing the data sets the possible problems can be reduced. In addition, the fact that the data has been collected from a number of sources, can further affect the accuracy of the well information. Care should be taken, especially, when using secondary data sources, such as project reports from previous investigations (Sander 1999, Solomon & Ghebreab 2008). Borehole data of Central Finland have been checked against the primary data source whenever possible.

As regards to the CF database, a comprehensive data-quality filtering has been done for the raw data in an initial stage, before the well information has been converted into a digital database. During this preliminary stage, many variable entries have been rejected because of doubtful or inaccurate data (e.g. depth or yield of wells given as approximations or within a range of values). During the years, many quality checkings and controls have been executed to reaffirm the correctness of the well information in the database. For instance, well construction data from more than 700 drilled wells were checked by phone calls with well owners in the summer of 2000. Indeed, a database should be regarded as a living organism, which needs steady care, use and development (Sander 1999).

Sometimes drillers do not report failed or 'dry' boreholes, such that the low-yield end of a yield-distribution curve may be under-representative (e.g. Ajayi & Abegunrin 1994, Banks et al. 2005). However, both wet and dry boreholes are a valuable source of data. Otherwise, any statistics compiled will be highly misleading and over-optimistic (e.g. Foster 1984, Lewis 1990, Cobbing & Davies 2008). The CF database covers wells regardless of their productivity, so also dry ones are included. Yields from hydrofractured or shooted boreholes have been rejected from the study material unless their initial yields have been available.

Seventeen drilled wells (0,7 %) in the CF database do not extend to the bedrock. Instead, they are left as open-bottom cased boreholes in the overlying soil mantle (maximum well depth 47 m). For inclusion in statistical analyses, a well had to be drilled into bedrock. Technical information given by the well-owner´s neighbor (n=62) or the new owner of the well (n=5) was eliminated from the statistical material of the study.

It is important that well data represent a random sample regarding geographic and geologic position. Likewise, areas having more detailed information should not be treated differently to those having very little data (Wladis 1995, Wladis et al. 1997). This kind of bias cannot entirely be eliminated from any data set (e.g. Knopman 1990). Private well borers usually construct a borehole as near the user as possible, generally not seriously taking into consideration other factors which might influence the yield (e.g. Morland 1997, Henriksen 2006b). Thus, private wells in the CF database can be considered randomly sited in a hydrogeological sense. To the contrary, test wells cannot be considered randomly sited, because geological and hydrogeological expertise and various geophysical methods have been applied in well siting (e.g. Mäkelä 1990b, 1990d, 1993, Penttinen et al. 2000).

In that what follows, the wells have been divided into three groups: private wells, test wells and all wells. In this way the data set may be considered to be compiled objectively.

In order to make a reliable statistical analysis from an initially large volume of data, the removal of outliers and extreme values may sometimes be necessary (e.g. Morland 1997, Henriksen 2003a, 2003b, 2006b). This kind of rejection of data thus usually causes a reduction in database records. For the purposes of this study, no outliers or extreme values were eliminated except a 505 meters deep well no. 182081 at Himos in the municipality of Jämsä, which initially was aimed not only at looking for water but also at testing the capability of the drilling equipments (KSML 1986). It is important to note that rejecting the aberrant values from a database when they are real may sometimes lead to serious errors. Indeed, they may be the only values that contain valuable information (e.g. Berthouex & Brown 1994). Outlier data should, therefore, be dealt with explicitly (Morland 1997, Henriksen 1995, 2003a, 200b).

In all, 84 wells were excluded and 2,425 wells were retained in the technical database for further statistical analyses.

6.2 Determination of hydraulic parameters

Pumping tests with water table measurements are rarely made in connection with crystalline-rock boreholes (e.g. Henriksen 2003a, Kenny et al. 2006). The well yields (Q) are rather obtained by various short-term tests carried out by drillers at the termination of the drilling and reported as liters per hour (Lhr^{-1}; Woolley 1982, Kaehler & Hsieh 1994, Loiselle & Evans 1995, Williams et al. 2004, Henriksen 2006b).

In Central Finland, well drillers carry out both air-lift tests and short-term pumping tests with submersible pumps. The amount of drawdown in the wells usually is not measured, nor is the duration of pumping time. The water blown out of the borehole during air-flush drilling is measured by channeling the water away from the borehole through a pipe into a bucket of known volume and taking the time to fill the bucket (Fig. 31). Yield estimations could be made for all the measurable water strikes every time a new strike of water is observed during drilling. In comparison, water well drillers in the United States commonly conduct a well-performance test by pumping a well with a constant rate and measuring the drawdown in the well until the change in water level is small (pseudo-steady state; Mace 2001).

Fig. 31. Yield determination during air-flushing of the Ristimäki test well no. 077081 in the municipality of Hankasalmi, Central Finland, in August of 2005. Drilled well depth at the moment the photo was taken was 61 m and Q ~ 2,500 Lhr^{-1}; the final well depth was 91 m and Q = 9,000 Lhr^{-1}. Photo Kari Illmer/CETECF.

The air-lift yield may only give a rough measure of the actual well yield, and air-lifting may overestimate small yields and underestimate high yields (Banks 1991, Johansson 2005, Henriksen & Braathen 2006). On the other hand, Paillet and Duncanson (1994) have found that the air-lift flow rates estimated during drilling agree qualitatively and quantitatively with the flow rates estimated during short-time test pumping. Short-term yield measurements usually reflect the local properties of the media and the state of the hydraulic system in the vicinity of the well (Banks 1992b, Henriksen 1995, Meier et al. 1999, Mace 2001, Banks et al. 2010).

Long-term sustainable well yields might be considerably lower than short-term yields (Daniel 1990, Banks et al. 1992a, 1994, 2005, Kaehler & Hsieh 1994, Kastrinos & Wilkinson 1994, Morland 1997, Acheampong & Hess 1998, Banks & Robins 2002, Misstear et al. 2006). In Piedmont crystalline rocks of North Carolina, USA, long-term well yields were found to be about 75 percent of those predicted on the basis of 24-hour pumping tests and only about 60 percent of the driller's reported yields (Harned & Daniel 1989). According to Mäkelä (1993), short-term yields are decent estimations of sustainable yields for crystalline rock aquifers in Central Finland when the yields are less than 1,000 Lhr^{-1}; high short-term well yields are on the average twice that big as the corresponding sustainable yields. Verma (2003) has estimated that sustainable well yields may only be from 20 to 30% of the accompanying air-lifted yields. Parizek and Siddiqui (1970) suggest that from 50 to 95% of a well´s short-term capacity may be lost as localized producing zones are dewatered.

Although short-term yields may well not be representative of long-term sustainable yields, they suit for calculation of hydraulic properties of bedrock aquifers, especially in light of their intended purpose of evaluating general, regional trends (Carlsson & Carlstedt 1977, Briz-Kishore 1993, Krásný 1993a, Kastrinos & Wilkinson 1994, Paillet & Duncanson 1994, Rohr-Torp 1994a, Henriksen 1995, 2003a, 2006b, Loiselle & Evans 1995, Wladis 1995, Sami 1996, Wallroth & Rosenbaum 1996, Morland 1997, Wladis & Gustafson 1999, Drew et al. 2001, Gustafson 2002, Moore et al. 2002a, Kenny et al. 2006).

6.2.1 Normalized yield

In this study, the normalized well yield (Q/d_s $Lhr^{-1}m^{-1}$) was determined by dividing the yield of a well (Q Lhr^{-1}) by the saturated thickness of open well section (SAT m; e.g. Davis & Turk 1964, Siddiqui & Parizek 1971), i.e.

$$Q/d_s = Q/SAT \tag{19}$$

The saturated open well section (SAT) was calculated as the difference between the depth to the static water level (GWT m) and the total depth of the well (DEPTH m) when the static water level was below the bottom of the casing, i.e.

$$SAT = DEPTH-GWT \tag{20}$$

If the length of the casing (CAS m) was greater than the depth to the static water level, the SAT was equal to the total depth of the well below the bottom of the casing, i.e.

$$SAT = DEPTH-CAS \tag{21}$$

(e.g. Davis & Turk 1964, Siddiqui & Parizek 1971). The yield of a dry well was entered into calculations as a nominal figure of 1 Lhr^{-1} (see e.g. Banks et al. 2005).

Missing overburden thickness (OVER, about 20% of the boreholes with depth and yield information) was calculated as the overall median of the overburden thickness (3 m). Missing water table (GWT, about 50% of the boreholes with depth and yield information) was calculated as the overall median of the water level data (5 m). Boreholes with missing casing data (CAS, about 45% of the boreholes with depth and yield information) were assigned a casing depth equal to the overburden thickness plus the overall median difference between casing depth and the overburden thickness (4 m). The wells with casing depth equal to overburden thickness were omitted from statistical analyses.

The approximations outlined above are largely the same as the ones used, for example, by Antal et al. (1998) and by Berggren (1998) for the Swedish evaluation of bedrock hydraulic properties and by Henriksen (2006b, 2008) for those of Norwegian and Swedish bedrock wells.

6.2.2 Specific capacity and well productivity

Short-term step-drawdown tests were carried out in 64 drilled wells in order to determine the specific capacities (Q/s_m $Lhr^{-1}m^{-1}$) of the wells. In step-drawdown tests the wells are pumped at a constant pumping rate until approximate stable drawdowns are reached (e.g. Kruseman & de Ridder 2000). Another type of step tests are constant-head drawdown tests, where a steady drawdown is reached and maintained by slightly varying the discharge Q (e.g. Labadie & Helweg 1975, Kath et al. 2004). Half of the wells were test wells intended for common water supply, the other half were private wells. The yield of the step-test wells was in a similar range than that of other wells in the CF database.

The step tests were done with electrical submersible pumps or with centrifugal pumps situated on the ground surface and driven by a combustion engine. In low-yield wells a variable frequency drive pump was used with an electronic controller to match motor speed to the need. A gauging-bucket was used to monitor the pumping discharge and a water table measuring tape or a water-level recorder to monitor the drawdown inside the borehole.

The tests consisted of 1 to 4 steps with stepwise increased pumping rates and drawdowns. The length of a single step was determined by how long it took for the water level in the well to reach a state of apparent equilibrium, that is, when the change in drawdown was minimal with time (e.g. Bradbury & Rothschild 1985). Because of this stabilization, no standardization of well specific capacity for pumping period was considered necessary (e.g. Brook 1988). The total duration of each test was from half an hour to six hours. Specific capacity for each well was calculated from the first step by dividing the pumping discharge (Q Lhr^{-1}) with the corresponding drawdown s_m (m). Private wells included cable-tool wells, which had been drilled and pumped with single-acting piston pumps mainly in the late 1950´s by the same well contractor.

In order to make comparisons between wells, recorded specific capacities should be adjusted to a common well radius. On the other hand, differences in diameter do not greatly affect specific capacities (Walton 1977). The dewatering of the most productive fractures near the top of the rock results in a pronounced decrease in specific capacity with increasing pumping rate (Leach 1982, Heath 1992, Mace 2001).

Well losses were determined in seven wells with the Hantush-Bierschenk´s method (Bierschenk 1963, Kruseman & de Ridder 2000) based on Jacob´s (1947) equation

$$s_w = BQ + CQ^2 \tag{22}$$

where s_w is the drawdown in the pumped well (m), B is the aquifer constant (sm^{-2}), C is the well-loss constant (s^2m^{-5}) and Q is the pumping rate (m^3s^{-1}).

The Jacob´s equation can be rewritten as

$$s_w/Q = B + CQ \tag{23}$$

Values for B and C were determined by computations involving increments of drawdown and pumping rate for successive steps (see Eagon & Johe 1972, Mandel & Shiftan 1981, Kruseman & de Ridder 2000). The values of specific drawdown s_w/Q for each step were plotted against the total pumping rate on plain coordinate paper and a straight regression line was fit through the plotted points by using least squares (Fig. 32). The value for B was determined from the intercept with the s_w/Q axis, and C was the slope of the line (Bruin & Hudson 1955, Eagon & Johe 1972, Clark 1977, Mandel & Shiftan 1981, Kruseman & de Ridder 2000). Because a steady state was reached in each step, the drawdown was not time dependent. The steady state drawdown in each step was then the true drawdown for that discharge rate (Eagon & Johe 1972, Uhl & Sharma 1978, Mandel & Shiftan 1981, Mackie 1982, Mace 1997, 2000, 2001). The well-loss corrected specific capacity Q/s_c was calculated as

$$Q/s_c = (Q/s_w)/(1-CQ^2/100) \tag{24}$$

where CQ^2 is the well loss in percent of total drawdown.

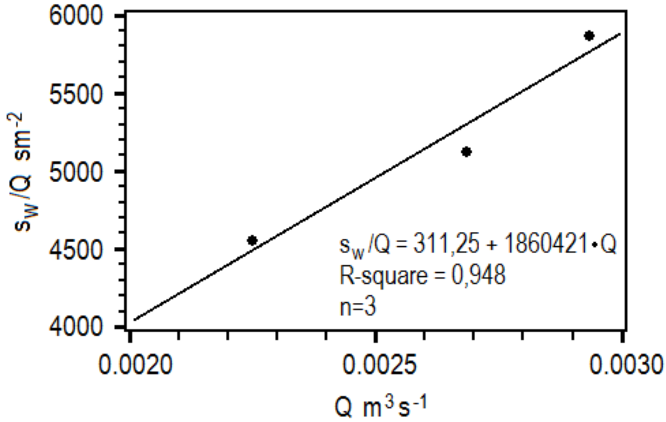


Fig. 32. Determination of well loss in the Jukojärvi test well no. 249037 in the municipality of Keuruu, Central Finland. Q = 10,000 Lhr^{-1} => $S_w = 0{,}86{*}Q + 14{,}36{*}Q^2$ => well loss = 94,3%.

Rorabaugh (1953) has proposed a more generalized form of Jacob´s equation

$$s_w = BQ + CQ^n \tag{25}$$

where the exponent n can range from 1,5 to 3,5.

Atkinson et al. (1994) give a review of the debate over the various methods to analyze step-drawdown tests. They argue that the exponent n in fractured rock aquifers should be in the range of 2 to 3. Labadie and Helweg (1975), Helweg (1994), Karami and Younger (2002) and Avci et al. (2010) have introduced improvements for existing methods when analyzing well losses with step-drawdown tests.

To define an empirical relationship between short-term specific capacity (Q/s_m) and normalized yield (Q/d_s), the regression between specific capacities and normalized yields was determined for the 64 step-test wells. The log-normal character of specific capacity and normalized yield suggested that a linear regression model was more appropriate to the logs of both variable values rather than the original values (e.g. Jetel & Krásný 1968, Carlsson & Carlstedt 1977, Banks 1998). Log-transformed values of each parameter were plotted against each other and a line was fit through the data by using least squares. The yield of a dry well was entered into calculations as a nominal figure of 1 Lhr^{-1}; the corresponding specific capacity was defined as 0,1 $Lhr^{-1}m^{-1}$. Based on the normalized yields (Q/d_s) defined earlier (Chapter 6.2.1) the regression equation was used to determine the specific capacities (Q/s) of the wells in the CF database. To the author´s knowledge, this technique has not been applied elsewhere.

The well productivity (Q_w ms^{-1}) was computed by dividing the specific capacity of a well (Q/s m^2s^{-1}) by the saturated thickness of open well section (SAT m; Chapter 6.2.1; Lattman & Parizek 1964, Siddiqui & Parizek 1971, 1974, Walton 1977, Brook 1988, Singhal & Gupta 1999), i.e.

$$Q_w = (Q/s)/SAT \tag{26}$$

Well productivity has been widely used as an approximate for bulk hydraulic conductivity. Follin et al. (1999) defined an average K-value for Swedish drilled wells as normalized yield per drilled rock depth divided by drilled rock depth (see also Carlsson & Carlstedt 1977, Carlsson & Olsson 1977a, Singhal 1977, Antal et al. 1998, Berggren 1998, Henriksen 2006b, 2008, Rotzoll & El-Kadi 2008), i.e.

$$K_{avg} = Q/d^2 \tag{27}$$

Mabee (1999) has questioned the use of well productivity as a variable for hydraulic properties of bedrock aquifers for the reason that it will probably exacerbate the differences between high and low yielding wells.

6.2.3 Transmissivity

Transmissivity (T_m m^2s^{-1}) was determined with the aid of long-term pumping tests conducted in several drilled wells by the CETECF since the mid-1980's. These wells had first been step-tested. Although the long-term pumping tests have been carried out carefully, the quality of the data for transmissivity determination greatly varies in different wells. This can partly be explained by the conception of the time: the primary intention of the pumping tests has been to investigate the sustainable yield and water quality of the wells, not necessarily to determine their hydraulic properties.

Mostly due to the great variations in the pumping rates during the first hours of the tests, only ten wells in all were included in transmissivity determinations. All the wells were aimed at common water supply. Thus they were of greater than average productivity. Long-term pumping tests are often only conducted on medium- to high-yielding boreholes and, consequently, median specific capacities and transmissivities solely from these data can be very misleading (Sander 1999, Nastev et al. 2004, Graham et al. 2009).

The wells included in transmissivity determinations were pumped at a constant rate with electrical submersible pumps. Despite the fact that the optimum pumping discharges for the long-term pumping tests were predetermined with the aid of the short-term step tests, it was necessary to slightly regulate the discharge rate in some wells at the beginning of the pumping. The drawdown within the pumped wells and the pumping discharge rate were measured with similar equipments as during the step tests.

The water table in the borehole was measured before the start of each pumping test (rest water-table), several times at the beginning of the test, at least once a day during the pumping, and during the recovery stage after the pumping was ceased. The recovery measurements were taken until the water level reached the original static water level or came close to that. In most cases no observation wells in bedrock were available. The duration of pumping tests varied from less than a week to three months depending on the intention of the study (Fig. 33).

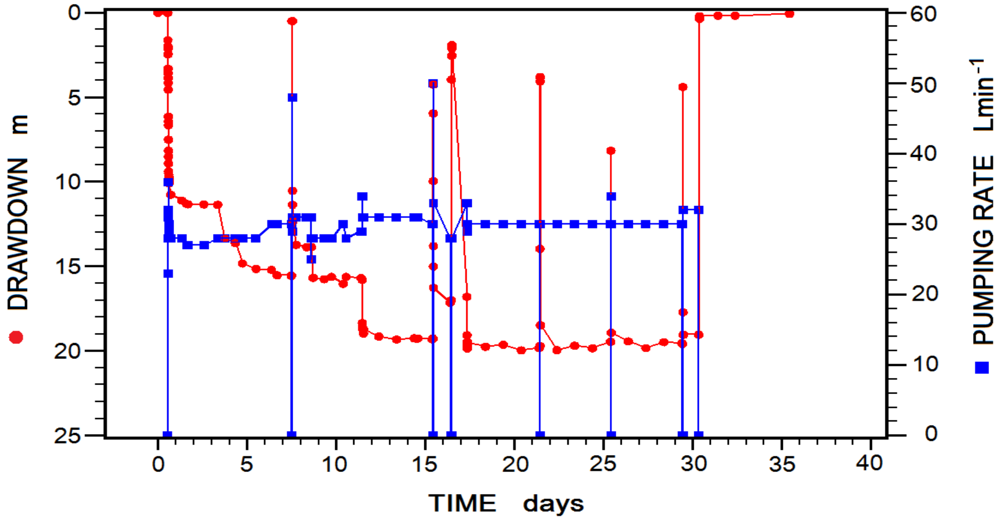


Fig. 33. The progress of pumping test in the Heiska test well no. 850036 in the municipality of Toivakka, Central Finland, in the fall of 2008. This well was not included in transmissivity determinations because of its great pumping rate variation during the first hours of the test.

The discharge-drawdown data were analyzed using the classical Cooper-Jacob (1946) single-well straight-line method for unsteady-state flow (e.g. Kruseman & de Ridder 2000). In this method, observed values of drawdown s_w versus the corresponding time log-t are plotted on a semi-log paper and a straight line is drawn through the plotted points (Fig. 34). The drawdown difference Δs_w per log cycle of time is determined and the values of pumping discharge Q (m^3d^{-1}) and Δs_w (m) are substituted into the equation

$$T_m = 2{,}30Q/4\pi\Delta s_w \qquad (28)$$

Single-well aquifer tests frequently are analyzed with the Cooper-Jacob (1946) method because of its simplicity. Another reason to use the method in this study was that it was considered to best equate the short-term step tests used for determining specific capacities of the wells. In addition, it is a well-known fact that the early part of the drawdown slope gives actual values of T (Driscoll 1987, Raj 1999). Moreover, Uhl and Sharma (1978) and Halford et al. (2006) have suggested that estimating hydraulic properties from single-well pumping tests with anything other than the Cooper-Jacob method is a waste of time.

Although the Cooper-Jacob method was originally derived from Theis (1935) equation for homogeneous isotropic porous media with one or more piezometers, Todd (1967) states that rock aquifers with secondary openings exhibit homogeneous characteristics when sufficiently large volumes are considered. On this basis it has even been observed that pumping tests of wells in rock aquifers yield excellent results (Todd 1967). Other assumptions and conditions for the use of this method are the following: the aquifer is confined and of uniform thickness and has a seemingly infinite areal extent, the aquifer is pumped at a constant discharge rate, the piezometric surface is horizontal, and the well penetrates the entire thickness of the aquifer and thus receives water by horizontal flow (e.g. Kruseman & de

Ridder 2000). The method is also applicable to unconfined aquifers for late-time drawdown data and as long as the water table drawdown is small compared to the original saturated thickness of the aquifer (Verweiji & Barker 1999, Kruseman & de Ridder 2000).

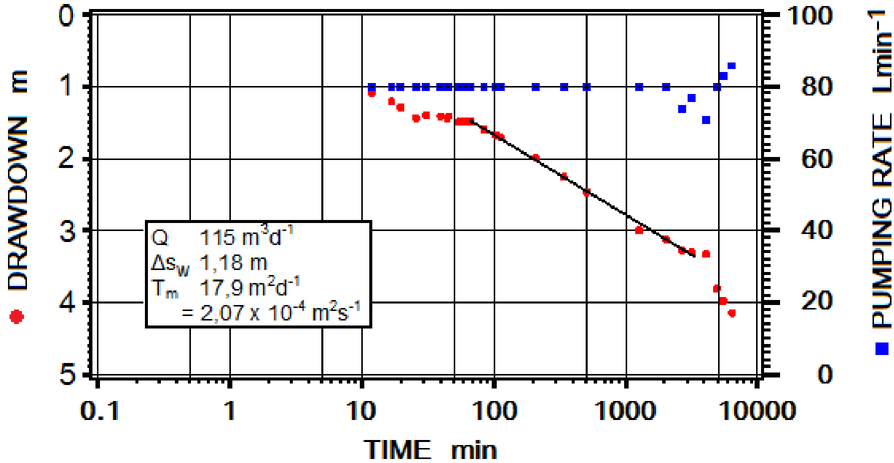


Fig. 34. Transmissivity (T_m) determination in the Oittila test well no. 179330 in the municipality of Jyväskylä, Central Finland, in the fall of 1998 with the Cooper-Jacob single-well straight-line method for unsteady-state flow. The slower drawdown during the first 70 minutes is due to the partial recovery of the well after higher discharge rate in the very beginning of the pumping test (< 10 min, not shown). At the end of the graph (time 5,000 min) the drawdown increases due to increased pumping rate.

Simulations of single-well aquifer tests made by Tumlinson et al. (2006) suggest that early-time drawdown data reflect rapidly changing, volumetric weighted mean T values proximal to the pumping well while late-time drawdown data reflect stabilized conditions and spatially averaged, volumetric weighted mean T out to a considerable distance from the pumping well. The measured drawdowns of the pumping tests might also show effects of delayed yield, the presence of double-porosity media or that of leaky surficial aquifers. In addition, the pumping drawdown curves of some boreholes may show effects of a recharge or impermeable boundaries (Raj 1999, Nastev et al. 2004).

Well losses do not influence the transmissivity analysis using the Cooper-Jacob method because the transmissivity is calculated from drawdown differences. However, the method can be applied only when the influence of wellbore storage has become negligible. To determine whether the early-time drawdown data were dominated by well-bore storage, a log-log plot of drawdown s_w versus pumping time t was made. If the early-time drawdowns plotted as a unit-slope straight line, it could be concluded that well-bore storage effects exist (e.g. Kruseman & de Ridder 2000). However, a longer period of pumping eliminates any effects of borehole storage (e.g. Acheampong & Hess 1999). Papadopulos and Cooper (1967) have observed that the influence of well-bore storage on the drawdown in a well decreases with time and becomes negligible at

$$t>25r_c^2/T \qquad (29)$$

where r_c is the radius (m) of the unscreened part of the well, where the water level is going down. This additional assumption for single well tests had to be satisfied when using the Cooper-Jacob method in this study.

An empirical relationship between T and Q/s was established by regressing the log-transformed transmissivity (T_m) values against well-loss corrected and log-transformed specific capacity values (Q/s_c) derived from the same wells. A line was fit through the plotted data by using least squares. Based on the specific capacities (Q/s) defined earlier (Chapter 6.2.2) the regression equation was used to determine the transmissivities (T) of the wells in the CF database.

In order to find out whether various empirical equations derived in other hard rock areas would suit for the estimation of T in Central Finland, the specific capacities (Q/s) of the CF database wells were entered as input data into these equations (Table 7).

The differences between various simplifications and transformations of T and Q/s listed in Table 7 are probably due to different drawdown, time of observation, well loss correction, diameter, casing depth, heterogeneity, lithology, weathering state of aquifers and due to the partial penetration ratio of the wells (e.g. Clark 1977, Meier et al. 2001, Razack & Lasm 2006, Rotzoll & El-Kadi 2008). However, as Jetel (1968) and Jetel and Krásný (1968) have admitted, such factors are relatively unimportant compared with the variation in aquifer transmissivity over several orders of magnitude. According to Uhl and Sharma (1978), determining transmissivity from specific capacity data is a bit risky and a poor substitute for constant rate pumping tests.

Table 7. Empirical equations derived in different hard rock areas to estimate transmissivity (T) from specific capacity (Q/s m^2d^{-1}). Aq=aquifer type: C=fractured crystalline aquifer, V=volcanic aquifer, L=limestone/ carbonate/karstic aquifer, S=sedimentary aquifer. WL=Q/s corrected (Yes)/uncorrected (No) for well loss.

Reference	Site	Aq	Equation m^2d^{-1}	WL
Eagon & Johe 1972	Ohio, USA	L	$T = 3,24(Q/s)^{0,81}$	Yes
Krásný 1975	Bohemian Massif, The Czech Republic	C	$T = 1,10(Q/s)$	No
Carlsson & Carlstedt 1977	Sweden	C	$T = 1,20(Q/s)$	No
Ericsson & Ronge 1986	Gideå area, Sweden	C	$T = 1,40\ (Q/s)$	No
Houston & Lewis 1988	Victoria Province, Zimbabwe	C	$T=0,507(Q/s)+0,424$	No
Banks 1991, 1992b	different sites	C	$T = 1,11(Q/s)$	No
Huntley et al. 1992	Peninsular Ranges, San Diego, USA	C	$T = 0,12(Q/s)^{1,18}$	No
El-Naqa 1994	Mujib Basin, Jordan	L	$T = 1,81(Q/s)^{0,92}$	No
Sayed & Al-Ruwaih 1995	Dammam aquifer, Kuwait	L	$T = 0,47(Q/s)^{1,13}$	No
Fabbri 1997	Veneto region, Italy	L	$T = 0,85(Q/s)^{1,07}$	No
Mace 1997	Edward aquifer, Texas, USA	L	$T = 0,76(Q/s)^{1,08}$	No
Rhén et al. 1997	different sites in Sweden	C	$T = 2,81(Q/s)^{0,98}$	No
Choi 1999	Jeju Island, Korea	V	$T = 0,45(Q/s)^{1,05}$	No
Mace et al. 1999	Carrizo-Wilcox aquifer, Texas, USA	S	$T = 1,36(Q/s)^{0,84}$	No
Wladis & Gustafson 1999	Fjällveden, Gideå, Sweden	C	$T = 2,22(Q/s)$	No
Jalludin & Razack 2004	The Republic of Djibouti	V	$T = 3,64(Q/s)^{0,94}$	Yes
Hamm et al. 2005	Jeju Island, Korea	V	$T = 0,99(Q/s)^{0,89}$	No
Patriarche et al. 2005	Carrizo aquifer, Texas, USA	S	$T = 2,09(Q/s)^{0,93}$	No
Razack & Lasm 2006	Man-Danane, Western Ivory Coast	C	$T = 0,33(Q/s)^{1,30}$	Yes
Rotzoll & El-Kadi 2008	Hawaii, USA	V	$T = 1,54(Q/s)^{1,002}$	No
Verbovšek 2008	Slovenia	C	$T = 1,08(Q/s)^{1,07}$	No
Yidana et al. 2008	Voltaian Basin, Ghana	S	$T = 0,77(Q/s)^{1,08}$	No
Banks et al. 2010	Fennoscandia	C	$T = 1,43(Q/s)$	No
Richard et al. 2011	Saguenay Basin, Quebec, Canada	C	$T = 0,97(Q/s)^{1,08}$	No

6.2.4 Hydraulic conductivity

The bulk hydraulic conductivity (K ms^{-1}) of a well was calculated with the aid of its transmissivity (T) value determined earlier (Chapter 6.2.3) by simply dividing the T-value (m^2s^{-1}) with the thickness of the aquifer (b m). This study adopted the common approach where the aquifer thickness was defined as the saturated open well section (Chapter 6.2.1), i.e.

$$K = T/SAT \tag{30}$$

That is, hydraulic conductivity was calculated as an average value over the whole length of the saturated open borehole (e.g. Kruseman & de Ridder 2000, Ratej & Brenčič 2005, Boutt et al. 2010, Courtois et al. 2010).

Acheampong and Hess (1999) estimated hydraulic conductivity by dividing the transmissivity (derived from pumping test analysis) by the sum of the thicknesses of the various fracture zones in the well with the knowledge that contributions from the various fractures to the well may vary due to the anisotropic nature of fractures (Paillet & Kapucu 1989). This was not, however, attempted in this study, though there were lithological logs available from test wells.

For roughly estimating groundwater flow velocities in bedrock aquifers Darcy´s law can be applied (e.g. Bowen 1986, Fetter 1994). Darcy´s law can be written as

$$Q = -KA_f\Delta h \tag{31}$$

where Q is the discharge (m^3s^{-1}), K is the hydraulic conductivity (ms^{-1}), A_f is the cross-sectional area of flow (m^2) and Δh is the hydraulic gradient (dimensionless). The negative sign shows that water flows in the direction of decreasing head. Rearrangement of Darcy´s equation yields

$$v = Q/n_eA_f = - K\Delta h/n_e \tag{32}$$

where v is the average linear flow velocity (ms^{-1}) and n_e is the effective porosity (dimensionless).

The effective porosity (n_e), also called the kinematic porosity, of a medium is defined as the volume of interconnected void space contributing to flow per unit bulk volume of rock (e.g. Gordon 1986). The effective porosity of crystalline rocks is generally very small. In Sweden, it has been estimated to lie around 0,0005 (Knutsson & Fagerlind 1977). According to Almén et al. (1986), the effective porosity of intact granite rocks is in the order of 10^{-4} and may be assumed to vary between 10^{-6} and 10^{-3}. Shevenell (1996) gives effective porosities of $1x10^{-4}$, $1x10^{-3}$, and $3x10^{-3}$ for conduits, fractures, and matrix elements, respectively, of the limestone and dolomite aquifer in Tennessee, USA. According to Lloyd (1999c), effective porosity values for hard rocks will rarely exceed 0,01 and are more likely to be of the order of 0,01-0,0001. Gbürek et al. (1999) estimate effective porosities of $5x10^{-3}$ for highly fractured rocks and $1x10^{-4}$ for poorly fractured rocks in Pennsylvania, USA. Healy and Cook (2002) doubt whether the values of n_e of crystalline rocks could be estimated with sufficient accuracy.

It has been suggested that, due to the high variation of hydraulic conductivity (K) values within a single well, it hardly serves any useful purpose to determine this parameter, unless aquifer modeling is being attempted and the model requires K (Shapiro 1993, Raj 1999). In addition, even multiwell pumping tests yield only an average or bulk hydraulic conductivity that is not adequate for predicting rapid ground water travel times through the fracture network (Muldoon & Bradbury 2005). Salmi (1985) has emphasized that hydraulic conductivity values available in different datasets are very heterogeneous owing to the diversity of measuring equipment, length of section measured, borehole diameter and mathematical solutions used.

Although approximate, hydraulic conductivity estimates obtained from specific capacity data have been considered acceptable for regional hydrogeological studies covering large areas (Jetel 1968, Jetel & Krásný 1968, Carlsson & Carlstedt 1977, Carlsson & Olsson 1977a, Antal et al. 1998, Nastev et al. 2004, Henriksen 2006b, 2008). Berggren (1998) has discovered that air-blow tests, recovery tests and test pumpings yield approximately the same median values for specific capacity and hydraulic conductivity independently of the type of capacity determination used.

6.3 Factors affecting well yield and hydraulic properties

6.3.1 General

In this study, factors affecting well yield and hydraulic properties have been divided into five groups: construction factors, geologic factors, topographic factors, lineament factors and catchment factors. The factors in different groups may be called with a common name 'well factors'.

Location of wells (NCOORD, ECOORD) in the CF database was assigned by Finnish national grid coordinate system (KKJ) with a precision of 10 meters. If necessary, the system could be converted into uniform grid coordinates (YKJ).

A grid of a circular cell was constructed around each well, so that the well itself was situated in the center of the circle (Fig. 35). The radius of the circle was $(1\ \text{km}/\pi)^{0,5} \approx 0{,}564$ km. Then the area of the circle was 1 km^2. Each well was associated only with the factors that occurred within the 0,564 km radius of the well´s coordinates. This circle area surrounding a well was defined as the well´s catchment area. Galanos and Rokos (2006) have employed the term well´s neighborhood. Brook et al. (1986) used a circular sample area 0,488 km in radius centered upon each well in their well productivity studies in Georgia, USA. Brook (1988) employed 1,5 km as a corresponding radius of his circular cell grid centered on each well; the area of the circle was about 7 km^2. Tam et al. (2004) used 0,450 km as radius of their cell grid; the cell grid was a square with an area of 0,09 km^2.

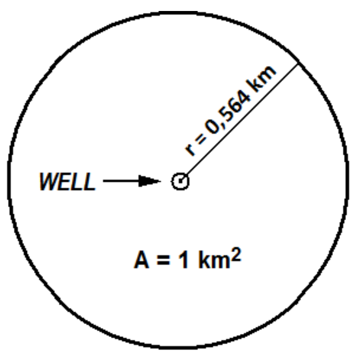


Fig. 35. A grid of a circular cell representing a well´s catchment area. The well is situated in the centre of the circle.

The literature contains little information on capture zone geometry in fracture-controlled systems (e.g. Risser & Barton 1995, Rayne et al. 2001). According to Bair and Roadcap (1992), capture zones computed by different methods center the wells and are circular or subcircular in shape. In real world, the situation is unlikely that simple. Instead, capture zone geometry in fractured systems can be very complex (Podgorney & Ritzi 1997).

Climatic factors were assumed uniform in the study area (see Chapter 5.1). For this reason they were not considered in this study.

6.3.2 Construction factors

Factors related to well construction (e.g. diameter, length of casing, total depth) have been considered important when accounting for differences in well yield before attempting to assess yield to natural factors, for example lithology (e.g. Knopman 1990). However, construction factors have only a limited value in predicting the well production properties in advance, that is until the drilling is completed.

Water-supply boreholes in the CF database have been drilled within the period of 1947 to 2008 by 35 different drilling contractors using cable-tool and, since the mid-1970's, down-the-hole hammer (DTH) methods (Fig. 36). Based on the construction year (CY), the wells were roughly divided into two groups (1947-1974 and 1975-2008) to see if the drilling method had any effect on the yield and hydraulic properties of the wells. According to MacDonald et al. (2002) and Moore et al. (2002a), there may be less clogging of fractures by bedrock cuttings when the well is drilled with the cable-tool rig compared with the DTH method. All wells except one test well (no. 931075) have been drilled vertically into the bedrock. For a detailed account of the various drilling techniques the reader is referred to Cruse (1979), Driscoll (1987), Anon (2003), Rebouças (2004) and Misstear et al. (2006). Banks (1992a) has addressed the optimal orientation of water-supply boreholes.

The elevation of a well site in meters above sea level (ASL m a.s.l.) was determined by interpolation from 1:20 000 topographic maps with 2,5 and/or 5 meter contours. The elevation accuracy of the maps is about one meter (Rahkila & Jokela 1988).

Fig. 36. DTH-drilling going on in the Tereniemi test well no. 931073 in the municipality of Viitasaari, Central Finland, in September of 1988. The final well depth was 43 m and Q=1,800 Lhr^{-1}. Photo the author.

The collapsible soil mantle above the bedrock is in most cases sealed off using a steel or plastic casing (CAS; Fig. 37). Drillers in Central Finland prefer to end the casing in compact bedrock to avoid the upper few meters, which may be weathered or highly fractured. The purpose of this is to keep the borehole open and prevent well contamination by near-surface waters (e.g. Harned & Daniel 1989). The weathered zone-bedrock interface may often be gradational, whereas in hard and massive bedrock the contact between overburden and relatively competent bedrock tend to be abrupt and can be easily detected by the driller from the response of the drilling machine. The boreholes in Central Finland are left uncased and unscreened open boreholes below weathered and fractured surficial zones. This is largely the practice in unweathered crystalline rocks in other parts of the world, too (e.g. Larsson et al. 1984, Loiselle & Evans 1995, Cohen et al. 1996, Moore et al. 2002a). According to Risberg and Thorsbrink (2002), Swedish standard well should be equipped with a six meters long casing in a minimum driven at least two meters into an intact bedrock. The corresponding driving length in Pennsylvania, United States, has been about 1,5 meters (Henry 1992).

The depth to the first (main) water strike (STR) or "principal water vein" (Clapp 1911a) has usually been recorded during drilling (Fig. 37). Larger cutting size and changes in color and texture of cuttings produced from the borehole indicate when the drill bit encounters a fracture or fracture zone (e.g. Henriksen & Braathen 2006). Transmissive fractures can be identified during DTH-drilling as water is being transported to the surface from different depth intervals (e.g. Kulkarni et al. 1997). However, narrow or mineral-filled water-bearing fractures might go unnoticed during DTH-drilling (Mäkelä 1990b). According to Paillet and Duncanson (1994), the depths of the fractures associated with water production during drilling are comparable to the depths of the fractures shown to be the source of water inflow in the geophysical log analysis.

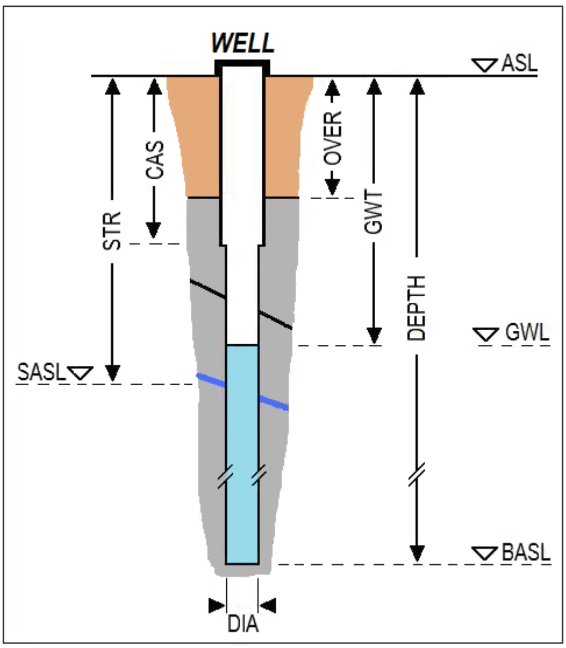


Fig. 37. Determination of construction factors for a drilled well.

The fracturing degree of bedrock at well site (FRAC) has been determined in connection with well drilling based on drilling experiences and observations. The FRAC is expressed in the CF database in a subjective, qualitative manner as (1) low (compact or intact rock), (2) medium (moderately fractured rock), and (3) high (highly fractured and possibly weathered rock).

After drilling to total depth, the DTH-drillers usually continue to airlift with the drill rod fixed at the bottom of the well and discharge water entering the well, until the return water is relatively clear and do not contain appreciable sand or rock fragments. After airlift is stopped the groundwater table in the borehole starts to rise. The static groundwater table (GWL) generally lies near the ground surface. For artesian water wells the depth to the groundwater table (GWT) is marked with a zero in the CF database. In cable-tool wells the bailer has been used to bail out the cuttings and water from the bottom of the well (e.g. Misstear et al. 2006). These well drilling routines are commonly used in other crystalline rock areas, too (e.g. Woolley 1982, Kaehler & Hsieh 1994, Loiselle & Evans 1995, Williams et al. 2004, Henriksen 2006b, Misstear et al. 2006). Afterwards the wells are commonly sampled for water quality determinations.

During the past four decades hydraulic fracturing has been used to develop low production water wells in fractured rock aquifers. Earlier also shooting with an explosive charge (blasting) was used to increase the yields of boreholes (Müllern & Eriksson 1977, Fagerlind 1979, 1982, Waltz & Decker 1981, Kandolin 1989, Mäkelä 1990b, Banks et al. 1992a, 1994, 1996, Less & Andersen 1994, Rohr-Torp 1994b, Lloyd 1999a, Devalpally et al. 2000, Cobbing & Ó Dochartaigh 2007). Hydraulic fracturing (or hydrofracturing) of a borehole involves injecting water under very high pressure into a sealed interval or part of borehole to flush and remove fine particles and rock fragments from existing bedrock fractures and/or increase the size and extent of existing fractures (Clark 1949, Hubbert & Willis 1957, Koenig 1960a, 1960b, 1960c, 1961, Stewart 1974, 1978, Williamson & Woolley 1980, Williamson 1982, Driscoll 1987, Howard et al. 1992, Joshi 1996, Banks & Less 1999, Banks & Robins 2002, Kalskin Ramstad et al. 2003, 2005, Kalskin Ramstad 2004, Misstear et al. 2006). Since 1947 hydraulic fracturing has successfully been used in oil wells to create reservoir fractures that improve well productivity (Howard & Fast 1970). Some 5% of the drilled water wells in the CF database have been developed by hydrofracturing or blasting.

Well construction factors (CAS, GWT, STR, DEPTH) in the CF database were measured from the ground surface (Fig. 37) and were rounded to the nearest meter. The height of well construction factors in meters above sea level could be calculated by subtracting them from the height of the well site (ASL m a.s.l.). For example, the level of the groundwater table (GWL m a.s.l.) was calculated as

$$GWL = ASL - GWT \tag{33}$$

The bottom level of a borehole (BASL m a.s.l.) was defined as

$$BASL = ASL - DEPTH \tag{34}$$

and the level of the first (main) water strike (SASL m a.s.l.) was calculated as

$$SASL = ASL - STR \tag{35}$$

6.3.3 Geologic factors

Soil type, soil thickness and extent, and the condition of the soil-bedrock interface may provide important impacts on well yield from crystalline rocks (Ellis 1906, Rand 1978, Olsson 1979, 1980, Omorinbola 1982, 1984, Pokki 1983, Acworth 1987, Olorunfemi et al. 1991, Olofsson 1993, Kastrinos & Wilkinson 1994, Henriksen 1995, Raj 1999, Olofsson et al. 2001, Banks & Robins 2002, Moore et al. 2002a, Caine & Tomusiak 2003).

Though the hydraulic conductivity in a saturated overburden is usually low, the total vertical leakage can equal the drilled well discharge during pumping when the cone of depression spreads out (Driscoll 1987) and can thus bring equilibrium in the pumping wells. Coarse and well-sorted soils, such as sand and gravel, which are directly deposited on the bedrock surface, favor flow into the bedrock (Kastrinos & Wilkinson 1994, Olofsson 1994, Cesano et al. 2000, Mabee et al. 2002). However, infiltration into the rock from such sediments is primarily a function of the hydraulic properties of the rock structures (Olofsson 1993).

In this study the surficial deposits at well sites (SOIL) were classified in five groups: (1) no overburden (outcrop of bedrock), (2) till, (3) coarse-grained sediment (glaciofluvial sand and gravel), (4) fine-grained sediment (silt, clay) and (5) peat. The soil type was interpreted from geologic and topographic maps (e.g. Vähäsarja 1971, Haavisto 1983). The thickness of the soil cover at well site (OVER m) was confirmed in connection with well drilling.

Several authors have pointed out the significance of lithology for the well yield and hydraulic properties (e.g. Larsson 1987, Rohr-Torp 1987, 2003, Knopman 1990, Henriksen 1995, Morland 1997, Olofsson et a. 2001). Some authors for their part keep the lithology of crystalline igneous and metamorphic rocks more or less non-significant from the well production point of view, at least on a regional scale (e.g. Wladis 1995, Havlík & Krásný 1998, Krásný 2002). In the present study, the bedrock geological information of the database wells was obtained by combining the well database with the digital bedrock map of Central Finland compiled at a scale of 1:200 000 (GTK 2007). This map relies mainly on the 1:400 000 bedrock geological map of Central Finland Granitoid Complex (Nironen et al. 2002a). It contains 22 mapped bedrock units (rock types) with post-kinematic intrusions coded with letter P (Fig. 38). The map made it possible to categorize each borehole with respect to bedrock type. Rock types and their grouping are shown in Table 8. The nomenclature of the magmatic rocks has been harmonized according to the IUGS recommendation (Streckeisen 1973, Nironen et al. 2002a, 2002b, Nironen 2003).

In order to find out if there are any relationships between the highest shore level of the Baltic Sea (HSL) and the well yield and hydraulic properties, the observations of the HSL made by Ristaniemi (1987; his Appendix I) were plotted on a map (scale 1:600 000) and the isolines were determined by manual interpolation and digitized (Fig. 27:A). The shore level values were then designated for the wells.

In Fennoscandia, the land uplift during the Holocene, caused by the unloading of the Late Weichselian ice sheet, has been considered an important regional factor for well yield (Rohr-Torp 1994a, 1995, 2000, Morland 1997, Gudmundsson 1999, Gudmundsson et al. 2002). It has been proposed that the result of the isostatic response would be significantly increased groundwater production in the areas of greatest uplift, but less marked yields of wells in the marginal areas. Thus, other things being equal, the total postglacial uplift should be a direct measure of the potential groundwater production from the crystalline bedrock (Rohr-Torp 1994a, 1995, 2000, Morland 1997, Gudmundsson 1999, Gudmundsson et al. 2002).

For the purpose of this study, the wells in the CF database were allocated a representative value according to the apparent annual land uplift (UPLIFT) isolines, which were digitized from the map of Mäkinen et al. (2003).

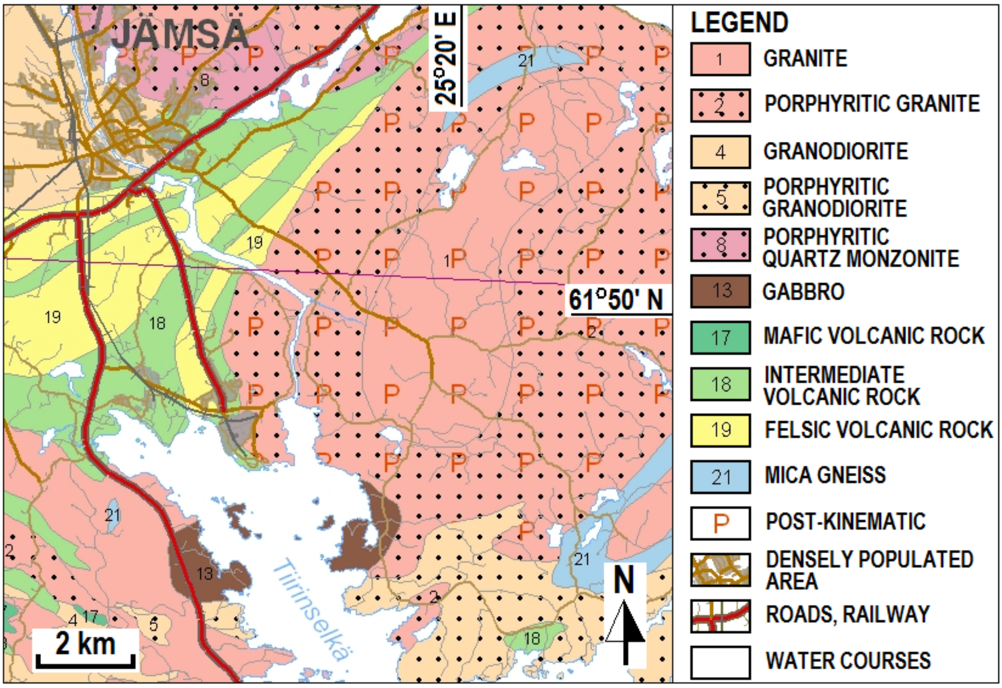


Fig. 38. Rock types in the Jämsä region, southern Central Finland. Modified from the bedrock map of Central Finland (GTK 2007).

Table 8. Rock types and their grouping in the CF database (see Nironen et al. 2002a, GTK 2007).

Grouping	Subgrouping	Code number	Rock type	Abbre-viation
Intrusive rocks	Felsic intrusive rocks	1	Granite	Gr
		2	Porphyritic granite	PGr
		3	Pyroxene-bearing granite	PxGr
		4	Granodiorite	GrDr
		5	Porphyritic granodiorite	PGrDr
	Intermediate intrusive rocks	6	Tonalite	Ton
		7	Quartz monzonite	QMZ
		8	Porphyritic quartz monzonite	PQMZ
		9	Quartz monzodiorite	QMDr
		10	Quartz diorite	QDr
		11	Monzodiorite	MDr
		12	Diorite	Dr
	Mafic intrusive rocks	13	Gabbro	Gb
		14	Ultramafic intrusive rock	UM
Subvolcanic rocks		15	Intermediate subvolcanic rock	ISv
		16	Felsic subvolcanic rock	FSv
Supracrustal rocks	Volcanic rocks	17	Mafic volcanic rock	MV
		18	Intermediate volcanic rock	IV
		19	Felsic volcanic rock	FV
		20	Quartz-feldspar schist and gneiss	QF
	Metasediments	21	Mica gneiss	MG
		22	Mica schist	MS

6.3.4 Topographic factors

Topography has been identified as an important indication of well yield, because it may reflect internal structures in the bedrock. Low altitude areas between ridges and hilltops have often been related to bedrock faults or fracture zones, and hence, to higher well yields. Areas of ridges and hilltops have expected to potentially be less fractured as they have been more resistant to erosion and, therefore, would have the potential of being areas where the water-yielding characteristics of wells would be lower. Indeed, many authors have noticed that wells on flat uplands or in valleys tend to yield larger amounts of water than wells on valley sides or sharp hilltops (Ellis 1906, LeGrand 1954, 1967, 1979, 1992, 2004, Davis & Turk 1964, Davis & DeWiest 1966, Poth 1968, Siddiqui & Parizek 1971, Larsson 1977, Uhl et al. 1979, Wyrick & Borchers 1981, Dijon 1984, Larsson et al. 1984, Daniel 1987, 1989, White et al. 1988, Knopman 1990, Lewis 1990, Huntley et al. 1991, Zewe & Rauch 1991, Barker et al. 1992, Henry 1992, McFarlane et al. 1992, Helvey & Rauch 1993, Henriksen & Kyrkjeeide 1993, Kastrinos & Wilkinson 1994, Henriksen 1995, 2006b, Braathen et al. 1999, Mabee 1999, Singhal & Gupta 1999, Krásný 2000, 2005, Moore et al. 2002a, Robinson 2002, Eftimi 2003, Kenny et al. 2006, Verbovšek & Veselič 2008, Holland & Witthüser 2011).

For the purpose of this study, four topographic settings (hilltop, slope, flatland, valley) were discriminated based on the location of the boreholes on 1:20 000 topographic maps. Several authors, for example LeGrand (1954, 1967, 1992), Daniel (1987, 1989) and Robinson (2002), have earlier used similar classifications. The differences between various classification practices have most often been due to different geologic and topographic environments.

Each well´s topographic setting (TOPO) was determined in the following way. Hilltop wells were located within two 5-meter contours of a hilltop, i.e. no more than 10 meters downwards from the hilltop (Fig. 39). This can be compared with Robinson´s (2002) hilltop wells, which were located within two 20-foot contours of a hilltop, i.e. about 12 meters downwards from the hilltop. The hilltops of this study must rise 25 meters above the surroundings, at least. Robinson´s (2002) slope wells were located between hilltop and valley positions with a slope equal or greater than 10 percent. In this study the slope was equal or greater than 2,5%, i.e. ≥25 meters per 1 km. The slope of Robinson´s (2002) upland wells was less than 10 percent. In this study the corresponding flatland slope was <2,5%, i.e. <25 meters per 1 km. Robinson´s (2002) valley wells were located within two 20-foot contours of a stream, valley bottom, or dry valley. In this study, the valley wells were actually lineament wells located no more than 150 meters away from the nearest lineament centre (e.g. Krishnamurthy et al. 2000, Lachassagne et al. 2001, Holland & Witthüser 2011). Valleys are sometimes wide so that wells drilled in valley floors may not encounter the narrow zone of structural weakness along which the valley originally developed (Brook 1988). In this study wide valleys without any distinct fault zones were interpreted as flatlands.

The height of the highest (ASLH) and lowest (ASLL) locations (m a.s.l.) in a well´s catchment area were determined by interpolation on 1:20 000 topographic maps with 2,5 and/or 5 meter contours. The relative height differences (m) in a well´s catchment area were defined with the following combination variables (Fig. 39):

$$\mathrm{RELA} = \mathrm{ASLH}-\mathrm{ASLL} \tag{36}$$

$$\mathrm{RELAH} = \mathrm{ASLH}-\mathrm{ASL} \tag{37}$$

$$\mathrm{RELAL} = \mathrm{ASL}-\mathrm{ASLL} \tag{38}$$

The relation between the variables RELAH and RELAL was defined as

$$RERE = RELAH/RELAL \qquad (39)$$

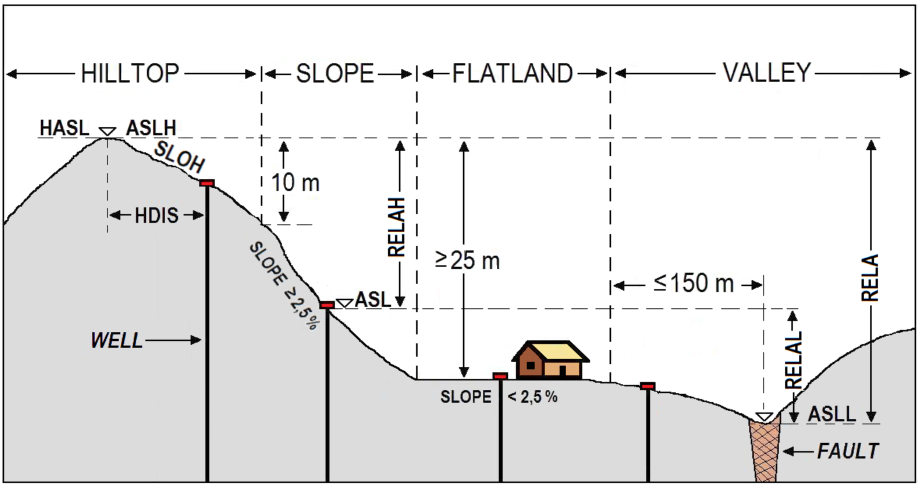


Fig. 39. Determination of topographic factors for drilled wells.

Distance from a well to the nearest (bedrock) hilltop (HDIS m) was measured on 1:20 000 topographic maps with a ruler. Then no fault zones were allowed between the well and the hilltop. The height of the hilltop (HASL m a.s.l.) was determined by interpolation from topographic maps with 2,5 and/or 5 meter contours. The height difference (m) between the peak of the hilltop and the well site was defined as

$$HAASL = HASL\text{-}ASL \qquad (40)$$

The rate of change in elevation (SLOH °) between the well site and the hilltop was calculated as

$$SLOH = \tan^{-1} (HAASL/HDIS) \qquad (41)$$

It has been assumed that steep slopes result in higher runoff rates and lower groundwater-recharge rates and, hence, lower well yields than gentle slopes (e.g. Clarke & McFadden 1991).

6.3.5 Lineament factors

Lineament factors, such as fracture densities and distance to lineaments and lineament intersections, have been applied with variable success to explain the variability of the yield and hydraulic properties of drilled wells. The basic idea of lineament analysis have been that if wells with high yields can be related to the location and properties of lineaments, then the lineament information could be useful in locating zones of high productivity in the aquifer (e.g. Larsson 1972, De La Garza & Slade 1986, Astier & Paterson 1989, Banks et al. 1992b, Yin & Brook 1992a, Briz-Kishore 1993, Gustafsson 1994, Mabee et al. 1994, 2002, Sidle & Lee 1995, Sami 1996, Magowe & Carr 1999, Sander et al. 1999, Kim et al. 2004, Tam et al. 2004).

The horizontal distance from a well to the nearest lineament (LDIS m) and that to the nearest lineament intersection (LIND m) were measured with a ruler on topographic maps at a scale of 1:20 000 with contours at 2,5 and/or 5 meter intervals (Fig. 40; e.g. Schowengerdt et al. 1981, Brook 1988). Each well was associated only with the nearest lineament and the lineament intersection that occurred within a 0,564 km radius of the well´s coordinates. The lineament centre was interpreted to locate where erosion (e.g. glacial abrasion, downcutting by a stream channel) had been the greatest (e.g. Henry 1992). The LDIS and LIND are measures of the proximity of the well to possible localized zones of higher aquifer permeability. Hence, wells on or close to such zones should have higher productivities (e.g. Setzer 1966, Brook 1988, Astier & Paterson 1989, Clarke & McFadden 1991, Yin & Brook 1992a, Wladis 1995, Banks et al. 1996, Braathen & Gabrielsen 1998, Braathen et al. 1999, Magowe & Carr 1999, Elfouly 2000, Nativ et al. 2003).

The nearest lineament was classified into one of the three groups (LPRO) based on its total (non-segmented) length on the 1:200 000 lineament map (Fig. 23): (1) > 10 km, (2) 2-10 km, and (3) <2 km. The actual length of a lineament may be difficult to find out. Indeed, shorter fault lengths may occasionally be a consequence of relatively poorly exposed faults in areas with considerable glacial deposits (e.g. Munier & Fenton 2004). On the other hand, the length is the only available measure to show the magnitude of a lineament (e.g. Tsuchida et al. 1990). Interpretation of the dip or width of lineaments was not attempted in this study.

Lineament density is usually represented as line density, length density or intersection density of lineaments (Vuorela 1982, Tossavainen 1992, Hung et al. 2002, Jia & Lin 2010). In this study, lineament-frequency density (LFRE) was defined as the number of all lineaments per a well´s catchment area of 1 km^2 (see Brook 1988, Edet et al. 1998, Elfouly 2000, Fernandes 2003, Hung et al. 2003, Tam et al. 2004). The cumulative length of all lineaments (LLEN m) within a well´s catchment area was measured on maps. Average lineament length (LLLF m) in a well´s catchment area was defined as

$$\mathrm{LLLF} = \mathrm{LLEN/LFRE} \qquad (42)$$

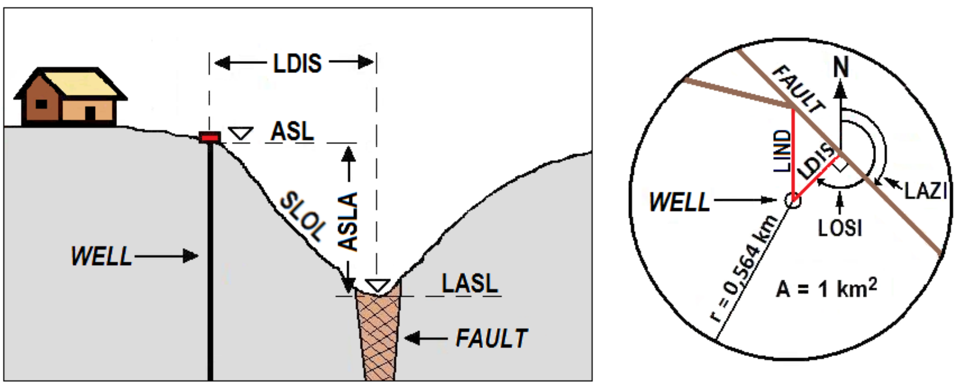


Fig. 40. Determination of lineament factors for a drilled well.

Lineament-length density (LDEN km^{-1}) is equal to the total length of all lineaments (km) per a well´s catchment area 1 km^2 (Tossavainen 1992, Edet 1996, Edet & Okereke 1997, Edet et al. 1998, Fernandes 2003, Tam et al. 2004). Lineament-intersection frequency (LINF) was defined as the number of intersections of lineaments per a well´s catchment area (see Brook 1988, Hung et al. 2002, 2003, Fernandes 2003). These lineament factors were defined only for wells with a nearest lineament extraction.

According to Schowengerdt et al. (1981), specific capacity and transmissivity values should be large where cumulative lineament length values are large, and vice versa. According to Brook (1988), the number and length of lineaments are measures of the average secondary permeability of the bedrock and therefore of the magnitude of groundwater flow to be expected in the region around the well. As these variables increase, well productivities might also be expected to increase. Hardcastle (1995) suggests that the number of photolineaments, the number of directional lineament families and the total length of lineaments within an area of defined radius can be used to rank the area´s potential to store and transmit large volumes of groundwater. According to Tam et al. (2004), zones of high lineament-length density can be used as an indication for zones of high concentrations of fractures or faults. It must also be admitted that in crystalline basement areas high lineament-length density usually corresponds to areas of outcropping bedrock and thin soil mantle, whereas low lineament-length density is indicative of buried bedrock and thick overburden (e.g. Edet et al. 1998).

The height of the bottom of the nearest lineament (LASL m a.s.l.) was determined by interpolation from 1:20 000 topographic maps with 2,5 and/or 5 meter contours. The height difference in meters between the well site (ASL) and the lineament bottom (LASL; Fig. 40) was calculated as

$$ASLA = ASL - LASL \tag{43}$$

The rate of change in elevation (SLOL °) between ASL and LASL was calculated as

$$SLOL = \tan^{-1} (ASLA/LDIS) \tag{44}$$

LWAT (1/0) informed whether there was a water course or not in the bottom of the nearest lineament.

The location of a well with regard to the nearest hilltop vs. the nearest lineament was described with the combination variable

$$HLDIS = HDIS/LDIS \tag{45}$$

The higher the HLDIS, the longer the distance from a well to the nearest hilltop and/or the smaller the distance to the nearest lineament and vice versa. Other combination factors were defined as

$$HALA = HASL/LASL \tag{46}$$

and

$$HLASL = HASL - LASL \tag{47}$$

It is important to know whether any preferred lineament directions exist that will provide high well yields or, at least, a lower risk of borehole failure than others (e.g. Greenbaum 1992). To evaluate this, the orientation of the lineaments (LAZI º) was examined from 1:20 000 topographic maps by measuring the strike of the nearest lineament of each well within a 0,564 km radius of the well´s coordinates with a protractor (Fig. 40). The strikes of the lineaments were then plotted with well yield and hydraulic parameters on azimuth plots.

The position of each well with respect to the nearest lineament (LOSI °) was defined as the azimuth of a straight line (normal) drawn perpendicular from a well to the nearest lineament (Fig. 40). With the parameter LOSI it was possible to express with a single variable in which side of the lineament of a certain orientation the well was situated. Both LAZI and LOSI were given in degrees clockwise from north. In statistical programs and rose diagrams the LOSI was replaced with the variable LPOS, which was defined as follows: if $0° < LOSI \leq 270°$ then $LPOS = LOSI + 90°$; else if $270° < LOSI \leq 360°$ then $LPOS = LOSI - 270°$.

Gustafsson (1994) suggests that the lack of any favored lineament directions in an area, when correlating borehole data with lineaments, would support the hypothesis of the near-surface bedrock zone dominated by tensile stresses as a result of gravitational unloading and lateral expansion during uplift and erosion. According to Sander (2007), statistically significant borehole populations with reliable information are very seldom available to give conclusive results regarding certain lineament azimuths as more promising than others.

6.3.6 Catchment factors

The horizontal distance from each well to the shoreline of the nearest water body (WDIS m) was measured on topographic maps at a scale of 1:20 000 with contours at 2,5 and/or 5 meter intervals with a ruler (Fig. 41). Only shorelines of water bodies with an area of ≥1,0 ha and those of perennial streams were used in this analysis. Each well was associated only with the nearest water course that occurred in a well´s catchment area within a 0,564 km radius of the well´s coordinates. The height of the water level of the nearest water body (WASL m a.s.l.) was in most cases marked on topographic maps. Otherwise it was interpolated from maps.

The height difference in meters between the well site (ASL m a.s.l.) and the nearest water course (WASL; Fig. 41) was calculated as

$$ASLW = ASL - WASL \tag{48}$$

The corresponding relation of these variables was defined as

$$ASWA = ASL/WASL \tag{49}$$

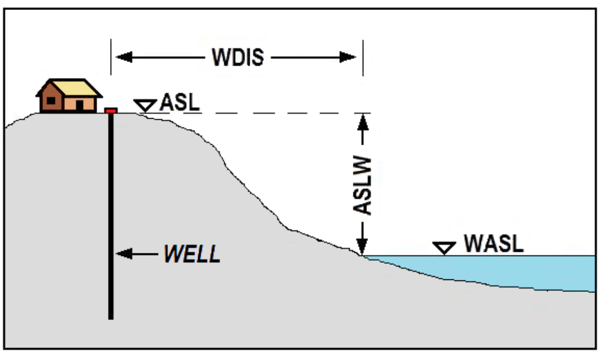


Fig. 41. Determination of catchment factors for a drilled well.

The effect of land use in a well´s catchment area on well yield and hydraulic properties was examined through the variables BED, FOR, PEA, CUL, SET and WAT (Table 3, Fig. 42). Initially these variables were aimed at investigating groundwater quality issues. However, they were thought to possibly have some (implicit) influence on the well production properties, too. The variables were interpreted and measured on 1:20 000 topographic maps with a well site -weighted sampling method consisting of 25 points (Fig. 42:A). The weighting method was chosen because the land use in the vicinity of a well was interpreted to have a more significant influence on the well than that of the more distant catchment area. The measuring results were multiplied by four to transform them as percentages (or hectares) of a well´s catchment area (1 km^2).

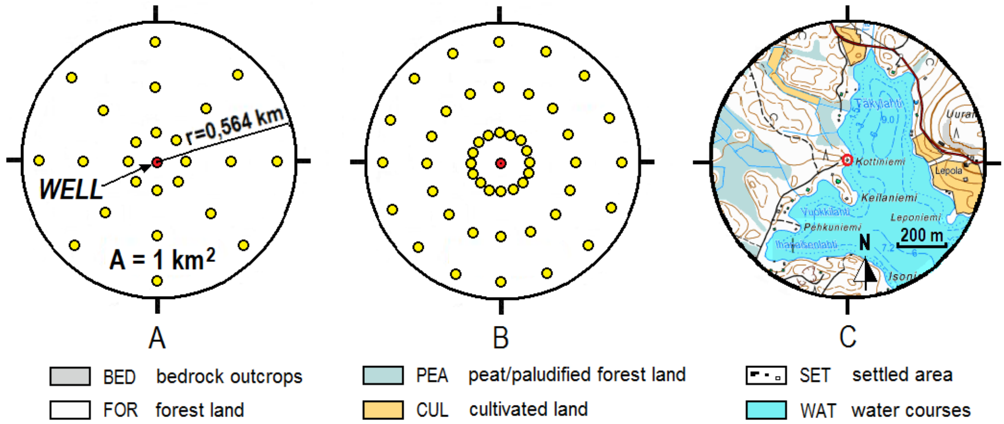


Fig. 42. The well site -weighted sampling devices consisting of A) 25 and B) 50 points for land use determination in a well´s catchment area (C). The well is placed in the center of the circle. Base map © National Land Survey of Finland permission no. 7/MML/11.

The author had tested the adequacy of 25 sampling points in 1993 with the non-parametric paired sample Wilcoxon sign rank test for 50 randomly sampled wells (Chapter 6.1.4). The measuring results were compared to those of a well site -weighted sampling method with 50 points multiplied by two (Fig. 42:B; the well site was measured twice). There were no statistically significant differences between the measuring results of the two methods. Thus it was concluded that land use could be determined accurately enough with 25 sampling points. The measuring results are attached in Appendix III. The test results are shown in Table 9.

Table 9. Results from testing the adequacy of 25 sampling points for the land use of wells´ catchment areas. The explanations of different abbreviations are given in Table 3.

Variable	n	Shapiro–Wilk test	p	Wilcoxon sign rank test	p
BED	50	0,8700	<0,0001	6,5	0,8516
FOR	50	0,9432	0,0286	17,5	0,8343
PEA	50	0,8883	<0,0001	10	0,7992
CUL	50	0,9628	0,2029	2,5	0,9752
SET	50	0,9091	0,0007	-30	0,5707
WAT	50	0,8457	<0,0001	-7	0,8422

6.4 Statistical methods

The data were analyzed and processed with the statistical analysis system of the SAS Institute Inc. that is available in the computer of the Finnish Environment Institute (FEI). SAS provides a wide range of statistical and graphical techniques and processes for the classification, analysis and interpretation of data (SAS 1990a, 1990b, 1990c, 1990d, 1990e).

For a detailed account of statistics with the SAS system the reader is referred to Cody and Smith (1987), Lewis and Ford (1987), SAS (1990a, 1990b, 1990c, 1990d, 1990e), Laine (1996), Korhonen (1998), and Khattree and Naik (1999). The basics of statistics are well described, for example, in Conover (1980), Draper and Smith (1981), Davis (1986), Ranta et al. (1989), and Berthouex and Brown (1994).

Contouring and surface mapping was done with the Surfer® Version 9 software (http://www.goldensoftware.com/products/surfer/surfer.shtml) and with the ArcView 9 geographic information system (GIS) software (http://www.esri.com/software/arcview/ index.html).

6.4.1 Data preparation

Many groundwater variables tend to be skewed right, that is, there are numerous small values and a few large values (e.g. Güler et al. 2002). Large departures from normality, particularly in the form of skewness, or lack of symmetry, can invalidate results. Statistical inference tests based on the normal distribution are sensitive to extreme values and outliers. Their presence may reduce the power of the test and also affect regional trends. Extreme values and data outliers should be checked before statistical analyses.

For Pearson product-moment correlation analysis, analysis of variance and multivariate statistical techniques the data should be transformed to correspond to normally distributed data (e.g. Reimann & Filzmoser 2000). In this study the Box-Cox method was used for normality transformations (Box & Cox 1964, Howarth & Earle 1979). In this method a new value z is calculated for each original value of the variable x based on the following equations:

$$z = (x^{\lambda} - 1)/\lambda \quad (\lambda \neq 0, x > 0) \tag{50}$$

and

$$z = \ln(x) \quad (\lambda = 0, x > 0) \tag{51}$$

All variables equal to zero except the well yield were replaced with 0,1 before transformations. The yield of a dry well was entered into transformations as a nominal figure of 1 Lhr^{-1}. The λ-values were iterated with a computer as long as the values of each variable transformed with the Box-Cox method were so close to normal distribution as possible. The normality of transformed variables was investigated during the iteration process with the Shapiro-Wilk W-statistics (Shapiro & Wilk 1965).

Transformatted variables were standardized to the mean 0,0 and variance 1,0 to avoid of having variables with large variances dominating, for example, factor loadings. This was done by calculating their standard scores (z-scores) as follows:

$$zi = xi - \bar{x}/s, \tag{52}$$

where zi = the standard score of the sample i, xi = the value of sample i, $\bar{x}$ = the mean and s = the standard deviation (e.g. Güler et al 2002).

Standardization scales the transformed data so that it centers about a mean of zero. In this way, each variable has equal weight in the statistical analyses. Besides normalizing and reducing outliers, the transformations also tend to homogenize the variance of the distribution (Güler et al. 2002).

6.4.2 Descriptive and bivariate statistics

The distribution characteristics of each variable in the database were evaluated by data screening with univariate and bivariate statistical methods. The different variables were evaluated using central tendency (mean, median) and dispersion (standard deviation, maximum, minimum) and by graphical displays such as frequency histograms, charts, scatter plots, etc. Interquartile range and confidence intervals about the quartiles are good descriptions of central tendency and variability, too (Knopman 1990). In addition to distribution characteristics, descriptive statistics could help to diagnose outliers, coding errors and the need for transformation of the data (e.g. Steinhorst & Williams 1985, Davis 1986, du Toit et al. 1986, Berthouex & Brown 1994, Güler et al. 2002).

Pearson´s product moment correlation and nonparametric Spearman´s rank correlation tests were performed between variables. Their null hypothesis is that there is no correlation. The p-values of each test showed the level of significance for which the test statistic lies on the boundary between acceptance and rejection of the null hypothesis. A “very highly significant” and "highly significant" trend referred to relationships significant at an alpha level of $\leq$0,001 and $\leq$0,01, respectively, and a "significant" trend referred to those at a significance level of $\leq$0,05 alpha. Data trends that were not significant at the $\leq$0,05 alpha were described as being "not significant". Spearman´s rank correlation test is resistant to the influence of outliers and avoids assumptions about the nature of the dependence, i.e. it can be helpful in detecting nonlinear relationships between attributes (Conover 1980, Knopman & Hollyday 1993, Banks et al. 2005). Partial correlation measures the strength of relationship between variables, controlling for the effect of one or more other variables (e.g. SAS 1990b).

Linear regression was used to estimate and test the significance of both continuous and categorical explanatory variables on each response variable. The coefficient of determination (also called the goodness of fit), R^2, was used to indicate how much of the variation in the response variable was explained by the regression model (e.g. Knopman & Hollyday 1993). The significance of the regression slope and its variance were tested against F- and τ-statistics (e.g. SAS 1990c).

For testing whether the means of various groups of data were statistically different, the rank sum tests, Wilcoxon-Mann-Whitney and Kruskal-Wallis, were used. These tests are the nonparametric equivalent of ANOVA on two groups and more than two groups of a single variable, respectively (e.g. Siddiqui & Parizek 1972, Davis 1986). The p-value gives the probability of making a Type I error: rejecting the null hypothesis when it is true that the means are about the same (e.g. Knopman & Hollyday 1993). A very low p-value means that it is likely that another sample of a similar size also would show a difference between groups. In addition, statistic with median scores was used. Two-sample data produced the two-sample median test. For more than two groups the test was equivalent to the Brown-Mood test. Median scores equaled 1 for observations greater than the median, and 0 otherwise (Conover 1980).

Chi-square goodness of fit test was used to test whether the observed proportions for a categorical variable differed from hypothesized proportions (e.g. SAS 1990b). As earlier has been described (Chapter 6.1.4), the reliability of the questionnaire data and the adequacy of land use sampling points were tested with the non-parametric paired sample Wilcoxon sign rank test.

6.4.3 Multivariate data analysis

Well yield and hydraulic properties are multivariate concepts. They are not thoroughly defined by any single constituent or summary value and conventional bivariate analyses do not reveal the relative importance of each factor that actually influences the well production properties (e.g. Steinhorst & Williams 1985, Lewis 1990, Srinivasa Rao et al. 2000, Henriksen 2003a, 2006b). The idea of multivariate analysis is to interpret the governing processes through data reduction and classification (e.g. Suk & Lee 1999, Meng & Maynard 2001).

Numerous multivariate techniques exist which may be used in statistical analysis. Factor, cluster and discriminant analyses, multivariate analysis of variance and multiple regression analysis are the most widely used in earth sciences (e.g. Hitchon et al. 1971, Draper & Smith 1981, Davis 1986, Zecharias & Brutsaert 1988, Berthouex & Brown 1994, Suk & Lee 1999). In R-mode analysis a large number of variables can be reduced to a minimum number of uncorrelated multidimensions, which thus define the multivariate population in a more simplified structure. Q-mode analysis is used to classify the cases (wells) into statistically independent groups, which define different units (e.g. Hitchon et al. 1971, Seyhan et al. 1985, Usunoff & Guzmán-Guzmán 1989).

In this study, stepwise multiple linear regression analysis was used to evaluate the relative importance of a series of factors on well yield and hydraulic properties of the crystalline bedrock. In parametric multiple regression, the simultaneous dependence of a hydraulic property variable on several other independent variables can be investigated (e.g. Brook 1988, Houston & Lewis 1988, Henriksen 2006b). Multiple regression analysis requires that parameters used as independent variables in regression relationships are uncorrelated with one another. The coefficient of determination, R^2, was used to indicate the variation explained by the regression model. However, this must be done with caution, because simply adding explanatory variables increases R^2 without any proof that the more complex model is better (e.g. Knopman & Hollyday 1993).

Multivariate regression is an efficient tool for predicting well yield simultaneously with many potentially independent variables (e.g. Jammallo 1984, Clarke & McFadden 1991, Moore et al. 2002a). Brook et al. (1986) subjected significant components of the principal component analysis to stepwise multiple regression analysis to determine their power to predict well productivity in a limestone aquifer in Georgia, USA.

The objective of principal factor analysis, the most commonly used factor analysis technique, is to reduce the data (R mode or Q mode) by detecting the possible underlying but unobservable pattern of relationships such that the given data may be rearranged into a smaller number of independent (common) factors which account for observed between variable interrelationships (e.g. Seyhan et al. 1985, Usunoff & Guzmán-Guzmán 1989, Suk & Lee 1999, Reimann et al. 2002).

In this study, factor analysis followed three main steps, namely: factor extraction, rotation of factors, and calculation of scores for each factor. At first, the correlation matrix for all variables was calculated. Then its principal components (eigenvectors) were obtained (e.g. Davis 1986, Hussein 2004). The size of the eigenvalue represents the variance of the original data that has been extracted on to each factor. The communality is a measure of the fraction of the variance of each variable that is explained by the factors extracted (Hitchon et al. 1971, Lawrence & Upchurch 1982). The first common factor is related to the eigenvalue having the highest contribution to the covariance relationship. The second common factor, which is orthogonal to the first, has the second highest contribution to the relationship and so forth. Generally the factors with eigenvalues less than 1,0 are not significant. Orthogonal varimax rotation is commonly used in hydrogeological studies (e.g. Seyhan et al. 1985, Usunoff & Guzmán-Guzmán 1989, Suk & Lee 1999, Reimann et al. 2002).

Discriminant analysis assumes a priori knowledge about the grouping of samples and it tests whether this grouping is statistically significant. Additionally, discriminant analysis can assign samples of unknown origin to a specific group (e.g. Dekkers et al. 1989). In this study, stepwise discriminant analysis was used to select those variables, which were best suited to differentiate between different groups of observations. Then discriminant analysis was assigned to examine various combinations of factors affecting well yield and hydraulic properties (e.g. Sander et al. 1996).

For reliable tests, statistical analyses need 50 samples, at least. For multivariate analyses there ought to be a sufficient number of samples for the number of variables. The rule of thumb for minimum number of samples is three times the number of variables. Incomplete data set (missing variable values) limits the use of multivariate analysis. The theory of multivariate analyses with practical examples is well described, for example, in Cooley and Lohnes (1985), Howarth (1985), Brown (1998), Güler et al. (2002), and Johnson and Wichern (2002).

7 RESULTS

7.1 Technical properties and water quality of drilled wells

7.1.1 General

The author discussed the amount of drilled wells in Central Finland with the local water well contractor in 1993. As a result of the discussion it could be estimated that about one third of the drilled wells of Central Finland was compiled in the CF database at that time (n=1399). In the beginning of 2009 there were 2,509 wells in the database. However, it is obvious that only a small part of the wells drilled in 2000's have been compiled in the database (Fig. 43). Hence, it is estimated that some 9,000 wells in all have been drilled in Central Finland during the last 60 years.

According to Natukka (1955), there were less than 100 drilled wells in Finland before 1940's. In the middle of 1950's the number of wells was about 1,000 (Natukka 1955). At least 60 of them were situated in Central Finland (Fig. 43). According to Laakso (1966a), the number of wells drilled 1951-1963 in Finland should have been approximately 5,000. From these numbers it is estimated that the present number of drilled wells in Finland should be around 150,000-160,000, which agrees well with earlier estimations (Rönkä & Niini 2005, Hatva et al. 2008). Around 2,000-5,000 new wells are drilled every year in Finland (Rönkä 1983, Olofsson & Rönkä 2007).

The spatial pattern of bedrock wells in Central Finland follows the population pattern, being highest in densely populated areas along the main roads and lowest where population is sparse (Fig. 44). Nearly one third of the wells have been drilled in the municipality of Jyväskylä and its neighborhood. However, there are wells in every part of study area. Hence, a good well distribution can be considered to exist throughout the study area. Nearly half of the wells have been drilled by the same contractor, and the three biggest contractors have drilled about 80% of the wells in Central Finland.

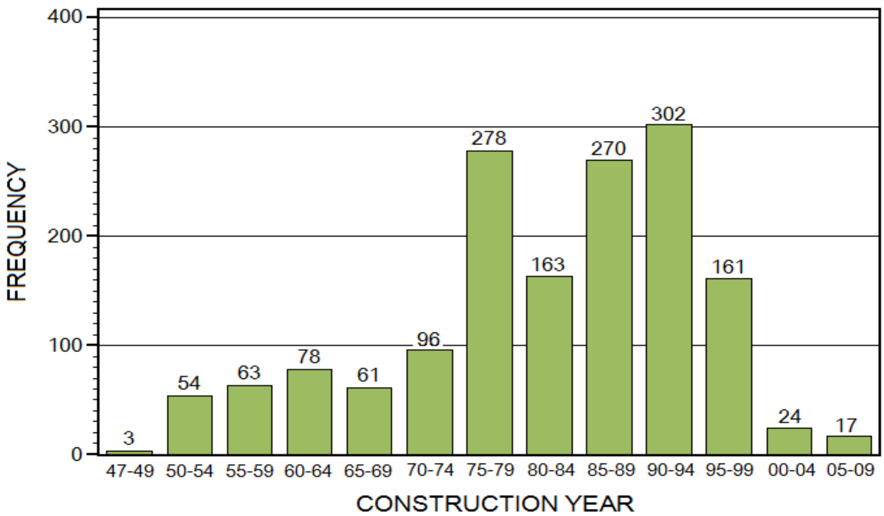


Fig. 43. Bar chart showing the frequency of private drilled wells in the CF database in five-year periods of construction (n=1570). The number of wells per bar is indicated above each bar.

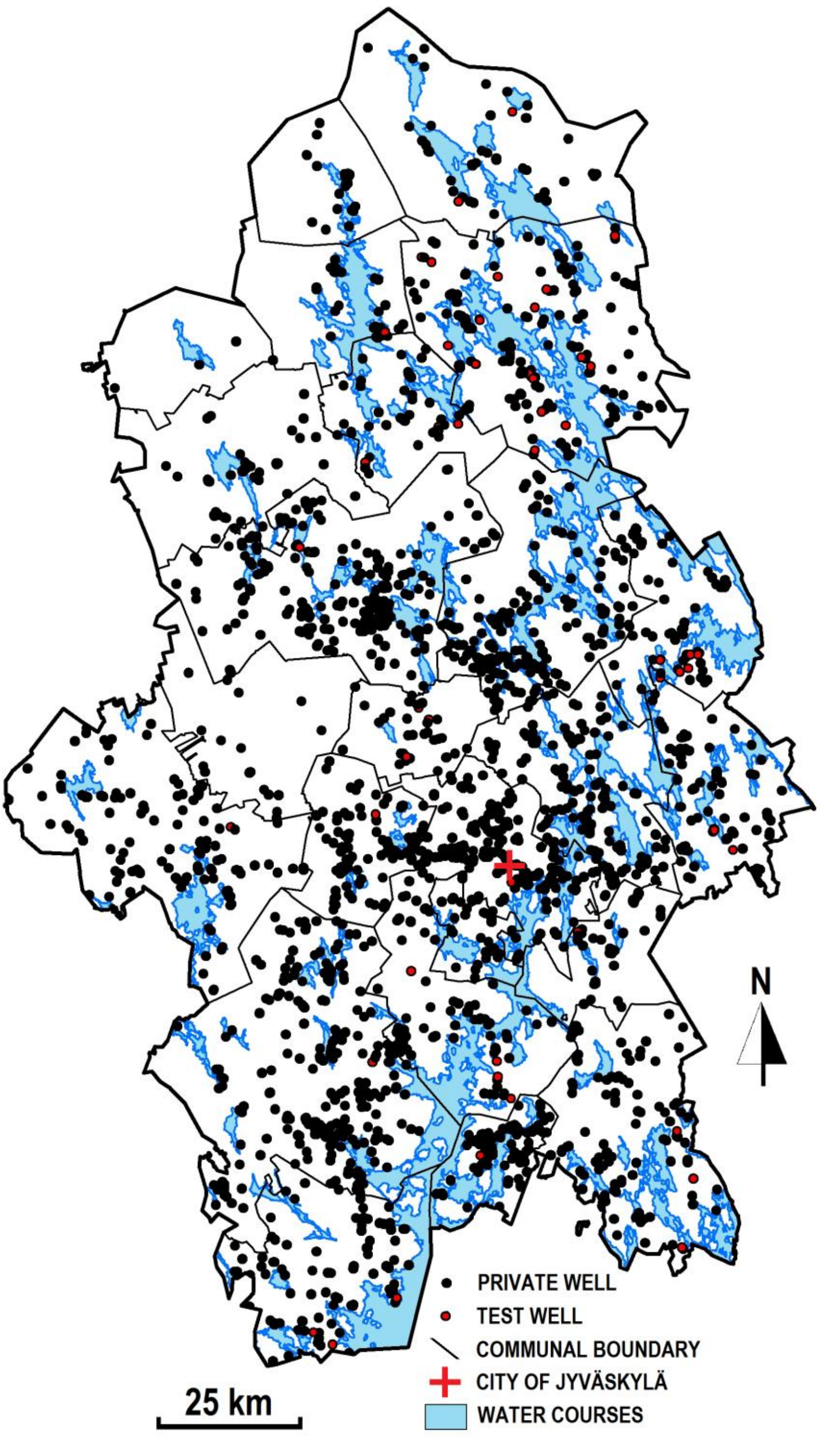


Fig. 44. The spatial distribution of the wells in the Central Finland drilled well database (n=2479).

In 1950's most wells were drilled for public buildings and private enterprises in urban areas (Fig. 45). From that onwards the majority of well drillings has been done for private farmhouses in rural areas, and especially from 1980's, for the needs of domestic households in suburban areas. Test well drillings began in 1980's.

In Sweden, A.E. Nordenskiöld suggested groundwater production from the bedrock by drilling wells for the first time in 1891 (Beyer 1968). The first practical experiment to explore bedrock groundwater was made in 1894 by drilling into gneisses on the coast of Östergötland 150 km southeast of Stockholm (Meier & Petersson 1951).

Carlstedt (1984) has estimated that in Sweden there were about 400,000 private wells in 1984, of which 300,000 were drilled into the bedrock. Between 5,000 and 10,000 new wells were drilled every year, mainly for farms and leisure houses (Gustafson & Krásný 1994). In the beginning of 2000´s there were 400,000...500,000 drilled wells in Sweden (Salonen et al. 2003) and the amount of wells increased, mostly due to energy wells, with almost 25,000 wells a year (Risberg 2004).

In Norway, there were about 800 bedrock boreholes in the beginning of the 1950's (Morland 1997). In the beginning of 2007, there were between 150,000 and 250,000 drilled wells in Norway (Gundersen & de Beer 2007; Pål Gundersen, written communication, 27/3/2007). Around 4,000 new boreholes are added to this total each year (Frengstad & Banks 2007).

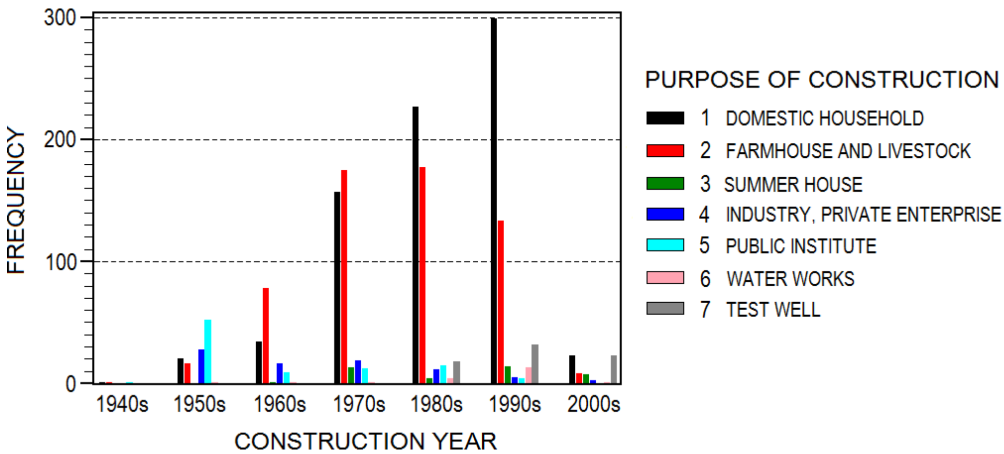


Fig. 45. Bar chart showing the purpose of construction of drilled wells in the CF database in ten-year periods of construction (n=1656, test wells included).

7.1.2 *Well construction data*

The median depth of private drilled wells in Central Finland is 73 m (n=1906). Test wells (n=73) are nearly 20 meters deeper (Table 10). The deepest well (505 m) is situated at Himos in the municipality of Jämsä, but because it was drilled not only for water supply (Chapter 6.1.5) it was left out from statistical analyses. The 90% depth quantile for private wells is around 150 m; the corresponding 10-percentile is 30 m (Table 11). The distribution of private wells in different depth groups is presented in Fig. 46.

Table 10. Well construction information for drilled wells in the CF database (A=private wells, B=test wells, C=all wells). Median values are in bold. The explanations of variable abbreviations are given in Table 3 and in Chapter 6.3.2.

A				PRIVATE WELLS				
Variable	**Dim**	**n**	**Mean**	**Std dev**	**Median**	**Min**	**Max**	**Range**
CY	a	1570	1982	12	**1984**	1947	2008	61
ASL	m a.s.l.	2322	130	33	**123**	79	255	176
CAS	m	780	10	7	**10**	0	57	57
GWT	m	709	6	5	**5**	0	40	40
GWL	m a.s.l.	702	121	31	**115**	79	235	156
STR	m	741	57	44	**45**	1	250	249
SASL	m a.s.l.	741	73	55	**76**	-124	238	362
DEPTH	m	1906	83	51	**73**	9	355	346
DIA	mm	1311	128	22	**115**	86	194	108
SAT	m	706	70	48	**58**	2	340	338
BASL	m a.s.l.	1894	47	61	**52**	-250	222	472
PUMP	m	864	55	33	**50**	0	200	200
WDEM	m^3d^{-1}	674	2,2	6	**0,9**	0,1	70	69,9
B				**TEST WELLS**				
Variable	**Dim**	**n**	**Mean**	**Std dev**	**Median**	**Min**	**Max**	**Range**
CY	a	73	1996	6	**1997**	1985	2008	23
ASL	m a.s.l.	73	124	28	**117**	82	209	127
CAS	m	67	12	7	**12**	2	36	34
GWT	m	63	3	3	**2**	0	14	14
GWL	m a.s.l.	63	122	28	**116**	79	206	127
STR	m	66	30	26	**22**	3	120	117
SASL	m a.s.l.	66	95	41	**98**	-17	198	215
DEPTH	m	73	93	35	**91**	19	215	196
DIA	mm	72	139	18	**140**	110	219	109
SAT	m	63	74	28	**75**	8	127	119
BASL	m a.s.l.	73	31	46	**33**	-93	179	272
PUMP	m	7	97	53	**110**	40	190	150
WDEM	m^3d^{-1}	68	63	71	**30**	4	250	246
C				**ALL WELLS**				
Variable	**Dim**	**n**	**Mean**	**Std dev**	**Median**	**Min**	**Max**	**Range**
CY	a	1643	1982	13	**1985**	1947	2008	61
ASL	m a.s.l.	2395	130	33	**123**	79	255	176
CAS	m	847	11	7	**10**	0	57	57
GWT	m	772	6	5	**5**	0	40	40
GWL	m a.s.l.	765	121	31	**115**	79	235	156
STR	m	807	55	43	**42**	1	250	249
SASL	m a.s.l.	807	74	54	**78**	-124	238	362
DEPTH	m	1979	84	50	**73**	9	355	346
DIA	mm	1383	129	22	**120**	86	219	133
SAT	m	769	70	46	**60**	2	340	338
BASL	m a.s.l.	1967	46	61	**51**	-250	222	472
PUMP	m	871	55	34	**51**	0	200	200
WDEM	m^3d^{-1}	742	8	28	**1**	0,1	250	249,9

The typical diameter (DIA) of private wells varies from 115 to 165 mm; test wells are most commonly 140-165 mm in diameter (Table 10). The median length of the borehole casing (CAS) is 10 meters. The thickest soil layer penetrated by drilling tools is 52 m with casing reaching 57 meters from the ground surface. The median depth to the groundwater table (GWT) is 5 meters. There are 51 artesian drilled wells in the CF database (6,7% of 765 wells).The median depth to the first (main) water strike (STR) in test wells (22 m) is half of that of private wells (45 m). Private wells have been drilled 25-30 meters below the first (main) water strike (Table 10).

The median open well section below the groundwater table (SAT) is 58 m for private wells and 75 meters for test wells. The height of the groundwater level (GWL) is at its lowest 79 m a.s.l., which is the same as the water level of Lake Päijänne, the largest and lowest surface water body in the study area. At its lowest the bottom of a borehole (BASL) extends 250 meters below the water level of the Baltic Sea. The median water demand (WDEM) of private drilled wells is 900 Lday^{-1}; the mean is 2,160 Lday^{-1} (Table 10).

Table 11. The distribution of drilled well depth (m) in the CF database. Median values are in bold.

Parameter or quantile	Private wells Depth m n=1906 96%	Test wells Depth m n=73 4%	All wells Depth m n=1979 100%
Mean	83	93	84
Std dev	51	35	50
Mode	40	91	40
Range	346	196	346
100% Max	355	215	355
99%	247	215	247
95%	181	151	181
90%	151	133	151
75% Q_3	109	121	109
50% Md	**73**	**91**	**73**
25% Q_1	46	72	46
10%	30	43	31
5%	23	40	23
1%	15	19	15
0% Min	9	19	9

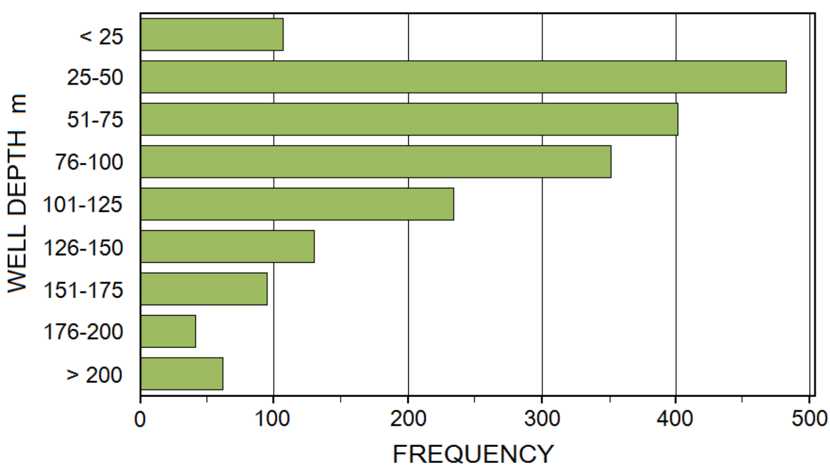


Fig. 46. Bar chart showing the distribution of private drilled wells in the CF database in different depth groups (n=1906).

In Finland, the average depth of the cable-tool wells was 58 m in the mid-50's (Table 12; n=555, Natukka 1955). According to Laakso (1966a, 1966b), the mean depth of cable-tool wells drilled by the same drilling company in 1951-1963 in Finland was 68 m (n=1108). In Laakso´s database there were 87 wells drilled in Central Finland with a mean depth of 51 m. Hyyppä (1984) considered (for water quality purposes) a data set of 908 wells around Finland, and recorded an average and a median depth of 48 m and 40 m, respectively. Lahermo (1970, 1971) studied drilled wells in Finnish Lapland and in the Rapakivi granite area of southeastern Finland. The average depth of wells was 33 and 65 m, respectively (n=213 and 147). In Rönkä´s material (1983,1988) the average depth of crystalline rock wells in southern and central Finland was 60 m (n=385).

Table 12. Mean and median depths for drilled wells in unweathered crystalline rocks around the world.

Location	Mean m	Median m	n	Reference
Finland	58	--	555	Natukka 1955
Finland	68	--	1108	Laakso 1966a, 1966b
Central Finland	51	--	87	Laakso 1966a
Finland	48	40	908	Hyyppä 1984
Finnish Lapland	33	--	213	Lahermo 1970
SE Finland	65	--	147	Lahermo 1971
Southern and central Finland	60	--	385	Rönkä 1983
Central Finland	**84**	**73**	**1979**	**This study**
Sweden	50	--	--	Meier & Petersson 1951
Sweden	--	64	59000	Fagerlind 1986
Sweden	--	75	102562	Fagerlind 1989
Sweden 1989	--	72	8337	Axelsson et al. 1991
Norway	--	56	12757	Morland 1997
North Carolina, USA	--	36	5221	Daniel 1987, 1989
Georgia, USA	126	--	257	Brook 1988
Virginia, USA	109	--	3561	Sutphin et al. 2001

Fagerlind (1986) and Gustafson (2002) document the findings from a statistical analysis of the Swedish data set, performed in around 1986. They found that the median depth of the wells was 64 m from a data set of 59,000 wells (Table 12). At the end of March 1989, the median depth of the wells in Sweden was 75 m with 25% and 75% percentiles at 51 m and 100 m, respectively (n=102562; Fagerlind 1989). The median depth of drilled wells in Norway is 56 m (n=12757; Morland 1997, Banks et al. 2005).

According to Daniel (1987, 1989), the median depth of wells drilled in crystalline rocks of the Piedmont and Blue Ridge Provinces of North Carolina, USA, is 36 m (n=5221). In crystalline bedrock of Georgia, USA, the well depth ranges 20-291 m with an average depth of 126 m (n=257; Brook 1988). The mean depth of wells drilled in crystalline rocks of Loudoun County, Virginia, USA, is 109 m (n=3561; Sutphin et al. 2001).

In the São Paulo weathered crystalline rock region, Brazil, the depth of wells drilled mainly for industry (n=1129) ranges between 50 and 300 m (Rebouças & Cavalcante 1990). In weathered basement aquifers of Eastern Senegal the average depth of wells is 58 meters (n=177; Diop & Tijani 2008), and an average borehole depth of 40-80 m is estimated for the crystalline basement rocks in Nigeria (Adelana et al. 2008). In Precambrian gneiss and schist area of Kandy, Sri Lanka, the median depth of drilled wells is 76 m (n=536; Johansson 2005).

Spearman rank correlation coefficients between well construction variables are shown in Table 13. They indicate that well diameter DIA, casing depth CAS and well DEPTH have increased during the years CY. At the same time the depth to the first (main) water strike

STR and the saturated open well section SAT have increased. The median well depth increases with STR (Fig. 47). This has been noticed in crystalline rocks of South Africa, too (Holland & Witthüser 2011). STR also correlates with GWT but not with GWL. The height of the well site ASL does not influence the well depth.

Table 13. Spearman rank correlation coefficients (r) with significance levels (p) and number of observations (n) in that order between well construction variables for all drilled wells in the CF database. Pairs having highly significant correlation coefficients ≥ |0,5000| are in bold.

	ASL	CY	DIA	CAS	GWT	GWL	BASL	SAT	STR	SASL	DEPTH
CY r p n	0,0544 0,0280 1633										
DIA	0,0393 0,1455 1372	**0,6331** **<0,0001** **1290**									
CAS	-0,1038 0,0026 842	0,3038 <0,0001 792	0,3350 <0,0001 780								
GWT	0,1338 0,0002 765	-0,0459 0,2250 702	0,0281 0,4694 667	-0,0816 0,0850 447							
GWL	**0,9853** **<0,0001** **765**	0,0953 0,0119 695	0,0718 0,0653 660	-0,0982 0,0384 445	-0,0090 0,8034 765						
BASL	**0,5677** **<0,0001** **1967**	-0,4139 <0,0001 1601	-0,4001 <0,0001 1367	-0,2681 <0,0001 839	-0,0592 0,1028 762	**0,5288** **<0,0001** **762**					
SAT	0,0600 0,0981 762	0,5425 <0,0001 700	0,4381 <0,0001 665	0,1715 0,0003 446	0,1931 <0,0001 769	0,0489 0,1778 762	**-0,7690** **<0,0001** **762**				
STR	-0,0208 0,5544 807	0,3176 <0,0001 754	0,3182 <0,0001 721	0,1287 0,0043 490	0,2320 <0,0001 460	-0,0298 0,5241 460	**-0,5826** **<0,0001** **805**	**0,6564** **<0,0001** **459**			
SASL	**0,5949** **<0,0001** **807**	-0,2274 <0,0001 754	-0,2227 <0,0001 721	-0,1481 0,0010 490	-0,1045 0,0250 460	**0,5906** **<0,0001** **460**	**0,8281** **<0,0001** **805**	**-0,5203** **<0,0001** **459**	**-0,7678** **<0,0001** **807**		
DEPTH	-0,0260 0,2497 1967	**0,5615** **<0,0001** **1611**	0,4977 <0,0001 1378	0,2484 <0,0001 845	0,1480 <0,0001 769	0,0404 0,2658 762	**-0,7990** **<0,0001** **1967**	**0,9829** **<0,0001** **769**	**0,6949** **<0,0001** **805**	**-0,5579** **<0,0001** **805**	
WDEM	-0,0682 0,0635 742	0,1154 0,0029 663	0,1467 0,0005 560	0,0927 0,0749 370	-0,1760 0,0015 321	0,0071 0,8988 321	-0,1785 <0,0001 706	0,1406 0,0120 319	-0,1083 0,0400 360	0,0860 0,1034 360	0,1806 <0,0001 706

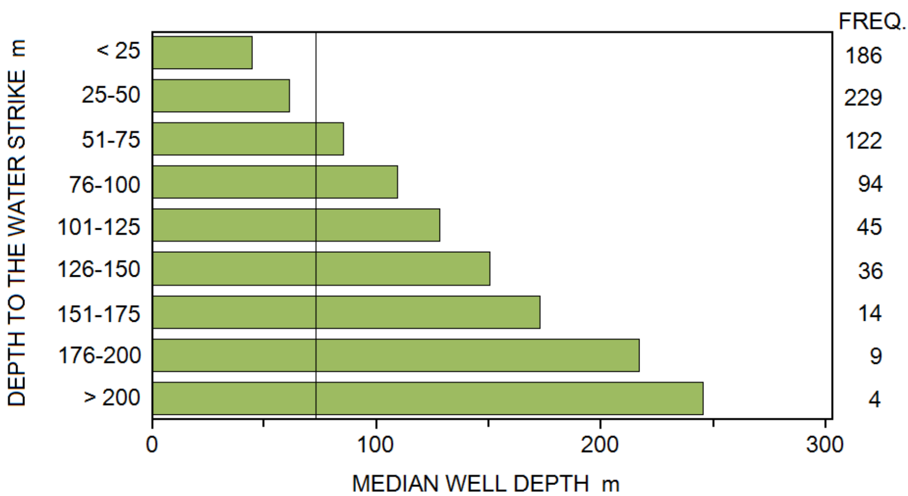


Fig. 47. Bar chart showing the median depth of private drilled wells in the CF database in different STR groups (n=739). The overall median depth of private wells (73 m) is marked with a vertical line. The number of wells per bar is indicated in the rightmost column.

Roughly one third of the wells are interpreted to be weakly-to-semi-confined having hydraulic heads reaching into the casing above the bedrock interface (e.g. Boutt et al. 2010). Groundwater level GWL closely follows ASL (Fig. 48) as is typical for crystalline rock terrains indicating low transmissivities and a topographically driven flow system that is not strongly confined (Lachassagne et al. 2001, Banks & Robins 2002, Owen et al. 2003, Marklund 2009, Boutt et al. 2010).

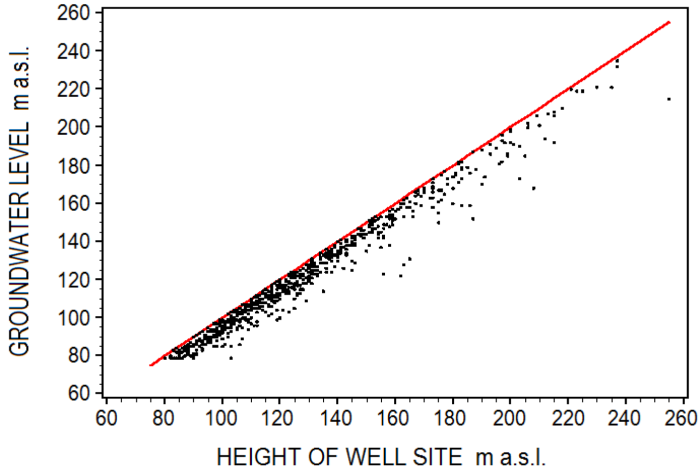


Fig. 48. The relation between the groundwater level (GWL) and the height of the drilled well site (ASL) in the CF database (n=765, test wells included). On the diagonal line the values are equal (=flowing artesian wells; n=51).

In drilling crystalline aquifers in Sweden, boreholes are normally abandoned after around 100 m if the required yield has not been achieved (Gustafson 2002). In New Hampshire, USA, a well depth of 120 m has become a common depth at which a driller stops drilling if the desired yield has not been reached (Moore et al. 2002a). In Central Finland, well drillers give a yield guarantee for an extra price, which means that if at a well depth of 120-150 m enough water has not been detected, the drillers continue drilling on their own account or they use hydraulic fracturing to increase the well yield. The guaranteed well yield though is commonly 1-2 m^3d^{-1}, i.e. 40-80 Lhr^{-1}.

7.1.3 Yield of wells

The median yield of private drilled wells in Central Finland is 650 Lhr^{-1}. The corresponding yield of test wells is tree times that big, 2,000 Lhr^{-1}. The overall median yield of drilled wells is 700 Lhr^{-1} (Table 14). The highest well yield so far detected in Central Finland is 24,000 Lhr^{-1} (private well no. 931093 in the municipality of Viitasaari).

Variations in well yields within the study area are large and the distribution of the borehole yields is heavily skewed towards low yields (Fig. 49). The mean value is forced up by the existence of a few boreholes with very high yields and the median yield value is considerably lower than the arithmetic mean (Table 14). The skewness in yield distribution is common in crystalline rock terranes (e.g. Bryn 1961, Davis & DeWiest 1966, Cederstrom 1972, Rohr-Torp 1987, Banks et al. 1992a, 1994, 2005, Kaehler & Hsieh 1994, Henriksen 2003a). Based on yield distribution one can quantify the probability of achieving the required yield from wells drilled in crystalline bedrock (e.g. Banks et al. 1994, Banks 1998).

Table 14. The distribution of drilled well yield (Q Lhr^{-1}) in the CF database. Median values are in bold.

Parameter or quantile	Private wells Q Lhr^{-1} n=1235 94%	Test wells Q Lhr^{-1} n=73 6%	All wells Q Lhr^{-1} n=1308 100%
Mean	1488	3348	1592
Std dev	2140	3810	2303
Mode	500	300	500
Range	24000	20000	24000
100% Max	24000	20000	24000
99%	10000	20000	10000
95%	6000	10000	6000
90%	4000	9000	4600
75% Q_3	1800	5000	2000
50% Md	**650**	**2000**	**700**
25% Q_1	300	400	300
10%	100	75	100
5%	40	5	40
1%	0	0	0
0% Min	0	0	0

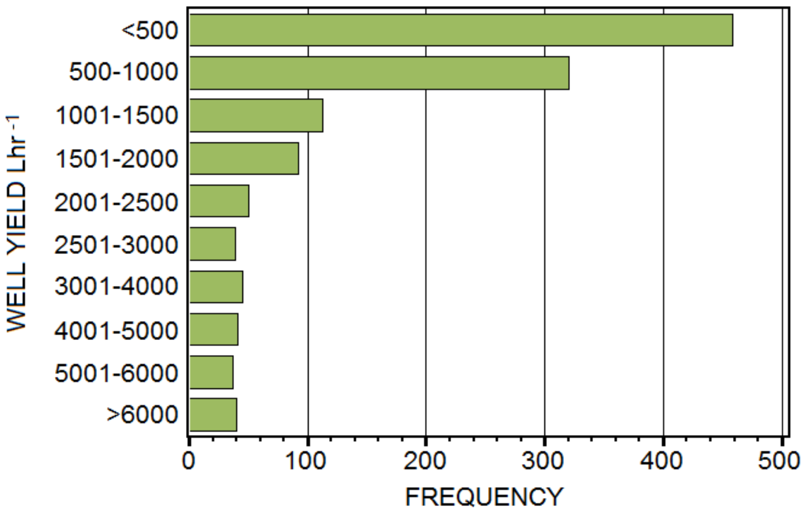


Fig. 49. Bar chart showing the distribution of private drilled wells in the CF database in different yield groups (n=1235). Note the different division on the vertical axis.

In Finland, the average yield of the cable-tool wells was 2,270 Lhr^{-1} in the mid-50’s (Table 15; n=555, Natukka 1955). According to Laakso (1966a, 1966b), the mean yield of cable-tool wells drilled by the same drilling company in 1951-1963 in Finland was 2,465 Lhr^{-1} (Md ca. 1,600 Lhr^{-1}; n=1108). In Laakso´s database there were 87 wells drilled in Central Finland with a mean yield of 1,683 Lhr^{-1} (Md ca. 800 Lhr^{-1}). According to Lahermo (1970, 1971), the average yield of wells in Finnish Lapland and in the Rapakivi granite area of southeastern Finland was 2,050 and 3,200 Lhr^{-1}, respectively (n=213 and 147).

Table 15. Mean and median yields for drilled wells in unweathered crystalline rocks around the world.

Location	Mean Lhr^{-1}	Median Lhr^{-1}	n	Reference
Finland	2270	--	555	Natukka 1955
Finland	2465	1600	1108	Laakso 1966a, 1966b
Central Finland	1683	800	87	Laakso 1966a
Finnish Lapland	2050	--	213	Lahermo 1970
SE Finland	3200	--	147	Lahermo 1971
Southern and central Finland	1188	--	368	Rönkä 1983
Central Finland	**1592**	**700**	**1308**	**This study**
Sweden	--	600	59000	Fagerlind 1986
Sweden	--	700	102562	Fagerlind 1989
Sweden 1989	--	700	8337	Axelsson et al. 1991
Norway, granites	--	700	87	Bryn 1961
Norway, gneisses	--	1050	908	Bryn 1961
Norway	--	600	12757	Morland 1997
Bohemian Massif, Upper Austria	2880	--	712	Stibitz & Fleischmann 1996
Sierra Nevada, USA	--	1800	239	Davis & Turk 1964
Colorado, USA	1120	300	138	Snow 1968
North Carolina, USA	--	2271	5234	Daniel 1987, 1989
Maine, USA	--	1470	2552	Caswell 1987, Mabee 1992
Virginia, USA	3640	--	3561	Sutphin et al. 2001
New Hampshire, USA	--	1300	20308	Moore et al. 2002a
Pennsylvania, USA	--	680-1135	--	Low 2004
Manitoba, Canada	--	700	196	Betcher et al. 1995
British Columbia, Canada	--	870	3240	Kohut 2006
Vancouver Island, Canada	--	900	433	Kenny et al. 2006

In Rönkä´s study material (1983, 1988), the average yield of crystalline rock wells in southern and central Finland was 1,188 Lhr^{-1} (n=368); 2,9% of the wells were completely dry and 1,5% gave less than 20 Lhr^{-1} of water. In the CF database, there are 27 dry wells (0 Lhr^{-1}; 2,1% of 1,308 wells) and 18 wells (1,4%) give less than 20 Lhr^{-1} of water. In the Gothenburg region, southwestern Sweden, 34 of 1,282 wells (2,7%) situated in crystalline rock area are listed as dry (Wladis 1995). Of 712 boreholes drilled in the Bohemian Massif in Upper Austria, nine wells (1,3%) did not yield any water; the maximum yield was 14,400 Lhr^{-1} and the average yield 2,880 Lhr^{-1} (Stibitz & Fleischmann 1996). In the crystalline rock area of Pinardville, New Hampshire, USA, zero-yield wells may comprise about 10% of the data (Drew et al. 1999, 2001). In weathered crystalline rock areas of Burkina Faso, 24% of drilled wells show a zero discharge (total n=14645; Courtois et al. 2010);

The average yield of wells drilled in crystalline bedrock areas of Sweden before 1950 was between 750 Lhr^{-1} and 3700 Lhr^{-1} with a maximum quantity of 25,000 Lhr^{-1} (Meier & Petersson 1951). The mean depth of drilled wells was about 50 m. It was found, however, that considerable amounts of water were still derived at depths of between 100 and 200 meters (see also Meier & Sund 1952). The importance of faults as remarkable bedrock aquifers was also noticed (Meier & Petersson 1951). Fagerlind (1986) found that the median yield was 600 Lhr^{-1} from a SGU-data set of 59,000 wells in Sweden (Table 15). The 25% and 75% percentiles were reported as 270 and 1,800 Lhr^{-1}, respectively. At the end of March 1989 the median yield of wells in Sweden was 700 Lhr^{-1} with 25%-ile at 300 Lhr^{-1} and 75%-ile at 2,000 Lhr^{-1} (n=102562; Fagerlind 1989). These figures, which are based on a large data set, are identical with those from all drilled wells in the CF database (Table 14). In Fagerlind´s (1989) data set, 19% of the wells had been developed with hydraulic fracturing.

In Bryn's (1961) study, the median yield of the wells situated near Oslo, Norway, was in granitic areas 700 Lhr^{-1} (n=87) and in gneissose areas 1,050 Lhr^{-1} (n=908; Table 15). According to Morland (1997), the median yield of drilled wells in Norway is 600 Lhr^{-1} (n=12757; see also Banks et al. 2005). Since Morland's (1997) analysis, the Geological Survey of Norway has expended considerable effort on acquiring high-quality data from drillers: the current database (as of March 2008) contains 26,811 records of bedrock well yields, with a median yield of 500 Lhr^{-1} and lower and upper quartile yields of 200 and 1500 Lhr^{-1} (Banks et al. 2010).

The median yield of wells drilled in unweathered crystalline rocks of the eastern part of Sierra Nevada region, USA, is 1,800 Lhr^{-1}; P90% is about 11,500 Lhr^{-1} (n=239, Davis & Turk 1964; Table 15). The median and average yield of 138 random wells in the metamorphic rocks of Colorado Front Range, USA, is 300 Lhr^{-1} and 1,120 Lhr^{-1}, respectively (Snow 1968). In crystalline rocks of Colorado, USA, the yield of drilled wells can vary from 0 to 22,700 Lhr^{-1} (Florquist et al. 1973). In crystalline rocks of Maine, USA, 86% of the drilled wells yield at least 450 Lhr^{-1}; 70% of wells yield less than 2,270 Lhr^{-1} (Rand 1978). The median yield of wells drilled in crystalline rocks in the Blue Ridge of Pennsylvania, USA, vary from 680-1,135 Lhr^{-1} (Low 2004). The well yields of metamorphic and plutonic crystalline rocks of New Hampshire, USA, range from zero to greater than 110,000 Lhr^{-1}. The median is about 1,300 Lhr^{-1}, 25% percentile is at 700 Lhr^{-1} and the 75% percentile at 3,400 Lhr^{-1}; P90% is around 9,000 Lhr^{-1} (n=20308, Moore et al. 2002a). In metamorphic and igneous rocks of Maine, USA, the median yield of drilled wells is 1,470 Lhr^{-1} with 25%-ile at 770 Lhr^{-1} and 75%-ile at 2,780 Lhr^{-1} (n=2552; Caswell 1987, Mabee 1992). According to Daniel (1987, 1989), the median yield of wells drilled in crystalline rocks of the Piedmont and Blue Ridge Provinces of North Carolina, USA, is 2,271 Lhr^{-1} (n=5234). The mean yield of wells drilled in crystalline rocks of Loudoun County, Virginia, USA, is 3,640 Lhr^{-1} (n=3561; Sutphin et al. 2001).

The yield of wells drilled in Precambrian igneous and metamorphic rocks of Manitoba, Canada, range from 0 to more than 50,000 Lhr^{-1}; the median yield is around 700 Lhr^{-1} and the P90%-ile is about 5,500 Lhr^{-1} (n=196; Betcher et al. 1995; Table 15). The median yield of drilled wells in crystalline rocks of British Columbia, southwestern Canada, is about 870 Lhr^{-1} (n=3240; Kohut 2006). In igneous and metamorphic rocks of Vancouver Island, Canada, the median yield of drilled wells is 900 Lhr^{-1} (n=433; Kenny et al. 2006). The median yield of wells drilled in the Precambrian metamorphic crystalline rocks of Kandy District, Sri Lanka, is about 600 Lhr^{-1} (n=2226; Johansson 2005). In the São Paulo weathered crystalline rock region, Brazil, the yield of wells drilled mainly for industry (n=1129) varies from 5,000 to 150,000 Lhr^{-1} (Rebouças & Cavalcante 1990). In the same region the median yield of 312 wells is 5,200 Lhr^{-1} (x=7,300 Lhr^{-1}; Yoshinaga Pereira & Kimmelmann e Silva 2004). In the crystalline rock aquifers of Paraiba State, Brazil, the 25%-ile yield of drilled wells is about 100 Lhr^{-1} (n=973; Filho & Rebouças 1995).

The median yield of wells drilled in crystalline basement of southern Zimbabwe is 1,200 Lhr^{-1} with 25%-ile at 240 Lhr^{-1} and 75%-ile at 3,600 Lhr^{-1}; the P90% is 7,200 Lhr^{-1} and P95% is 18,000 Lhr^{-1} (Owen et al. 2003). In weathered basement aquifers of Eastern Senegal the well yield categorization is as follows: $Q \leq 1{,}000$ Lhr^{-1} 30% of the wells and $Q \geq 10{,}000$ Lhr^{-1} 16% of the wells (n=177; Diop & Tijani 2008). An average yield of 1,440 Lhr^{-1} is estimated for the crystalline basement rocks in Nigeria (Adelana et al. 2008). Nearly 80% of the boreholes in the Precambrian crystalline bedrock of northeastern Ghana yield more than 1,000 Lhr^{-1}, while 9% of them yield higher than 5,000 Lhr^{-1} (n=2458, Agyekum & Dapaah-Siakwan 2008). In Burkina Faso, almost 35% of drilled wells have a discharge lower than or equal to 700 Lhr^{-1} (n=14645); the weathered zone represents some 40% of the total well length (Courtois et al. 2010). Moreover, wells with a yield higher than or equal to 3,600 Lhr^{-1} and 10,000 Lhr^{-1} represent 23% and 6% of the total well population, respectively.

Two boreholes drilled a few meters apart in a fractured aquifer may produce yields that differ totally (Clapp 1911b, Schoeller 1975, Quist 1990, Kaehler & Hsieh 1994, Havlík & Krásný 1998, Drew et al. 1999, Van Tonder 1999, Lachassagne 2008).

7.1.4 Success rate and high-yield wells

Groundwater exploration in crystalline rock areas commonly experiences low success rates and a wide range of yields and high failure rates for boreholes (e.g. Sander 1999). The term 'success rate' can take different meanings. A successful well may be defined as a well that yields more than 10 $Lmin^{-1}$ (600 Lhr^{-1}), which is the minimum requirement to equip the well with a hand pump (Sander et al. 1996, 1997, Acheampong & Hess 1998, Darko & Krásný 2000, Johansson 2005). In addition, a commonly accepted definition of a successful borehole is a sustainable yield of around 500 Lhr^{-1} for a handpump equipped well and 5,000 Lhr^{-1} for a mechanized well (Bernard & Valla 1991, Boeckh 1992, Sami 1996). However, a lower yield than 500 Lhr^{-1} can be accepted for a private well, provided that borehole storage can cover peak demands (Sander 1999). In Sweden, the yield of a successful well for a single home is considered to be 250 Lhr^{-1} (Gustafson 2002).

In Ghana, a successful borehole with a handpump yields not less than 790 Lhr^{-1} (Gyau-Boakye et al. 2008). In Malawi, a successful borehole in relation to hand pump requirement is defined as one that yields at least 900 Lhr^{-1} (Grey et al. 1985). In Senegal, Diop and Tijani (2008) define a successful well with a yield of 1,000 Lhr^{-1} or more. According to Teeuw (1995), a successful borehole yields 1,200 Lhr^{-1}, at least, which is sufficient for hand-pump installation. A successful borehole may also be the one whose yield and water quality satisfy the needs of a particular project (Anon 1989a). In other words, a well that may properly be called successful where but a small water supply is needed is a failure if a large amount is required (Clapp 1911b). In Brazil, wells producing less than 100 Lhr^{-1} have usually been considered unsuccessful (Filho & Rebouças 1995, Rebouças 1999b).

Kent and Gombar (1990) define a successful well in Zimbabwe as a well with specific capacity of not less than 40 $Lhr^{-1}m^{-1}$. Lewis (1990) suggests that an internationally accepted definition of a successful borehole would be 30 $Lhr^{-1}m^{-1}$.

The success rates of drilled wells in the CF database defined in different ways are shown in Table 16.

Table 16. The success rate of drilled wells in the CF database.

YIELD Lhr^{-1}	SUCCESS RATE		
	Private wells	Test wells	All wells
	% of 1235 wells	% of 73 wells	% of 1308 wells
≥100	90,7	87,7	90,5
≥250	78,0	84,9	78,4
≥600	54,8	67,1	55,5
≥1000	42,3	61,6	43,4
≥4000	11,4	34,3	12,7

In this study, the capacity of a high-yield well is 4,000 Lhr^{-1} or more (P90% of private well yields in the CF database; Table 14); the capacity of low-yield wells is ≤100 Lhr^{-1} (P10%). According to this definition, one third of the test wells in Central Finland are high-yield wells (Table 16).

In India, bore wells yielding 3,000 Lhr^{-1} or more are considered high-yield wells (Naik et al. 2001). Also Batte et al. (2008a) consider a well with a yield of greater than 3,000 Lhr^{-1} to be high yielding in Kamuli District, eastern Uganda; low-yield wells produce less than 500 Lhr^{-1}. Kastrinos and Wilkinson (1994) defined a high-yield well in Massachusetts, USA, as a well with yield of 4500 Lhr^{-1} or more. According to Lachassagne et al. (2001), high-yield wells in hard rock aquifers can produce more than 5,000 Lhr^{-1} of water for a long time. Wells yielding more than 6,800 Lhr^{-1} are considered high-yielding wells in Eastern Kentucky Coal Field, USA (7% of 6075 wells; Dinger et al. 2002). In New Hampshire, USA, industry standardized high-yield wells have yields of 9,000 Lhr^{-1} or more; very low-yield wells produce $\leq$100 Lhr^{-1} (Drew et al. 1999, 2001, 2002, 2003, Moore et al. 2002a). In South Africa, high-yield boreholes yield more than 10,800 Lhr^{-1}; a very low borehole yield is less than 360 Lhr^{-1} (Bell & Maud 2000).

According to Lewis (1990), high-yielding boreholes are those with a specific capacity greater than 360 $Lhr^{-1}m^{-1}$. In São Paulo district, Brazil, high-capacity wells have specific capacities Q/s >$1,4x10^{-4}$ m^2s^{-1} (Fernandes & Rudolph 2001). According to Batte et al. (2008a), the transmissivity of high-yield wells in Kamuli District, eastern Uganda, is greater than $9,26x10^{-5}$ m^2s^{-1}.

7.1.5 Well development

In the CF database there exist well development data from 76 wells, most of which are low-yield wells and 14 of them test wells. Both hydraulic fracturing and explosive shooting (blasting) have been used. Well yields from airlift or short-term test pumpings before and after the treatment can be used as a measure of the success of the development. In the following, the success rate is calculated as the percentage improvement in the volume extracted. For this purpose, the yield of a dry well was entered into calculations as a nominal figure of 1 Lhr^{-1} (e.g. Banks et al. 2005). In hydrofracturing, the maximum hydraulic pressure has varied 10-25 MPa, and the amount of water pumped into each borehole has been around 5,000-6,000 liters. No propping agents, e.g. sand (Banks et al. 1996), have been injected into the wells.

The results show that due to the hydraulic fracturing (n=66) the median (short-term) well yield has increased from 45 to 525 Lhr^{-1}, i.e. by more than tenfold (Table 17). Then in percentages the median success rate is 990%. In liters per hour the median yield increase is 410 Lhr^{-1}. The method has increased the yield in all dry wells (n=16) and in about 80% of them by 200 Lhr^{-1} or more. In five wells (8%) the hydrofracturing has not increased the yield. The corresponding results from explosive shooting (n=10) are not as convincing as those from hydrofracturing. The median increase in yield has been only 11 Lhr^{-1}. Half of the wells have not increased their yield at all. The distribution of short-term yields before and after well development with both methods is shown in Fig. 50.

In Australia, Williamson and Woolley (1980) have found that hydraulic fracturing with water-only should be used, as no improvements cannot be accomplished by using the viscous fluid-propping agent mixture. Moreover, if propping agents are used, the grains should be relatively coarse. Hydraulic fracturing was proved most effective for wells yielding less than 900 Lhr^{-1}. In a study of hydraulic fracturing performed in low yielding boreholes in the crystalline basement rocks of Masvingo Province, Zimbabwe, reported by Herbert et al. (1993), 12 boreholes were stimulated with hydraulic fracturing using a single or double packer unit. In 50% of the cases the borehole yield was increased by an average of 80% in the range of 10 to 240%. Hydrofracturing results from South Africa show that 47% of 170 boreholes in crystalline rocks have experienced a 20% yield improvement, at least (Less & Andersen 1993, 1994). In Scotland, hydrofracturing of two public water supply boreholes in Precambrian crystalline bedrock has increased borehole yields by at least one order of magnitude (Cobbing & Ó Dochartaigh 2007).

Banks and Robbins (2002) emphasize that hydraulic fracturing at shallow depths (<25-30 meters) runs the risk of creating fractures to the surface, which would be vulnerable to contamination and thus should be avoided. Kalskin Ramstad (2004) examined borehole hydrofracturing in two hard rock sites in Norway. Hydraulic fracturing with water-only resulted in an overall increase in water yield of the hard rock boreholes. In addition, it was suggested that the increased borehole yields and the improved, reliable and cost-effective hydraulic fracturing techniques in crystalline bedrock would probably increase the interest for groundwater as a domestic water supply for small- to medium sized water works.

Table 17. Well development results of the drilled wells in the CF database.

Method	Variable	Dim	n	Mean	Std dev	Median	Min	Max
Hydraulic fracturing	YIELD before = A	Lhr^{-1}	66	207	516	45	1	3600
	YIELD after = B	Lhr^{-1}	66	872	944	525	10	5000
	INCREASE (B-A)	Lhr^{-1}	66	665	766	410	0	3560
	SUCCESS RATE (B-A)/A*100	%	66	15221	36829	992	0	179900
Explosive shooting	YIELD before = A	Lhr^{-1}	10	120	193	29	1	600
	YIELD after = B	Lhr^{-1}	10	254	293	220	1	900
	INCREASE (B-A)	Lhr^{-1}	10	134	264	11	0	840
	SUCCESS RATE (B-A)/A*100	%	10	2498	6185	138	0	19900

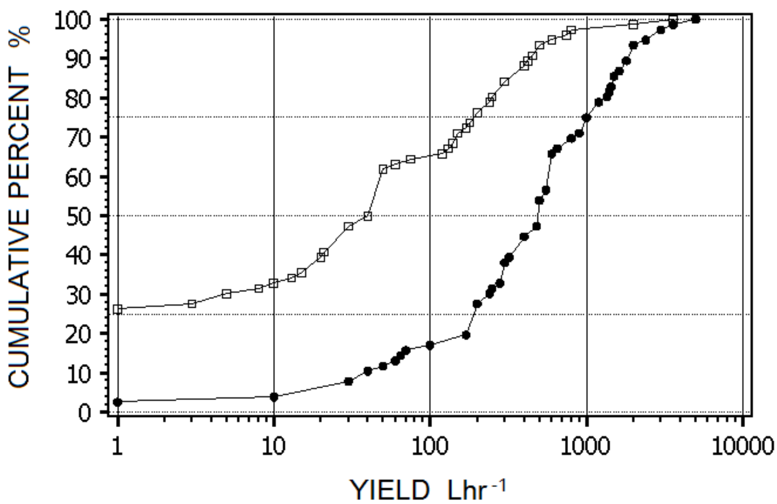


Fig. 50. The cumulative percent frequency of short-term well yields in the CF drilled well database (n=76) before (□) and after (●) well development. The yield of a dry well is expressed as a nominal figure of 1 Lhr^{-1}.

7.1.6 Groundwater quality

Groundwater quality of drilled wells in the CF database has been analyzed in 31 different laboratories between 1952 and 2009. The statistics for groundwater quality are performed in Table 18. These results are based on the samples analyzed in accredited laboratories under public supervision between 1972 and 2008 (total number of sampled wells n=2007, i.e. 80% of the CF database wells). Most samples (75%) have been analyzed in the former laboratory of CETECF, in the Ambiotica laboratory of the Institute for Environmental Research in University of Jyväskylä and in the laboratory of the Geological Survey of Finland.

The sampling and analysis have followed the Finnish standard rules (National Board of Waters 1984). The turbidity of the samples if determined has been < 1,0 FNU. Heavily polluted wells and wells clearly contaminated by surface waters (n=77; 4%) have been excluded from the statistical examination. Most of the metal samples have been passed through a 0,45-µm filter in the field before taking them to the laboratory. For water quality data that are censored, i.e. concentrations of some elements are reported as non-detected, less-than or greater-than (e.g. Güler et al. 2002), the actual values have been used without any replacements (e.g. Berthouex & Brown 1994).

Water quality judgment of drilled wells in Central Finland according to the Decree of the Ministry of Social Affairs and Health in Finland relating to the quality and monitoring of water intended for human consumption in single households (STM 2001; Table 18) is shown in Table 19. A majority of the wells yields good quality groundwater. The most common quality problems in groundwater of drilled wells in Central Finland are manganese, pH, colour, fluoride, iron and bacteria, in this order. The average groundwater quality of the CF database wells is similar to that of other crystalline rock areas (e.g. Hyyppä 1984, 1986, Lahermo et al. 1990, Aastrup et al. 1995, Banks et al. 1998, Frengstad 2002, Knutsson & Olofsson 2002, Moore 2004).

Table 18 (next page). The statistics for groundwater quality of the drilled wells in the CF database. Median values are in bold. Maximum concentrations and guide levels of water intended for human consumption in single households are shown in column STM (2001). Turbidity <1,0 FNU if determined; total number of sampled wells n=2007 (80% of the CF database wells); test wells included. The explanations of variable abbreviations are given in Table 4.

Variable	Dimension	n	Mean	Std dev	Median	Min	Max	STM (2001)
TEMP	º C	530	7,6	2,9	**7,0**	2,3	19,0	--
COLO	mgl^{-1} Pt	662	8	24	**5**	0	500	≤5
TURB	FNU	519	0,36	0,27	**0,28**	0	0,98	≤1
O2	mgl^{-1}	304	4,4	3,4	**3,4**	0	12,9	--
CO2	mgl^{-1}	234	19,4	24,7	**9,0**	0	160,0	--
PH	--	844	7,1	0,8	**7,2**	5,2	9,0	6,5...9,5
COND	mSm^{-1} 25º C	831	23,5	27,7	**18,4**	2,6	380,0	<250
ALK	$mmoll^{-1}$	517	1,31	0,82	**1,30**	0,04	6,20	--
HCO3	mgl^{-1}	517	80,2	50,2	**79,3**	2,4	378,3	--
HARD	$mmoll^{-1}$	618	0,75	0,74	**0,58**	0,06	9,20	--
COD	mgl^{-1}	1167	1,4	2,1	**0,8**	0,08	29,1	≤5
KMNO4	mgl^{-1}	1167	5,4	8,4	**3,0**	0,3	115,0	≤20
TOC	mgl^{-1}	138	1,1	1,0	**0,8**	0,4	8,0	--
NO3	mgl^{-1}	879	5,3	10,8	**1,0**	0	130,0	≤50
NO2	mgl^{-1}	690	0,02	0,18	**0,006**	0	4,52	≤0,5
NH4	mgl^{-1}	592	0,04	0,16	**0,008**	0	2,65	≤0,5
TOTP	mgl^{-1}	295	0,02	0,03	**0,006**	0	0,23	--
PO4P	mgl^{-1}	80	0,02	0,06	**0,006**	0,001	0,48	--
S	mgl^{-1}	6	3,9	3,2	**3,0**	1,4	10,0	--
SO4	mgl^{-1}	539	11,5	12,4	**8,8**	0,4	204,0	≤250
CL	mgl^{-1}	685	27,6	117,8	**6,8**	0,6	1800,0	≤100
BR	mgl^{-1}	86	0,3	0,9	**0,1**	0,009	6,8	--
F	mgl^{-1}	596	0,81	0,85	**0,49**	0,04	5,50	≤1,5
SIO2	mgl^{-1}	369	16,5	4,9	**16,1**	4,7	33,0	--
CA	mgl^{-1}	476	20,7	19,6	**16,1**	1,8	207,0	--
MG	mgl^{-1}	470	6,3	6,4	**4,7**	0,3	74,1	--
NA	mgl^{-1}	487	19,3	35,2	**9,9**	1,4	400,0	--
K	mgl^{-1}	473	3,1	4,7	**2,0**	0,3	71,0	--
AG	μgl^{-1}	87	0,01	0,003	**0,01**	0	0,03	--
AL	μgl^{-1}	263	18	20	**20**	0	160	≤200
AS	μgl^{-1}	224	5	21	**<2**	1	294	≤10
B	μgl^{-1}	115	29	31	**20**	0	213	≤1000
BA	μgl^{-1}	91	25	65	**10**	1	575	--
BE	μgl^{-1}	87	0,12	0,06	**0,1**	0,1	0,3	--
BI	μgl^{-1}	87	0,03	0,003	**0,03**	0,02	0,03	--
CD	μgl^{-1}	320	<1	<1	**<1**	<1	3	≤5
CO	μgl^{-1}	133	<1	<1	**<1**	0	5	--
CR	μgl^{-1}	253	<1	1	**<1**	<1	3	≤50
CU	μgl^{-1}	319	34	60	**10**	<1	400	≤2000
FE	μgl^{-1}	1227	639	2768	**83**	0	47000	<400
HG	μgl^{-1}	24	0,4	0,2	**0,5**	<0,1	0,5	≤1
I	μgl^{-1}	6	4,3	1,4	**4,4**	2,3	6,4	--
LI	μgl^{-1}	36	7,9	9,5	**6,2**	0,6	50,7	--
MN	μgl^{-1}	1156	153	338	**40**	0	4100	<100
MO	μgl^{-1}	87	4	6	**<2**	<1	44	--
NI	μgl^{-1}	319	3	3	**<3**	<1	26	≤20
PB	μgl^{-1}	319	<1	0,8	**<1**	<1	11	≤10
RB	μgl^{-1}	87	4	5	**<2**	<1	34	--
SB	μgl^{-1}	111	<1	0,4	**<1**	<1	3	≤5
SE	μgl^{-1}	111	<1	0,6	**<1**	<1	5	≤10
SN	μgl^{-1}	6	<0,5	0	**<0,5**	<0,5	<0,5	--
SR	μgl^{-1}	91	151	248	**92**	20	1770	--
TH	μgl^{-1}	87	0,02	0,003	**0,02**	0,01	0,03	--
TL	μgl^{-1}	87	0,03	0,02	**0,02**	0,01	0,12	--
U	μgl^{-1}	330	27	141	**3**	<1	2400	--
V	μgl^{-1}	87	<1	<1	**<1**	<1	10	--
ZN	μgl^{-1}	321	115	382	**29**	<1	5500	--
RA	Bql^{-1}	80	0,02	0,02	**0,01**	0	0,13	--
RN	Bql^{-1}	582	435	770	**190**	5	8140	<1000
TA	Bql^{-1}	118	0,6	1,2	**0,2**	0	7,8	--

Table 19. Water quality judgment of drilled wells in the CF database according to the Decree of the Ministry of Social Affairs and Health relating to the quality and monitoring of water intended for human consumption in single households (STM 2001; Table 18). Turbidity <1 FNU if determined; total number of sampled wells n=2007 (80% of the CF database wells); test wells included. The percent of good quality wells is in bold. The explanations of different abbreviations are given in Table 4.

Variable	Sampled wells		Good quality		Poor quality	
	n	%	n	%	n	%
COLO	662	33,0	517	**78,1**	145	21,9
PH	844	42,1	637	**75,5**	207	24,5
COND	831	41,4	828	**99,6**	3	0,4
COD	1167	58,1	1115	**95,5**	52	4,5
KMNO4	1167	58,1	1115	**95,5**	52	4,5
NO3	879	43,8	874	**99,4**	5	0,6
NO2	690	34,4	687	**99,6**	3	0,4
NH4	592	29,5	583	**98,5**	9	1,5
SO4	539	26,9	539	**100,0**	0	0,0
CL	685	34,1	663	**96,8**	22	3,2
F	596	29,7	493	**82,7**	103	17,3
AL	263	13,1	263	**100,0**	0	0,0
AS	224	11,2	215	**96,0**	9	4,0
B	115	5,7	115	**100,0**	0	0,0
CD	320	15,9	320	**100,0**	0	0,0
CR	253	12,6	253	**100,0**	0	0,0
CU	319	15,9	319	**100,0**	0	0,0
FE	1227	61,1	1017	**82,9**	210	17,1
HG	24	1,2	24	**100,0**	0	0,0
MN	1156	57,6	777	**67,2**	379	32,8
NI	319	15,9	317	**99,4**	2	0,6
PB	319	15,9	318	**99,7**	1	0,3
SB	111	5,5	111	**100,0**	0	0,0
SE	111	5,5	111	**100,0**	0	0,0
RN	582	29,0	525	**90,2**	57	9,8
BACT	420	20,9	359	**85,5**	61	14,5

7.2 Hydraulic properties of drilled wells

7.2.1 Normalized yield

The median normalized yield (Q/d_s) of private drilled wells in Central Finland is 12 $Lhr^{-1}m^{-1}$ (Table 20). The Q/d_s-value for high-yield wells is ≥102 $Lhr^{-1}m^{-1}$ (P90%) and that for low-yield wells ≤1 $Lhr^{-1}m^{-1}$ (P10%). Q/d_s-values range nearly six orders of magnitude. The median Q/d_s-value of test wells is 2,3 times higher than that of private wells. The median normalized well yield defined as yield per drilled rock depth (Q/d) is 11 $Lhr^{-1}m^{-1}$ for all wells in the CF database. The yield of a dry well was entered into calculations as a nominal figure of 1 Lhr^{-1}.

In the Gothenburg crystalline rock region, southwestern Sweden, the average normalized yields defined as yield per drilled rock depth (Q/d) are around 10 $Lhr^{-1}m^{-1}$ (Wladis & Rosenbaum 1994, 1995, Wladis 1995). In Norway, the corresponding median normalized yield (Q/d) of drilled wells is 12 $Lhr^{-1}m^{-1}$ (n=12757; Morland 1997, Banks et al. 2005). In the crystalline bedrock of Sunnfjord region, Norway, the median Q/d of boreholes is somewhat lower, i.e. 7,1 $Lhr^{-1}m^{-1}$; the mean is 16,6 $Lhr^{-1}m^{-1}$ (n=373; Braathen et al. 1999).

Table 20. The distribution of normalized yield (Q/d_s) of drilled wells in the CF database. Median values are in bold.

Parameter or quantile	Private wells Q/d_s n=1227 94%		Test wells Q/d_s n=73 6%		All wells Q/d_s n=1300 100%	
	$Lhr^{-1}m^{-1}$	m^2s^{-1}	$Lhr^{-1}m^{-1}$	m^2s^{-1}	$Lhr^{-1}m^{-1}$	m^2s^{-1}
Mean	46	$1,27x10^{-5}$	51	$1,43x10^{-5}$	46	$1,28x10^{-5}$
Std dev	165	$4,59x10^{-5}$	58	$1,60x10^{-5}$	161	$4,47x10^{-5}$
Mode	13	$3,70x10^{-6}$	33	$9,26x10^{-6}$	13	$3,70x10^{-6}$
Range	3500	$9,72x10^{-4}$	208	$5,79x10^{-5}$	3500	$9,72x10^{-4}$
100% Max	3500	$9,72x10^{-4}$	208	$5,79x10^{-5}$	3500	$9,72x10^{-4}$
99%	400	$1,11x10^{-4}$	208	$5,79x10^{-5}$	400	$1,11x10^{-4}$
95%	188	$5,21x10^{-5}$	159	$4,42x10^{-5}$	180	$4,99x10^{-5}$
90%	102	$2,82x10^{-5}$	145	$4,03x10^{-5}$	109	$3,03x10^{-5}$
75% Q_3	35	$9,75x10^{-6}$	73	$2,03x10^{-5}$	37	$1,04x10^{-5}$
50% Md	**12**	**$3,23x10^{-6}$**	**28**	**$7,87x10^{-6}$**	**12**	**$3,31x10^{-6}$**
25% Q_1	4	$9,92x10^{-7}$	5	$1,30x10^{-6}$	4	$9,96x10^{-7}$
10%	1	$2,81x10^{-7}$	0,6	$1,72x10^{-6}$	1	$2,78x10^{-7}$
5%	0,3	$7,98x10^{-8}$	0,07	$1,88x10^{-6}$	0,3	$7,84x10^{-8}$
1%	0,008	$2,28x10^{-9}$	0,01	$4,15x10^{-9}$	0,008	$2,28x10^{-9}$
0% Min	0,004	$1,18x10^{-9}$	0,01	$4,15x10^{-9}$	0,004	$1,18x10^{-9}$

7.2.2 Specific capacity and well productivity

The specific capacities (Q/s_m) of drilled wells determined by short-term pumping tests are shown in Table 21. The specific capacity of a dry well has been defined as 0,1 $Lhr^{-1}m^{-1}$. The total number of wells is 64, half of which are private wells and the other half test wells. The median yield of the wells is 900 Lhr^{-1}, which is slightly higher than that of all wells in the CF database (700 Lhr^{-1}). The Q/s_m ranges more than five orders of magnitude. The Q/s_m-values decrease with well depth. However, the relation is statistically not significant.

Table 21. Results from short-term pumping tests to determine specific capacities (Q/s_m) of drilled wells in the CF database (n=64). *) The specific capacity of a dry well has been defined as 0,1 $Lhr^{-1}m^{-1}$.

PRIVATE WELLS n=32				TEST WELLS n=32			
Well id	Pumping rate Q Lhr^{-1}	Drawdown s_m m	Q/s_m $Lhr^{-1}m^{-1}$	Well id	Pumping rate Q Lhr^{-1}	Drawdown s_m m	Q/s_m $Lhr^{-1}m^{-1}$
077004	12000	1,00	12000	077077	70	10,50	7
172093	10420	50,00	208	077080	920	10,00	92
179003	1200	2,00	600	077081	2470	5,85	422
179004	200	7,00	29	172070	870	8,00	109
179018	1200	17,00	71	179037	20000	1,18	16900
179056	100	25,00	4	179327	8	9,00	1
179106	12000	1,50	8000	179330	4800	3,43	1400
179112	380	1,50	253	216019	1000	8,00	125
179141	450	50,00	9	216037	1050	1,00	1050
179335	90	18,40	5	216043	1110	27,65	40
182204	--	--	0,1*)	249037	9500	10,46	908
249115	240	3,00	80	275054	800	6,15	130
275020	1000	12,00	83	275055	--	--	0,1*)
410065	190	35,00	5	275056	100	11,80	8
410071	2200	9,70	227	275058	8	9,00	1
410072	4500	24,50	184	275059	3000	2,00	1500
410073	550	44,00	13	275061	10	7,50	1
410074	1300	29,50	44	291093	2400	3,18	754
410075	750	35,00	21	291096	1680	5,60	300
410079	2500	70,00	36	435103	3050	29,40	104
410081	780	15,00	52	592130	500	16,70	30
410082	2000	16,70	120	729148	8400	5,53	1520
410083	2000	24,00	83	850036	1020	21,80	47
410084	600	34,50	17	892018	40	10,00	4
410086	400	34,50	12	931023	460	9,00	51
592011	5300	28,00	189	931024	1080	4,46	242
592029	260	68,00	4	931132	77	38,06	2
592030	150	30,00	5	931150	3200	2,36	1354
601031	200	4,00	50	931161	6010	19,70	305
729128	1000	46,00	22	931162	200	100,00	2
931124	1600	3,00	533	931163	1670	2,11	790
931125	600	7,00	86	931164	3855	9,80	393

Regression between specific capacity (Q/s_m) and normalized yield (Q/d_s) was first calculated separately for private wells and test wells. Because the regressions were almost identical, the final regression was based on both private and test wells. The coefficient of determination, R^2, for log-transformed values was more than twice of that for non-transformed values supporting a log-log relationship between the variables (Fig. 51).

The data points are scattered about the regression line within the entire range of observations. The regression equation can be written as

$$\log(Q/s_m) = 0{,}545 + 0{,}986 \log(Q/d_s) \tag{53}$$

where Q/d_s is expressed in m^2s^{-1}. The equation (53) explains 78% of the variation in the specific capacity values, which is considered good. The equation (53) can also be written as

$$Q/s_m = 3{,}504\,(Q/d_s)^{0{,}986} \tag{54}$$

where Q/d_s is expressed in m^2s^{-1} and

$$Q/s_m = 4{,}338\,(Q/d_s)^{0{,}986} \tag{55}$$

where Q/d_s is expressed in $Lhr^{-1}m^{-1}$.

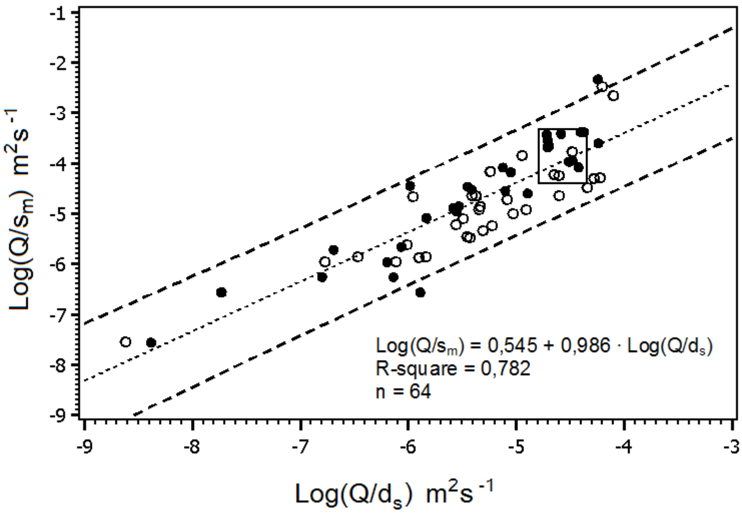


Fig. 51. The log-log regression between specific capacity (Q/s_m) and normalized yield (Q/d_s) with 95% confidence limits for individual predicted values for drilled wells in the CF database (○=private well, ●=test well). The black box includes wells with transmissivity (T_m) determinations (Fig. 52).

The distribution of specific capacities (Q/s) calculated with the equations (54) and (55) for drilled wells in the CF database is shown in Table 22.

The median specific capacity (Q/s) of private drilled wells in Central Finland is 49 $Lhr^{-1}m^{-1}$, i.e. $1{,}4 \times 10^{-5}$ m^2s^{-1}. The Q/s-value for high-yield wells is ≥413 $Lhr^{-1}m^{-1}$ (P90%) and that for low-yield wells ≤4 $Lhr^{-1}m^{-1}$ (P10%). Specific capacity ranges more than five orders of magnitude. The median Q/s-value of test wells is 2,4 times higher than that of private wells.

Specific capacities (Q/s) are clearly higher than the corresponding normalized well yields (Q/d_s or Q/d). For example, for a 105 meters deep well with saturated open section of 100 m and with a yield of 1,000 Lhr^{-1}, the normalized well yield is 10 $Lhr^{-1}m^{-1}$ and the corresponding specific capacity is 42 $Lhr^{-1}m^{-1}$, i.e. Q/s is 4,2 times the value of Q/d_s. This is predominantly because the actual drawdown per a unit well yield is generally much less than the available drawdown based on the saturated open well section or on the drilled rock depth. The reason behind this might be, for example, that air-lifting at the end of the DTH drilling partly prevents water to enter the borehole. Occasionally also the reopening of the water-bearing fractures may increase the well yield afterwards compared to that of the drilling-time well yield. Of course, drilling deep wells in intact rock may significantly decrease normalized well yields.

Specific capacities determined in other parts of the world coincide well with the results of this study. The average specific capacity of the wells drilled in metamorphic rocks of

Colorado Front Range, USA, is ca. 6,4x10^{-5} m^2s^{-1} (n=37, Snow 1968). In crystalline rock area of Rothschild in Wisconsin, USA, Summers (1972) reports specific capacity values ranging between 15 and 450 $Lhr^{-1}m^{-1}$. According to Daniel (1987, 1989), the median specific capacity of wells drilled in crystalline rocks of the Piedmont and Blue Ridge Provinces of North Carolina, USA, is 60 $Lhr^{-1}m^{-1}$ (n=5221). The average specific capacity of wells drilled in crystalline bedrock of Georgia, USA, is 570 $Lhr^{-1}m^{-1}$ (n=257; Brook 1988). In fractured metamorphic and igneous rocks in Pennsylvania, USA, specific capacity values normally range between 110 and 1120 $Lhr^{-1}m^{-1}$, but values as low as 8 $Lhr^{-1}m^{-1}$ have been recorded (n=4391; Knopman & Hollyday 1993).

Table 22. The distribution of specific capacity (Q/s) of drilled wells in the CF database. Median values are in bold.

Parameter or quantile	Private wells Q/s n=1227 94%		Test wells Q/s n=73 6%		All wells Q/s n=1300 100%	
	$Lhr^{-1}m^{-1}$	m^2s^{-1}	$Lhr^{-1}m^{-1}$	m^2s^{-1}	$Lhr^{-1}m^{-1}$	m^2s^{-1}
Mean	184	5,12x10^{-5}	209	5,81x10^{-5}	186	5,16x10^{-5}
Std dev	644	1,79x10^{-4}	232	6,45x10^{-5}	628	1,74x10^{-4}
Mode	56	1,50x10^{-5}	138	3,80x10^{-5}	56	1,50x10^{-5}
Range	13530	3,76x10^{-3}	838	2,33x10^{-4}	13530	3,76x10^{-3}
100% Max	13530	3,76x10^{-3}	838	2,33x10^{-4}	13530	3,76x10^{-3}
99%	1594	4,43x10^{-4}	838	2,33x10^{-4}	1594	4,43x10^{-4}
95%	755	2,10x10^{-4}	642	1,78x10^{-4}	724	2,01x10^{-4}
90%	413	1,15x10^{-4}	587	1,63x10^{-4}	443	1,23x10^{-4}
75% Q_3	145	4,02x10^{-5}	298	8,28x10^{-5}	154	4,27x10^{-5}
50% Md	**49**	**1,35x10^{-5}**	**117**	**3,26x10^{-5}**	**50**	**1,39x10^{-5}**
25% Q_1	15	4,23x10^{-6}	20	5,50x10^{-6}	15	4,24x10^{-6}
10%	4	1,22x10^{-6}	3	7,52x10^{-7}	4	1,21x10^{-6}
5%	1	3,52x10^{-7}	0,3	8,46x10^{-8}	1	3,46x10^{-7}
1%	0,04	1,06x10^{-8}	0,07	1,91x10^{-8}	0,04	1,06x10^{-8}
0% Min	0,02	5,54x10^{-9}	0,07	1,91x10^{-8}	0,02	5,54x10^{-9}

In Manitoba, Canada, the specific capacities of wells drilled in Precambrian igneous and metamorphic rocks range mostly from 3 to 36 $Lhr^{-1}m^{-1}$ (n=196; Betcher et al. 1995). In the Precambrian terrain of eastern Ontario, Canada, metasedimentary and metavolcanic rocks (n=383) and intrusive rocks (n=749) have a mean specific capacity of 2,3x10^{-5} m^2s^{-1} and 3,0x10^{-5} m^2s^{-1}, respectively (Gleeson & Novakowski 2009).

Median specific capacity of boreholes in weathered crystalline basement aquifers in Zimbabwe, Nigeria and Malawi varies from 1,9x10^{-5} to 1,0x10^{-4} m^2s^{-1} (Chilton & Foster 1995). The specific capacity in fractured-bedrock aquifers of Uganda varies from 1,2x10^{-5} to 2,0x10^{-4} m^2s^{-1}; the mean Q/s is 8,0x10^{-5} m^2s^{-1} (Taylor & Howard 2000). In the weathered crystalline and metamorphic bedrock aquifers of Ivory Coast the specific capacity ranges from 8 to 1185 $Lhr^{-1}m^{-1}$; the average is 213 $Lhr^{-1}m^{-1}$ (Razack & Lasm 2006). In weathered basement aquifers of Eastern Senegal the specific capacity is between 10-4,000 $Lhr^{-1}m^{-1}$ (n=177; Diop & Tijani 2008).

Near São Paulo, Brazil, the specific capacity of drilled wells in crystalline and sedimentary rocks ranges from 2,8x10^{-7} to 4,4x10^{-3} m^2s^{-1}, with a median of 3,3x10^{-5} m^2s^{-1} (Fernandes & Rudolph 2001). In another study in the São Paulo crystalline rock region the median specific capacity of 312 wells is 100 $Lhr^{-1}m^{-1}$ ($\bar{x}$=280 $Lhr^{-1}m^{-1}$; Yoshinaga Pereira & Kimmelmann e Silva 2004). The specific capacity of drilled wells (n~15000) in the crystalline basement of northeastern Brazil is distributed as follows: < 100 $Lhr^{-1}m^{-1}$ 34%, 100-500 $Lhr^{-1}m^{-1}$ 34%, 500-1,000 $Lhr^{-1}m^{-1}$ 13% and >1,000 $Lhr^{-1}m^{-1}$ 13% (Rebouças 1999a).

Well losses were determined in one private well and in six test wells. The well loss results are shown in Table 23. The well losses depend on pumping discharges and they range from 20% to 95% of total drawdown. The corresponding well efficiencies vary from 80% to 5%. This suggests that limited fracturing in the vicinity of the boreholes may severely impede the movement of water to the wells (e.g. Hahn 1977). The mean and median well losses are 55% and 42%, respectively. The well loss and well efficiency percentages are in a similar range than in crystalline rock areas of the Delaware Piedmont, USA (Hahn 1977, Talley & Hahn 1978), and in southwestern Nigeria (Izinyon & Anyata 2006).

Table 23. Well loss determinations for drilled wells in the CF database (n=7).

Well id	Step-test			Driller´s well yield Lhr^{-1}	Well loss %	Well efficiency %
	Pumping rate Q Lhr^{-1}	s_w m	Well loss %			
077004	1380 12000 18000	0,08 1,00 2,00	7,8 42,1 52,1	12000	42	58
172070	1560 1680 2040 2280	20,36 23,16 30,16 34,16	30,8 32,4 36,8 39,4	1500	30	70
179335	45 90	8,25 18,40	11,5 20,7	90	21	79
249037	8100 9660 10560	10,26 13,77 17,24	93,1 94,1 94,6	10000	94	6
291093	2420 6000	3,20 10,22	19,4 37,3	5500	35	65
729148	2940 4320	1,25 2,28	51,3 60,9	8700	76	24
850036	520 830 1000 1060	5,96 14,44 21,64 21,82	81,9 88,0 89,7 90,3	800	88	12
Mean Median					55 42	45 58

The median well productivity (Q_w) of private drilled wells in Central Finland is $2{,}1 \times 10^{-7}$ ms^{-1}, i.e. 0,02 md^{-1} (Table 24). The Q_w-value for high-yield wells is $\geq 3{,}7 \times 10^{-6}$ ms^{-1} (P90%) and that for low-yield wells $\leq 9{,}6 \times 10^{-9}$ ms^{-1} (P10%). Well productivities range nearly eight orders of magnitude. The median Q_w-value of test wells is 1,7 times higher than that of private wells.

Table 24. The distribution of well productivity (Q_w) of drilled wells in the CF database. Median values are in bold.

Parameter or quantile	Private wells Q_w n=1227 94%		Test wells Q_w n=73 6%		All wells Q_w n=1300 100%	
	md^{-1}	ms^{-1}	md^{-1}	ms^{-1}	md^{-1}	ms^{-1}
Mean	0,45	$5{,}17 \times 10^{-6}$	0,10	$1{,}14 \times 10^{-6}$	0,43	$4{,}95 \times 10^{-6}$
Std dev	6,20	$7{,}17 \times 10^{-5}$	0,14	$1{,}60 \times 10^{-6}$	6,02	$6{,}97 \times 10^{-5}$
Mode	0,76	$8{,}79 \times 10^{-6}$	0,09	$1{,}03 \times 10^{-6}$	0,76	$8{,}79 \times 10^{-6}$
Range	162,36	$1{,}88 \times 10^{-3}$	0,67	$7{,}78 \times 10^{-6}$	162,36	$1{,}88 \times 10^{-3}$
100% Max	162,36	$1{,}88 \times 10^{-3}$	0,67	$7{,}78 \times 10^{-6}$	162,36	$1{,}88 \times 10^{-3}$
99%	3,64	$4{,}21 \times 10^{-5}$	0,67	$7{,}78 \times 10^{-6}$	3,48	$4{,}03 \times 10^{-5}$
95%	0,76	$8{,}79 \times 10^{-6}$	0,42	$4{,}85 \times 10^{-6}$	0,76	$8{,}79 \times 10^{-6}$
90%	0,32	$3{,}65 \times 10^{-6}$	0,25	$2{,}85 \times 10^{-6}$	0,31	$3{,}64 \times 10^{-6}$
75% Q_3	0,08	$8{,}88 \times 10^{-7}$	0,14	$1{,}58 \times 10^{-6}$	0,08	$9{,}59 \times 10^{-7}$
50% Md	**0,02**	$\mathbf{2{,}05 \times 10^{-7}}$	**0,03**	$\mathbf{3{,}54 \times 10^{-7}}$	**0,02**	$\mathbf{2{,}08 \times 10^{-7}}$
25% Q_1	0,004	$4{,}46 \times 10^{-8}$	0,005	$5{,}54 \times 10^{-8}$	0,004	$4{,}46 \times 10^{-8}$
10%	0,0008	$9{,}64 \times 10^{-9}$	0,0007	$8{,}02 \times 10^{-9}$	0,0008	$9{,}50 \times 10^{-9}$
5%	0,0002	$2{,}28 \times 10^{-9}$	0,0001	$1{,}16 \times 10^{-9}$	0,0002	$2{,}24 \times 10^{-9}$
1%	0,000007	$8{,}66 \times 10^{-11}$	0,00002	$2{,}85 \times 10^{-10}$	0,000007	$8{,}66 \times 10^{-11}$
0% Min	0,000002	$2{,}36 \times 10^{-11}$	0,00002	$2{,}85 \times 10^{-10}$	0,000002	$2{,}36 \times 10^{-11}$

7.2.3 Transmissivity

Transmissivity (T_m) was determined in one private well and in nine test wells with the Cooper-Jacob single-well straight-line method for unsteady-state flow. Specific capacities (Q/s_m) determined by short-term pumping tests for the same wells were corrected for well loss with the approximate average well loss 50% as $Q/s_c=(Q/s_m) \times 2$. The results are shown in Table 25. Q/s_c-values are 3,5-5,0 times higher than the corresponding transmissivities. The median yield of the wells is 7,000 Lhr^{-1}, which is one order of magnitude higher than that of all drilled wells in the CF database. However, the wells fit the general regression line in Figure 51.

The regression between T_m and Q/s_c was calculated. The coefficient of determination, R^2, for log-transformed values was almost twice of that for non-transformed values supporting a log-log relationship between the variables (e.g. Razack & Huntley 1991). The regression equation between transmissivity T_m and specific capacity corrected for well loss Q/s_c is

$$\log T_m = -0{,}572 + 0{,}986 \log(Q/s_c) \qquad (56)$$

where Q/s_c is expressed in m^2d^{-1}.

The data points are scattered about the regression line within one and a half log-cycle range of observations (Fig. 52). The regression equation explains 80% of the variation in transmissivity values, which is considered good. The estimated line slope close to 1,0 indicates that there is no correlation between Q/s and well efficiency (Christensen 1995).

The equation (56) can also be written as

$$T_m = 0{,}268\,(Q/s_c)^{0{,}986} \qquad (57)$$

where Q/s_c is expressed in m^2d^{-1}.

Table 25. Results from the transmissivity determinations of drilled wells in the CF database (n=10).

Well identity number	Driller´s well yield Lhr^{-1}	Diameter of well mm	Pumping rate Q m^3d^{-1}	Drawdown difference per log cycle of time Δs_w m	Time condition $t > 25r_c^2/T$ min	Saturated open well section m	Transmissivity $T_m=2,30Q/4\pi\Delta s_w$		Specific capacity corrected for well loss $Q/s_c=(Q/s_m)*2$	
							m^2d^{-1}	m^2s^{-1}	m^2d^{-1}	m^2s^{-1}
077080	3600	140	30	3,40	>109	78	1,6	$1,88x10^{-5}$	4,4	$5,11x10^{-5}$
179330	8000	140	115	1,18	> 9	85	17,9	$2,07x10^{-4}$	67,2	$7,78x10^{-4}$
216037	5500	140	184	1,81	> 10	78	18,6	$2,16x10^{-4}$	50,4	$5,83x10^{-4}$
249037	10000	160	190	4,98	>29	48	7,0	$8,09x10^{-5}$	43,6	$5,04x10^{-4}$
275060	5000	140	238	13,72	> 56	108	3,2	$3,67x10^{-5}$	13,4	$1,56x10^{-4}$
729148	8700	140	97	1,58	>15	61	11,3	$1,30x10^{-4}$	73,0	$8,44x10^{-4}$
931009	6000	150	44	6,60	>166	155	1,2	$1,41x10^{-5}$	7,9	$9,11x10^{-5}$
931161	9000	140	138	6,15	>42	66	4,1	$4,76x10^{-5}$	14,6	$1,69x10^{-4}$
931163	4600	140	127	0,95	>18	63	24,4	$2,83x10^{-4}$	37,9	$4,39x10^{-4}$
931164	9000	140	125	5,85	>45	81	3,9	$4,52x10^{-5}$	18,9	$2,18x10^{-4}$
Mean	6940	143	129	4,62	>51	82	9,3	$1,08x10^{-4}$	33,1	$3,83x10^{-4}$
Median	7000	140	126	4,19	>36	78	5,6	$6,42x10^{-5}$	28,4	$3,29x10^{-4}$

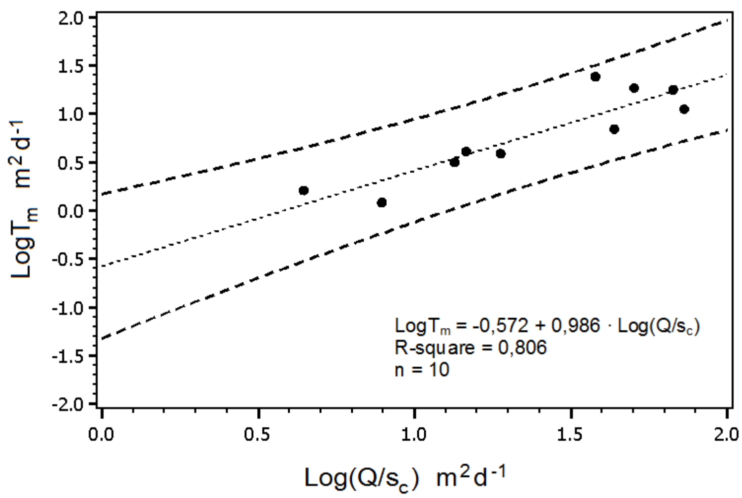


Fig. 52. The log-log regression between transmissivity T_m and specific capacity corrected for well loss Q/s_c with 95% confidence limits for individual predicted values for drilled wells in the CF database (n=10).

Transmissivities (T) for drilled wells in the CF database were determined on the basis of this regression equation (57). The specific capacities Q/s of the database wells (Table 22) were first corrected for well loss by multiplying them with 2.

The median transmissivity (T) of private drilled wells in Central Finland is $7{,}2 \times 10^{-6}$ m^2s^{-1}, i.e. 0,6 m^2d^{-1} (Table 26). The T-value for high-yield wells is $\geq 5{,}9 \times 10^{-5}$ m^2s^{-1} (P90%) and that for low-yield wells $\leq 6{,}7 \times 10^{-7}$ m^2s^{-1} (P10%). Transmissivities range nearly six orders of magnitude. The median T-value of test wells is 2,4 times higher than that of private wells. The highest T-values in the data set ($> 1 \times 10^{-3}$ m^2s^{-1}) are of the same order of magnitude as those given by Leveinen (1996, 2000) for a high-yield drilled well at the Pohjukansalo water work in Leppävirta, east-central Finland (T=$1{,}3 \ldots 6{,}4 \times 10^{-3}$ m^2s^{-1}, Q=30,000 Lhr^{-1}).

The transmissivities of drilled wells in the CF database range for most part ($\bar{x} \pm s$) from very low (V) to low (IV) with a very large standard deviation (s>0,8) in Krásný´s (1993a) classification of transmissivity magnitude and variation (Fig. 53). The spatial distribution of transmissivities of the CF database wells is presented in Fig. 54.

Table 26. The distribution of transmissivity (T) of drilled wells in the CF database. Median values are in bold.

Parameter or quantile	Private wells T n=1227 94%		Test wells T n=73 6%		All wells T n=1300 100%	
	m^2d^{-1}	m^2s^{-1}	m^2d^{-1}	m^2s^{-1}	m^2d^{-1}	m^2s^{-1}
Mean	2,26	$2{,}61 \times 10^{-5}$	2,58	$2{,}99 \times 10^{-5}$	2,28	$2{,}64 \times 10^{-5}$
Std dev	7,63	$8{,}83 \times 10^{-5}$	2,84	$3{,}29 \times 10^{-5}$	7,44	$8{,}62 \times 10^{-5}$
Mode	0,71	$8{,}18 \times 10^{-6}$	1,72	$2{,}00 \times 10^{-5}$	0,71	$8{,}18 \times 10^{-6}$
Range	159,14	$1{,}84 \times 10^{-3}$	10,24	$1{,}19 \times 10^{-4}$	159,14	$1{,}84 \times 10^{-3}$
100% Max	159,14	$1{,}84 \times 10^{-3}$	10,24	$1{,}19 \times 10^{-4}$	159,14	$1{,}84 \times 10^{-3}$
99%	19,31	$2{,}23 \times 10^{-4}$	10,24	$1{,}19 \times 10^{-4}$	19,31	$2{,}23 \times 10^{-4}$
95%	9,24	$1{,}07 \times 10^{-4}$	7,88	$9{,}12 \times 10^{-5}$	8,86	$1{,}03 \times 10^{-4}$
90%	5,10	$5{,}90 \times 10^{-5}$	7,21	$8{,}34 \times 10^{-5}$	5,47	$6{,}33 \times 10^{-5}$
75% Q_3	1,81	$2{,}10 \times 10^{-5}$	3,69	$4{,}28 \times 10^{-5}$	1,92	$2{,}23 \times 10^{-5}$
50% Md	**0,62**	**$7{,}16 \times 10^{-6}$**	**1,47**	**$1{,}70 \times 10^{-5}$**	**0,63**	**$7{,}34 \times 10^{-6}$**
25% Q_1	0,20	$2{,}27 \times 10^{-6}$	0,25	$2{,}95 \times 10^{-6}$	0,20	$2{,}28 \times 10^{-6}$
10%	0,06	$6{,}66 \times 10^{-7}$	0,04	$4{,}14 \times 10^{-7}$	0,06	$6{,}59 \times 10^{-7}$
5%	0,02	$1{,}96 \times 10^{-7}$	0,004	$4{,}80 \times 10^{-8}$	0,02	$1{,}93 \times 10^{-7}$
1%	0,0005	$6{,}17 \times 10^{-9}$	0,001	$1{,}10 \times 10^{-8}$	0,0005	$6{,}17 \times 10^{-9}$
0% Min	0,0003	$3{,}26 \times 10^{-9}$	0,001	$1{,}10 \times 10^{-8}$	0,0003	$3{,}26 \times 10^{-9}$

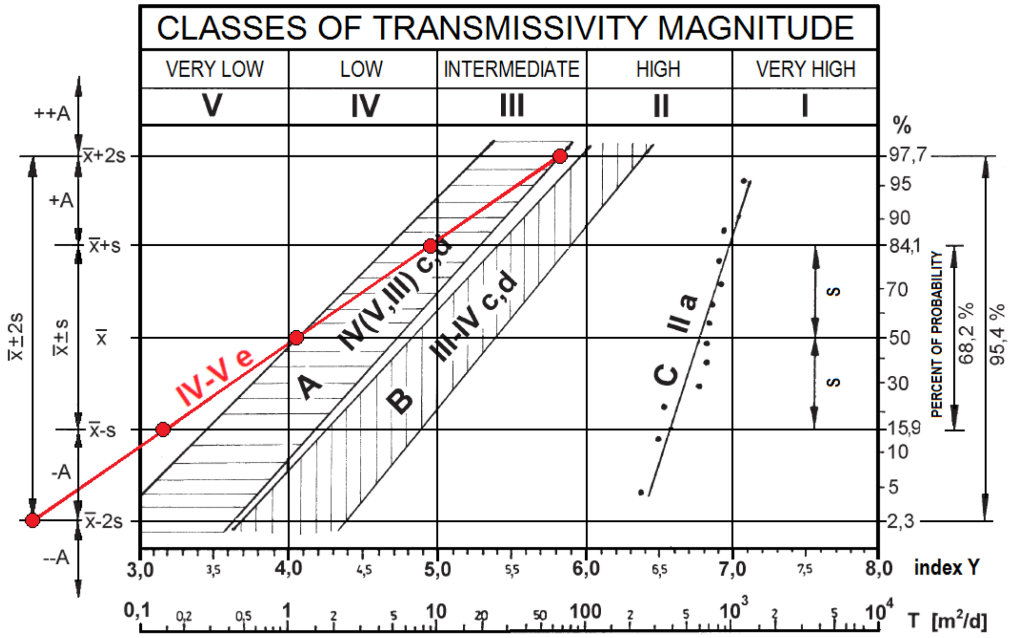


Fig. 53. The transmissivity distribution of drilled wells in the CF database (●—●) on Krásný´s (1993a) classification of transmissivity magnitude and variation (n=1300, test wells included). Index Y was calculated as $Y = \log_{10}(10^6 \cdot Q/s)$, where Q/s is expressed in $Ls^{-1}m^{-1}$. $\bar{x}$ = arithmetic mean, s = standard deviation, $\bar{x}\pm s$ = interval of prevailing transmissivity (68,2% of observations), $\bar{x}\pm 2s$ = 95,4% of observations, +A = positive anomalies, –A = negative anomalies, ++A = extreme positive anomalies, --A = extreme negative anomalies, **A-C** = background transmissivity values from the Czech Republic (**A** = majority of hard rock types, **B** = crystalline limestones, **C** = fluvial deposits). Modified from Krásný (2002).

The transmissivity values in Central Finland are similar to those of other crystalline rock areas. The average transmissivity values in gneiss and granite areas of Sweden range from $2,8x10^{-6}$ to $3,6x10^{-5}$ m^2s^{-1} (n=194, Carlsson & Carlstedt 1977). In Äspö crystalline rock area, Sweden, the geometric mean of transmissivity is $1,4x10^{-5}$ m^2s^{-1} (Rhén et al. 1997). An examination of yield distributions of drilled wells in Fennoscandia allowed Banks et al. (2010) to make an estimate of the bulk transmissivity of the wells as being approximately $6,4x10^{-6}$ m^2s^{-1}, which is almost identical with the present study.

The transmissivities of gneiss and granite rocks of the Black Forest region, Germany, vary from $1,0x10^{-7}$ to $4,7x10^{-3}$ m^2s^{-1} to a depth of 3,5 km (Stober 1997). The median transmissivity of the drilled wells in metamorphic rock area of Maine, USA, is $4,5x10^{-5}$ m^2s^{-1} (n=35; Mabee 1992) and $3,0x10^{-6}$ m^2s^{-1} in the Blue Ridge of Pennsylvania, USA (Low 2004). The mean transmissivities of drilled wells in granite and gneiss areas of Saguenay-Lac-Saint-Jean area near Quebec, Canada, are $8,48x10^{-5}$ and $9,36x10^{-5}$ m^2s^{-1}, respectively (n=382 and 313; Richard et al. 2011).

The transmissivity in fractured-bedrock aquifers of Uganda varies from $3,8x10^{-6}$ to $1,6x10^{-4}$ m^2s^{-1}; the mean is $5,4x10^{-5}$ m^2s^{-1} and the median $1,16x10^{-5}$ m^2s^{-1} (Howard et al. 1992, Taylor & Howard 2000). According to Batte et al. (2008b), the transmissivity of high-yield wells in Kamuli District, eastern Uganda, is greater than $9,26x10^{-5}$ m^2s^{-1}. Test pumping of boreholes tapping crystalline bedrock in Zimbabwe show a wide range of transmissivity ($6,0x10^{-6}$…$1,0x10^{-3}$ m^2s^{-1}), although most values are within the range $2,0x10^{-5}$…$6,0x10^{-5}$ m^2s^{-1} (Wright 1992). The average transmissivity of Precambrian granite rocks in Ghana is around 7,5x10^{-5} m^{2}s^{-1} with a range of 3,4x10^{-6}...1,3x10^{-3} m^{2}s^{-1} (Darko & Krásný 1998, Gyau-Boakye et al. 2008). Transmissivity values in weathered crystalline and metamorphic rocks of Ivory Coast range from 1,11x10^{-6} to 1,14x10^{-5} m^{2}s^{-1}; the average is 4,0x10^{-5} m^{2}s^{-1} (Razack & Lasm 2006). In weathered basement aquifers of Eastern Senegal the transmissivity varies 10^{-6}-10^{-3} m^{2}s^{-1} (n=177; Diop & Tijani 2008).

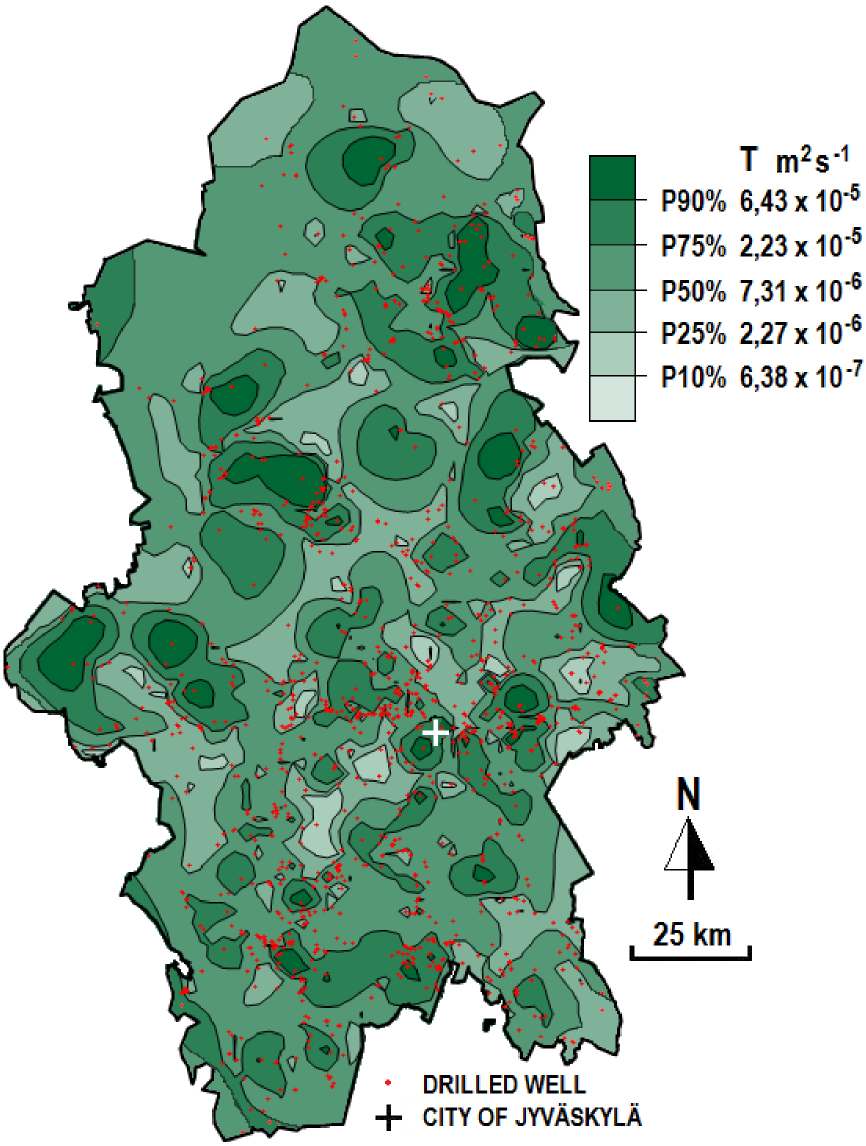


Fig. 54. The spatial distribution of transmissivity (T m^{2}s^{-1}) of drilled wells in the CF database (n=1288, test wells included). Class boundaries: P10% (low-yield wells), P25% (Q_1), P50% (Md), P75% (Q_3), P90% (high-yield wells). Surfer gridding with the natural neighbor method by Kari Illmer/CETECF.

Naik et al. (2001) give transmissivity values from $6,6x10^{-4}$ to $1,1x10^{-3}$ m^2s^{-1} for basalts with variable weathering and jointing in India, whereas transmissivity of a partly weathered granite aquifer ranges from $3,5x10^{-5}$ to $1,9x10^{-3}$ m^2s^{-1} (Subrahmanyam & Yadaiah 2001). Transmissivity in the fracture zones of the Northern Territory of Australia was found to be different in the strike (up to $9,26x10^{-4}$ m^2s^{-1}) to the across-strike direction (up to $1,16x10^{-4}$ m^2s^{-1}; Verma 2003).

Empirical equations between T and Q/s from various studies in hard rock areas were used to calculate T-values for drilled wells in the CF database (Table 27). The T-values given by the equations have a high range in each aquifer type. The median T-value determined with the equation of the present study is from 0,1 to 4,3 times that with other equations in crystalline rock areas. It should be noted that the specific capacities Q/s in different studies have not necessarily been determined in the same way.

Table 27. Mean and median transmissivities for drilled wells in the CF database (n=1300) in ascending order of median values according to various empirical regression equations between T and Q/s (m^2d^{-1}) for fractured crystalline aquifers (C), volcanic aquifers (V), limestone/carbonate/karstic aquifers (L) and sedimentary aquifers (S). WL=Q/s corrected for well loss (Yes/No).

Reference	Aq	Equation (m^2d^{-1})	WL	Transmissivity T for 1300 drilled wells in the CF database			
				Mean		Median	
				m^2d^{-1}	m^2s^{-1}	m^2d^{-1}	m^2s^{-1}
Huntley et al. 1992	C	$T = 0,12(Q/s)^{1,18}$	No	0,92	$1,07x10^{-5}$	0,15	$1,72x10^{-6}$
Choi 1999	V	$T = 0,45(Q/s)^{1,05}$	No	2,31	$2,68x10^{-5}$	0,54	$6,30x10^{-6}$
Sayed & Al-Ruwaih 1995	L	$T = 0,47(Q/s)^{1,13}$	No	3,06	$3,54x10^{-5}$	0,58	$6,67x10^{-6}$
THIS STUDY	**C**	$\mathbf{T = 0,27(Q/s)^{0,99}}$	**Yes**	**2,28**	$\mathbf{2,64x10^{-5}}$	**0,63**	$\mathbf{7,34x10^{-6}}$
Mace 1997	L	$T = 0,76(Q/s)^{1,08}$	No	4,28	$4,95x10^{-5}$	0,92	$1,07x10^{-5}$
Yidana et al. 2008	S	$T = 0,77(Q/s)^{1,08}$	No	4,33	$5,02x10^{-5}$	0,94	$1,08x10^{-5}$
Razack & Lasm 2006	C	$T = 0,33(Q/s)^{1,30}$	Yes	9,45	$1,09x10^{-4}$	1,03	$1,19x10^{-5}$
Fabbri 1997	L	$T = 0,85(Q/s)^{1,07}$	No	4,64	$5,37x10^{-5}$	1,03	$1,19x10^{-5}$
Houston & Lewis 1988	C	$T=0,507(Q/s)+0,424$	No	2,69	$3,12x10^{-5}$	1,04	$1,21x10^{-5}$
Hamm et al. 2005	V	$T = 0,99(Q/s)^{0,89}$	No	3,29	$3,81x10^{-5}$	1,16	$1,35x10^{-5}$
Richard et al. 2011	C	$T = 0,97(Q/s)^{1,08}$	No	5,46	$6,32x10^{-5}$	1,18	$1,37x10^{-5}$
Verbovšek 2008	C	$T = 1,08(Q/s)^{1,07}$	No	5,90	$6,83x10^{-5}$	1,31	$1,52x10^{-5}$
Krásný 1975	C	$T = 1,10(Q/s)$	No	4,90	$5,68x10^{-5}$	1,32	$1,53x10^{-5}$
Banks 1991, 1992b	C	$T = 1,11(Q/s)$	No	4,95	$5,73x10^{-5}$	1,33	$1,54x10^{-5}$
Carlsson & Carlstedt 1977	C	$T = 1,20(Q/s)$	No	5,35	$6,19x10^{-5}$	1,44	$1,66x10^{-5}$
Mace et al. 1999	S	$T = 1,36(Q/s)^{0,84}$	No	4,00	$4,63x10^{-5}$	1,58	$1,83x10^{-5}$
Ericsson & Ronge 1986	C	$T = 1,40(Q/s)$	No	6,24	$7,22x10^{-5}$	1,68	$1,94x10^{-5}$
Banks et al. 2010	C	$T = 1,43(Q/s)$	No	6,38	$7,38x10^{-5}$	1,71	$1,98x10^{-5}$
Rotzoll & El-Kadi 2008	V	$T = 1,54(Q/s)^{1,002}$	No	6,89	$7,98x10^{-5}$	1,84	$2,13x10^{-5}$
El-Naqa 1994	L	$T = 1,81(Q/s)^{0,917}$	No	6,45	$7,47x10^{-5}$	2,14	$2,47x10^{-5}$
Patriarche et al. 2005	S	$T = 2,09(Q/s)^{0,93}$	No	7,70	$8,92x10^{-5}$	2,47	$2,86x10^{-5}$
Wladis & Gustafson 1999	C	$T = 2,22(Q/s)$	No	9,90	$1,15x10^{-4}$	2,66	$3,08x10^{-5}$
Rhén et al. 1997	C	$T = 2,81(Q/s)^{0,98}$	No	11,85	$1,37x10^{-4}$	3,36	$3,88x10^{-5}$
Eagon & Johe 1972	L	$T = 3,24(Q/s)^{0,81}$	Yes	15,56	$1,80x10^{-4}$	6,58	$7,61x10^{-5}$
Jalludin & Razack 2004	V	$T = 3,64(Q/s)^{0,94}$	Yes	26,43	$3,06x10^{-4}$	8,28	$9,58x10^{-5}$

7.2.4 Hydraulic conductivity

Bulk hydraulic conductivities (K) for the wells in the CF database were calculated by dividing their corresponding transmissivity (T) values with the saturated open well sections. The results are shown in Table 28.

The median bulk hydraulic conductivity (K) of private drilled wells is 1,1x10^{-7} ms^{-1}, i.e. 0,01 md^{-1}. The K-value for high-yield wells is ≥1,9x10^{-6} ms^{-1} (P90%) and that for low-yield wells ≤5,3x10^{-9} ms^{-1} (P10%). The hydraulic conductivity values range nearly eight orders of magnitude. The median K-value of test wells is 1,7 times higher than that of private wells. The highest K-values in the data set are of the same order of magnitude as those given by Leveinen et al. (1998) for a high-yield drilled well at the Pohjukansalo water work in Leppävirta, east-central Finland (K=5,0x10^{-6}...7,9x10^{-5} ms^{-1}, Q=30,000 Lhr^{-1}).

Table 28. The distribution of bulk hydraulic conductivity (K) of drilled wells in the CF database. Median values are in bold.

Parameter or quantile	Private wells K n=1227 94%		Test wells K n=73 6%		All wells K n=1300 100%	
	md^{-1}	ms^{-1}	md^{-1}	ms^{-1}	md^{-1}	ms^{-1}
Mean	0,22	2,58x10^{-6}	0,05	5,87x10^{-7}	0,21	2,47x10^{-6}
Std dev	3,04	3,52x10^{-5}	0,07	8,18x10^{-7}	2,96	3,42x10^{-5}
Mode	0,39	4,51x10^{-6}	0,05	5,30x10^{-7}	0,39	4,51x10^{-6}
Range	79,57	9,21x10^{-4}	0,34	3,97x10^{-6}	79,57	9,21x10^{-4}
100% Max	79,57	9,21x10^{-4}	0,34	3,97x10^{-6}	79,57	9,21x10^{-4}
99%	1,86	2,15x10^{-5}	0,34	3,97x10^{-6}	1,77	2,04x10^{-5}
95%	0,39	4,51x10^{-6}	0,21	2,47x10^{-6}	0,39	4,51x10^{-6}
90%	0,16	1,87x10^{-6}	0,13	1,49x10^{-6}	0,16	1,87x10^{-6}
75% Q_3	0,04	4,64x10^{-7}	0,07	8,14x10^{-7}	0,04	5,04x10^{-7}
50% Md	**0,01**	**1,08x10^{-7}**	**0,02**	**1,85x10^{-7}**	**0,01**	**1,10x10^{-7}**
25% Q_1	0,002	2,39x10^{-8}	0,003	2,98x10^{-8}	0,002	2,40x10^{-8}
10%	0,0005	5,27x10^{-9}	0,0004	4,42x10^{-9}	0,0005	5,22x10^{-9}
5%	0,0001	1,26x10^{-9}	0,00006	6,67x10^{-10}	0,0001	1,25x10^{-9}
1%	0,000004	5,06x10^{-11}	0,00001	1,65x10^{-10}	0,000004	5,06x10^{-11}
0% Min	0,000001	1,39x10^{-11}	0,00001	1,65x10^{-10}	0,000001	1,39x10^{-11}

Hydraulic conductivities are in a similar range as in other crystalline rocks elsewhere. In the five study sites for the final disposal of the spent nuclear fuel in Finland the hydraulic conductivity of the upper part of crystalline bedrock varies from 10^{-11} to 10^{-4} ms^{-1} (Anon 1992). According to Brusila (1983), the mean hydraulic conductivity in Finland is near the ground surface 1,8x10^{-7} ms^{-1} and at the depth of 200 m 2,0x10^{-8} ms^{-1} (n=856 water loss measurements).

Hydraulic conductivity of granite and gneiss areas in south Sweden is around 10^{-8}...10^{-6} ms^{-1} (Carlsson & Olsson 1976, 1977a, 1977b, Carlsson & Carlstedt 1977, Hult et al. 1978). In Swedish Precambrian bedrock, Carlsson et al. (1979) have measured hydraulic conductivities of 10^{-6}…10^{-5} ms^{-1} from 400 m below the ground surface. According to Henriksen (2008), the prevailing hydraulic conductivity of drilled wells in the large dataset from Norway and Sweden (n~140000) is 10^{-8}...10^{-6} ms^{-1}. Based on median yields of Fennoscandian bedrock wells Banks et al. (2010) have suggested a bulk hydraulic conductivity of around 1,1x10^{-7} ms^{-1}, which is exactly the same as in this study. They also presented an overall median K value of 6,75x10^{-8} ms^{-1} for wells drilled in pre-Caledonian basement areas of

Sweden (n=93300; e.g. Antal et al. 1998). Banks et al. (2010) argued that the slight discrepancy between their data and the Swedish data may, in fact, be consistent with the Swedish practice of estimating hydraulic conductivity by dividing yield by depth x depth (e.g. Follin et al. 1999) rather than by depth x drawdown, which Banks et al. (2010) assumed to be 2/3 of the saturated depth.

The permeabilities of gneiss and granite rocks of the Black Forest region, Germany, vary from $3{,}5 \times 10^{-10}$ to $8{,}7 \times 10^{-5}$ ms^{-1} to a depth of 3,5 km (Stober 1997). In crystalline basement of northern Switzerland hydraulic conductivity ranges from 1×10^{-13} to 4×10^{-4} ms^{-1} or nine orders of magnitude between 1 and 2,5 km depth (Stober & Bucher 2007).

The average hydraulic conductivity over the upper 100 m of the schist and granite bedrock of the Mirror Lake Site in New Hampshire, USA, is 3×10^{-7} ms^{-1} (Shapiro & Hsieh 1996). The median hydraulic conductivity of the crystalline rocks in the Blue Ridge of Pennsylvania, USA, is $3{,}5 \times 10^{-8}$ ms^{-1} (Low 2004). In crystalline bedrock of New Hampshire, USA, hydraulic conductivity ranges over 6 orders of magnitude, from the lower detection limit of $1{,}0 \times 10^{-10}$ ms^{-1} to $5{,}0 \times 10^{-5}$ ms^{-1} (Johnson 1999). The hydraulic conductivity values range over five orders of magnitude in one well in a Silurian dolomite aquifer in Wisconsin, USA ($7{,}0 \times 10^{-9} \ldots 8{,}0 \times 10^{-4}$ ms^{-1}; Davison 1990, Rayne et al. 2001).

Dickin et al. (1984) consider that in the Canadian Shield the permeability range in the upper parts of bedrock (a few hundred meters thick) is between 10^{-8} and 10^{-6} ms^{-1}. In the crystalline bedrock of the Whiteshell Research Area, southeastern Manitoba, hydraulic conductivity varies from 10^{-16} to 10^{-6} ms^{-1} or 10 orders of magnitude; the K values of fracture zones vary from 1×10^{-12} to 5×10^{-6} ms^{-1} (Stevenson et al. 1995). The mean hydraulic conductivities of drilled wells in granite and gneiss areas of Saguenay-Lac-Saint-Jean area near Quebec, Canada, are $2{,}29 \times 10^{-6}$ and $2{,}06 \times 10^{-6}$ ms^{-1}, respectively (n=382 and 313; Richard et al. 2011).

7.2.5 Relations between well yield and hydraulic parameters

The well yield and hydraulic parameters correlate very highly significantly with each other (Table 29). Spearman rank correlation coefficients are in general higher than Pearson product moment correlation coefficients, which indicates that the relations between various parameters are not necessarily linear. In addition, the close relationship between transmissivity, normalized yield and specific capacity and that between hydraulic conductivity and well productivity can clearly be seen.

Table 29. Pearson product moment (P) and Spearman rank correlation coefficients (S) between well yield and hydraulic parameters for all drilled wells in the CF database (n=1300). All correlation coefficients are very highly significant (p <0,0001).

		Q	Q/d_s	Q/s	Q_w	T
Q/d_s	P	0,48823				
	S	0,91892				
Q/s	P	0,49673	0,99991			
	S	0,91892	1,00000			
Q_w	P	0,14031	0,86432	0,85885		
	S	0,79618	0,96771	0,96771		
T	P	0,50492	0,99963	0,99991	0,85341	
	S	0,91892	1,00000	1,00000	0,96771	
K	P	0,14134	0,86543	0,85999	0,99999	0,85458
	S	0,79463	0,96703	0,96703	0,99999	0,96703

The mutual relations between different hydraulic parameters are presented in Table 30 and in Fig. 55. Typical normalized well yields are about one quarter of the corresponding specific capacities and less than half of the corresponding transmissivities. On the other hand, the specific capacities and well productivities are nearly twice those of T- and K-values, respectively. In the study of Huntley et al. (1992), the Q/s values were approximately three times those of the corresponding T-values in the Peninsular Ranges batholith of San Diego County, USA. Because T-values estimated from specific capacity data apply only to the aquifer adjacent to the well, they frequently are greater than those obtained from aquifer tests (Singhal & Gupta 1999).

Table 30. Relations between different hydraulic parameters of all drilled wells in the CF database (n=1300).

Variable relation	Dimension	DRILLED WELLS IN THE CF DATABASE n=1300				
		Mean	Std dev	Median	Minimum	Maximum
$(Q/d_s)/(Q/s)$	m^2s^{-1}/ m^2s^{-1}	0,24	0,01	0,24	0,21	0,26
$(Q/d_s)/T$	m^2s^{-1}/ m^2s^{-1}	0,45	0,02	0,45	0,36	0,53
$(Q/s)/T$	m^2s^{-1}/ m^2s^{-1}	1,89	0,05	1,89	1,70	2,04
$(Q_w)/K$	ms^{-1}/ms^{-1}	1,89	0,05	1,89	1,70	2,04

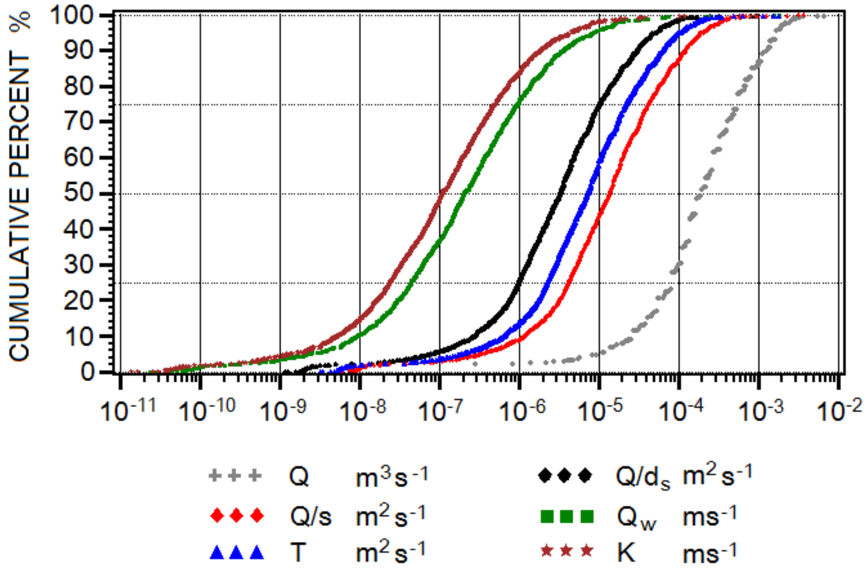


Fig. 55. Cumulative percent frequencies of well yield and hydraulic parameters of all drilled wells in the CF database (n=1300).

The regression equation

$$\log (Q/s) = -1{,}48 + 1{,}11 \log Q \qquad (58)$$

where Q/s is expressed in $Lhr^{-1}m^{-1}$ and Q in Lhr^{-1}, can be used for roughly estimating specific capacities of drilled wells in Central Finland (Fig. 56). The equation (58) can also be written as

$$Q/s = 0{,}033\ Q^{1{,}112} \tag{59}$$

Then the corresponding drawdown (s) in a well can be estimated as

$$s = Q/(Q/s) \tag{60}$$

where s is expressed in meters.

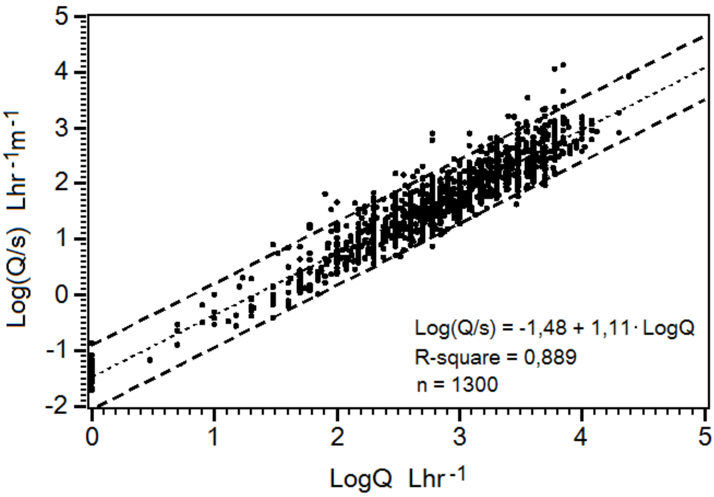


Fig. 56. The log-log regression between specific capacity (Q/s) and well yield (Q) with 95% confidence limits for individual predicted values for drilled wells in the CF database (n=1300, test wells included).

For roughly estimating the bulk T and K values of drilled wells in Central Finland, specific regression equations are given in Table 31. They are based on logarithmic transformations of well yield and hydraulic parameters for all wells in the CF database.

Table 31. Regression equations between the response variables T and K and the explanatory variables Q, Q/d_s and Q/s for all drilled wells in CF database (n=1300).

Equation	R^2
$T\ [m^2s^{-1}] = 5{,}377 \times 10^{-9} * Q^{1{,}097}\ [Lhr^{-1}]$	0,8885
$T\ [m^2s^{-1}] = 6{,}591 \times 10^{-7} * (Q/d_s)^{0{,}972}\ [Lhr^{-1}m^{-1}]$	1,0000
$T\ [m^2s^{-1}] = 0{,}454 * (Q/s)^{0{,}986}\ [m^2s^{-1}]$	1,0000
$T\ [m^2s^{-1}] = 1{,}718 * K^{0{,}778}\ [ms^{-1}]$	0,9463
$K\ [ms^{-1}] = 3{,}828 \times 10^{-11} * Q^{1{,}224}\ [Lhr^{-1}]$	0,7048
$K\ [ms^{-1}] = 6{,}545 \times 10^{-9} * (Q/d_s)^{1{,}185}\ [Lhr^{-1}m^{-1}]$	0,9463
$K\ [ms^{-1}] = 0{,}450 * (Q_w)^{0{,}989}\ [ms^{-1}]$	1,0000
$K\ [ms^{-1}] = 0{,}222 * T^{1{,}218}\ [m^2s^{-1}]$	0,9463

The median yield of high-yield wells is 120 times that of low-yield wells (Table 32). The corresponding difference in normalized well yield is 300-fold. Also other hydraulic parameters are 2-3 orders of magnitude greater in high-yield wells compared to low-yield wells. The differences in well yield and hydraulic parameters between low-yield and high-yield wells are statistically very highly significant (p <0,0001). The yield of a dry well has been entered into calculations as a nominal figure of 1 Lhr^{-1}.

Table 32. Statistics for well yield and hydraulic parameters in low-yield ($Q \leq 100\ Lhr^{-1}$) and high-yield ($Q \geq 4000\ Lhr^{-1}$) drilled wells in the CF database (n=317, test wells included).

Variable	Dimension	LOW-YIELD WELLS $Q \leq 100\ Lhr^{-1}$ n=152				
		Mean	**Std dev**	**Median**	**Minimum**	**Maximum**
Q	Lhr^{-1}	47	36	50	1	100
Q/d_s	$Lhr^{-1}m^{-1}$	0,9	1,8	0,4	0,004	16
Q/s	m^2s^{-1}	$1{,}05x10^{-6}$	$2{,}15x10^{-6}$	$4{,}57x10^{-7}$	$5{,}54x10^{-9}$	$1{,}85x10^{-5}$
Q_w	ms^{-1}	$5{,}87x10^{-8}$	$3{,}30x10^{-7}$	$3{,}80x10^{-9}$	$2{,}36x10^{-11}$	$3{,}71x10^{-6}$
T	m^2s^{-1}	$5{,}67x10^{-7}$	$1{,}14x10^{-6}$	$2{,}53x10^{-7}$	$3{,}26x10^{-9}$	$9{,}77x10^{-6}$
K	ms^{-1}	$3{,}13x10^{-8}$	$1{,}74x10^{-7}$	$2{,}10x10^{-9}$	$1{,}39x10^{-11}$	$1{,}95x10^{-6}$
Variable	**Dimension**	**HIGH-YIELD WELLS $Q \geq 4{,}000\ Lhr^{-1}$ n=165**				
		Mean	**Std dev**	**Median**	**Minimum**	**Maximum**
Q	Lhr^{-1}	6474	3078	6000	4000	24000
Q/d_s	$Lhr^{-1}m^{-1}$	213	399	120	22	3500
Q/s	m^2s^{-1}	$2{,}36x10^{-4}$	$4{,}29x10^{-4}$	$1{,}35x10^{-4}$	$2{,}59x10^{-5}$	$3{,}76x10^{-3}$
Q_w	ms^{-1}	$3{,}02x10^{-5}$	$1{,}93x10^{-4}$	$2{,}63x10^{-6}$	$1{,}16x10^{-7}$	$1{,}88x10^{-3}$
T	m^2s^{-1}	$1{,}19x10^{-4}$	$2{,}11x10^{-4}$	$6{,}93x10^{-5}$	$1{,}36x10^{-5}$	$1{,}84x10^{-3}$
K	ms^{-1}	$1{,}49x10^{-5}$	$9{,}46x10^{-5}$	$1{,}34x10^{-6}$	$6{,}08x10^{-8}$	$9{,}21x10^{-4}$

7.3 Factors affecting well yield and hydraulic properties

7.3.1 Construction factors

Until the mid-70's the median depth of private drilled wells in Central Finland was around 45 meters (Fig. 57). From that onwards the well depths began to steadily increase till in the early 2000’s the median well depth reached 120 m. The main reason for the depth increase has been the introduction of more powerful drilling equipments. The wells constructed before the mid-70’s were drilled with the cable tool method, which was displaced by the more efficient DTH-method (Mäkelä 1994a). Although the median well depths have significantly increased from the mid-1970’s, the corresponding well yields have remained unchanged or even decreased (Fig. 58).

Also in Sweden the median depth of drilled wells has slowly increased from the 1970’s (Gierup 1994). In Stockholm county, for instance, the median depth of wells was 57 m in 1976; 13 years later it was 75 m. Gierup suggests that some reasons for this might have been more effective drilling technique, price fixing with drilling depth and increased water demand. In Norway, the median depth of drilled wells has increased since 1974. At the same time the median yield of wells has decreased (Morland 1997). Morland considers that the reason for this has been the introduction of faster and more efficient drilling techniques, which may result in clogging of fissures and cracks during drilling. In crystalline rocks of

Loudoun County, Virginia, USA, the mean depth of wells drilled in 1953-1996 has increased with time, while at the same time the mean yield per foot of drilling has decreased (Sutphin et al. 2001).

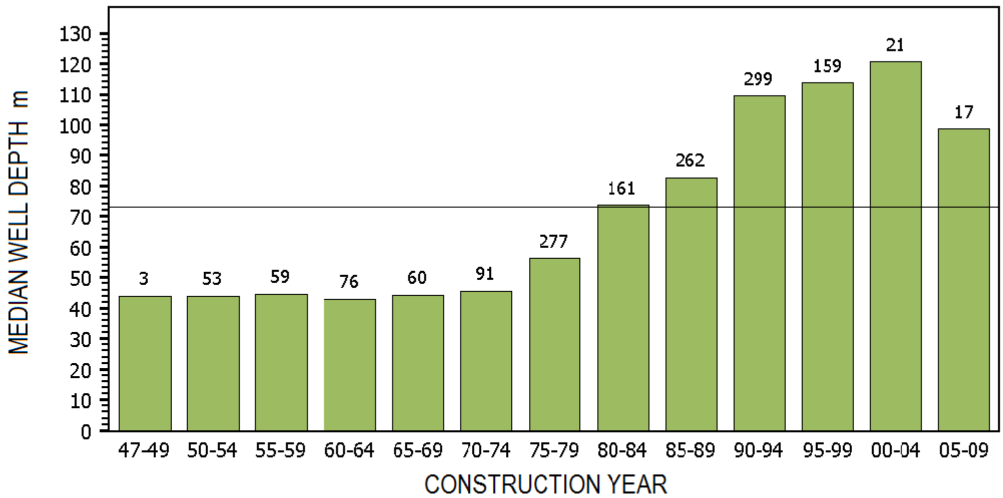


Fig. 57. Bar chart showing the median depth (m) of private drilled wells in the CF database in five-year periods of construction (n=1538). The number of wells per bar is indicated above each bar and the overall median depth of private wells (73 m) is marked with a horizontal line.

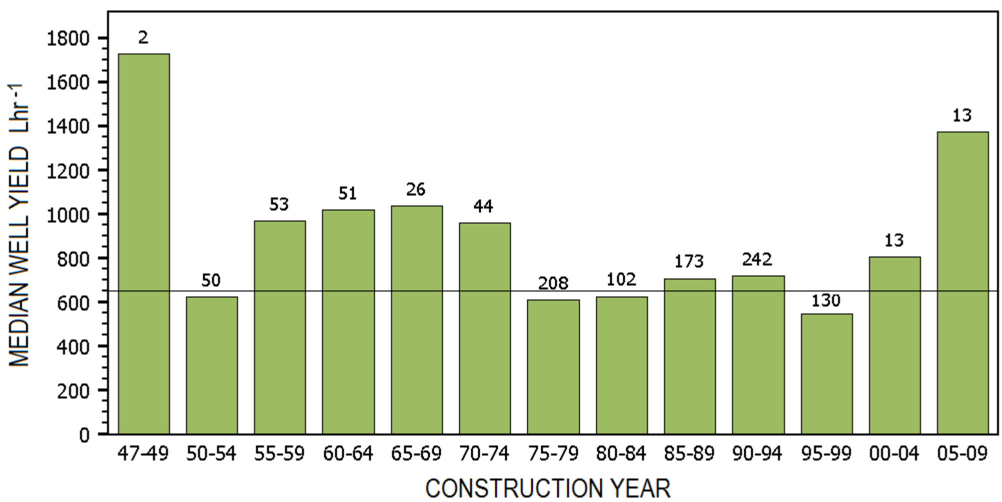


Fig. 58. Bar chart showing the median yield (Q Lhr^{-1}) of private drilled wells in the CF database in five-year periods of construction (n=1107). The number of wells per bar is indicated above each bar and the overall median yield of private wells (650 Lhr^{-1}) is marked with a horizontal line.

To test if the drilling technique had some influence on the well hydraulic properties, the private well data were roughly divided into two groups: cable tool wells (CY 1947-1974) and DTH-wells (CY 1975-2008). The median Q/d_s, Q/s and T values of cable tool wells are twice those of DTH-wells and the median Q_w and K values are more than 4-fold in cable tool wells compared to DTH-wells (Table 33). The differences are statistically significant.

The DTH-wells are significantly deeper than the cable-tool wells in the CF database (Table 33). The well depth inevitably affects the hydraulic properties of the wells that are normalized with the saturated open well section, which in turn is directly proportional to the total well depth (Table 13). In addition, deeper casing may decrease well yield by preventing the water in the surficial part of bedrock to enter the well. The median casing depth of DTH-wells (10 m) is twice that of cable tool wells (5 m); the corresponding difference in median overburden thicknesses is one meter only.

Interestingly, all dry wells (Q 0 Lhr^{-1}) in the CF database (n=27) have been drilled with the DTH-method. On the other hand, the most productive wells in the CF database are DTH-wells. The median depth of DTH-wells drilled since 1990 is 109 m and the corresponding Q and Q/d_s values are 600 Lhr^{-1} and 7 $Lhr^{-1}m^{-1}$, respectively (n=451).

Table 33. Statistics for well depth, yield and hydraulic parameters of private wells in the CF database drilled with the cable tool and DTH-methods (n=1103). Highest values are in bold. Comparisons significant at the α≤0.05 level between the groups tested with the nonparametric Kruskal-Wallis test for group means and with the median one-way analysis for group medians are indicated by different letters $^{A, B}$.

Variable	Parameter	Cable tool wells CY 1947-1974 n=225 20%	DTH-wells CY 1975-2008 n=878 80%
DEPTH m	mean median	53^{B} 49^{B}	**99^{A}** **91^{A}**
Q Lhr^{-1}	mean median	**1640^{A}** **800^{A}**	1433^{B} 600^{B}
Q/d_s $Lhr^{-1}m^{-1}$	mean median	**72^{A}** **18^{A}**	38^{B} 9^{B}
Q/s m^2s^{-1}	mean median	**$8{,}05x10^{-5\ A}$** **$2{,}10x10^{-5\ A}$**	$4{,}23x10^{-5\ B}$ $1{,}02x10^{-5\ B}$
Q_w ms^{-1}	mean median	**$1{,}18x10^{-5\ A}$** **$6{,}37x10^{-7\ A}$**	$3{,}66x10^{-6\ B}$ $1{,}41x10^{-7\ B}$
T m^2s^{-1}	mean median	**$4{,}09x10^{-5\ A}$** **$1{,}11x10^{-5\ A}$**	$2{,}16x10^{-5\ B}$ $5{,}43x10^{-6\ B}$
K ms^{-1}	mean median	**$5{,}86x10^{-6\ A}$** **$3{,}36x10^{-7\ A}$**	$1{,}82x10^{-6\ B}$ $7{,}50x10^{-8\ B}$

Jammallo (1984), Lewis (1990) and Moore et al. (2002a) have suggested a negative correlation between elevation and yield in crystalline rock areas, i.e. high elevations tend to have low well yields and vice versa. In this study, the height of the well site ASL has no statistically significant influence on well yield or hydraulic parameters (Table 34). Neither does the well construction year CY nor the well diameter DIA correlate with the well yield. The depth to the groundwater table GWT has a negative correlation with Q and hydraulic parameters. Also in crystalline basement areas of southern Zimbabwe higher well yields correlate with shallow water tables (Owen et al. 2003).

DTH-wells drilled especially in 80's and 90's are deeper and larger in diameter than wells drilled earlier, which explains the strengthened negative correlations of hydraulic parameters with CY, DIA and DEPTH compared to those with Q (Table 34). Also STR has negative correlations with well yield and hydraulic parameters: the deeper the first (main) water strike the lower the yield and hydraulic values.

Table 34. Spearman rank correlation coefficients (r) with significance levels (p) and number of observations (n) between well construction factors and well yield and hydraulic parameters for all drilled wells in the CF database. Pairs having correlation coefficients ≥ |0,5000| are in bold. The explanations of variable abbreviations are given in Table 3 and in Chapter 6.3.2.

		Q	Q/d_s	Q/s	Q_w	T	K
ASL	r	-0,04528	-0,04009	-0,04009	-0,03543	-0,04009	-0,03551
n=1288	p	0,1043	0,1504	0,1504	0,2038	0,1504	0,2028
CY	r	-0,02959	-0,21177	-0,21177	-0,31897	-0,21177	-0,32010
n=1176	p	0,3106	<0,0001	<0,0001	<0,0001	<0,0001	<0,0001
DIA	r	-0,00887	-0,19655	-0,19655	-0,30218	-0,19655	-0,30335
n=1089	p	0,7701	<0,0001	<0,0001	<0,0001	<0,0001	<0,0001
CAS	r	0,07320	0,02432	0,02432	-0,01511	0,02432	-0,01543
n=727	p	0,0485	0,5127	0,5127	0,6843	0,5127	0,6778
GWT	r	-0,14119	-0,16330	-0,16330	-0,15499	-0,16330	-0,15510
n=610	p	0,0005	<0,0001	<0,0001	0,0001	<0,0001	0,0001
STR	r	-0,31798	-0,49651	-0,49651	**-0,57748**	-0,49651	**-0,57810**
n=643	p	<0,0001	<0,0001	<0,0001	**<0,0001**	<0,0001	**<0,0001**
SAT	r	0,29416	**0,60498**	**0,60498**	**0,76961**	**0,60498**	**0,77105**
n=610	p	<0,0001	**<0,0001**	**<0,0001**	**<0,0001**	**<0,0001**	**<0,0001**
DEPTH	r	-0,30431	**-0,62038**	**-0,62038**	**-0,78006**	**-0,62038**	**-0,78158**
n=1300	p	<0,0001	**<0,0001**	**<0,0001**	**<0,0001**	**<0,0001**	**<0,0001**
WDEM	r	0,21177	0,12610	0,12610	0,06518	0,12610	0,06438
n=536	p	<0,0001	0,0035	0,0035	0,1318	0,0035	0,1366

In crystalline rocks, there is often dependence of well yield on well depth, with decreasing well yield coinciding with increasing well depth (LeGrand 1954, Davis & Turk 1964, Lewis & Burghy 1964, Summers 1972, Landers & Turk 1973, Woolley 1982, Jammallo 1984, Larsson et al. 1984, Brook 1988, Lewis 1990, Banks 1992b, Henriksen 1995, 2003a, Singhal & Gupta 1999, Kenny et al. 2006, Courtois et al. 2010).

Ellis (1906) found that below a depth of 200 feet (61 m) the chances of obtaining water greatly decrease. Daniel (1989) discovered that, for a well of a given diameter, the yield per foot of hole is inversely proportional to the depth of the well, indicating that the amount of additional water obtained by drilling deeper is continuously decreasing. On the other hand, some authors have found no change in well yield with increasing depth (Loiselle & Evans 1995, Acheampong & Hess 1998, Mabee 1999, Verbovšek & Veselič 2008). According to Fagerlind (1986), in highly fractured rock the yield of wells increases with increasing depth. Cederstrom (1972) states that deeper wells generally yield more water than shallower wells. In granitic aquifers of South India, Briz-Kishore (1993) found an approximately positive linear relationship between yield and depth down to a particular level with yield remaining constant thereafter.

In Central Finland, the well depth has a very highly significant negative correlation with well yield Q and hydraulic parameters (Table 34). Median well yields decrease gradually until the wells reach the depth of around 150 m with the yield remaining nearly constant from that downwards (Fig. 59). The decrease of hydraulic parameters with depth is even more dramatic (Figs. 60 and 61). This is partly because of the parameter normalization with the saturated open well section. Similar results have been gained, for example, in Sweden (e.g. Ahlbom et al. 1991a).

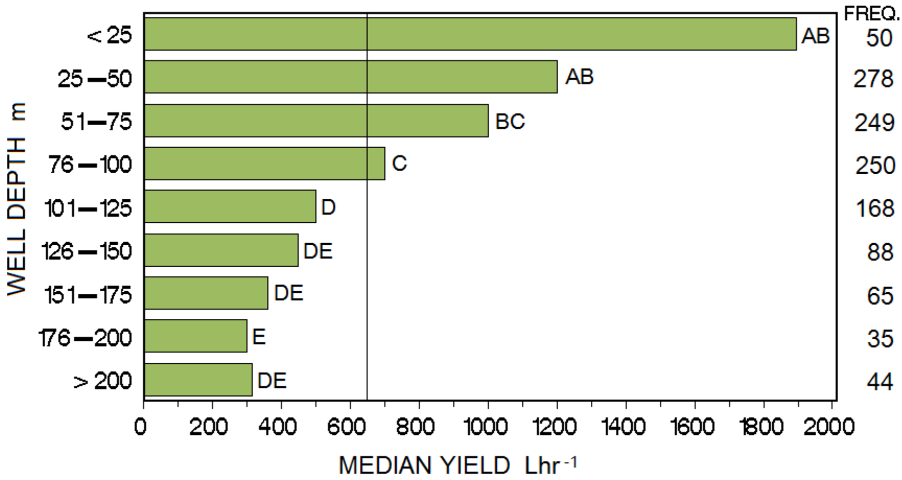


Fig. 59. Bar chart showing the median yield of private drilled wells in the CF database in different depth groups (n=1227). The overall median yield of private wells in the CF database (650 Lhr^{-1}) is marked with a vertical line. Comparisons significant at the α≤0.05 level between the groups tested with the nonparametric median one-way analysis for group medians are indicated by different letters A, B, etc. The number of wells per bar is marked in the rightmost column.

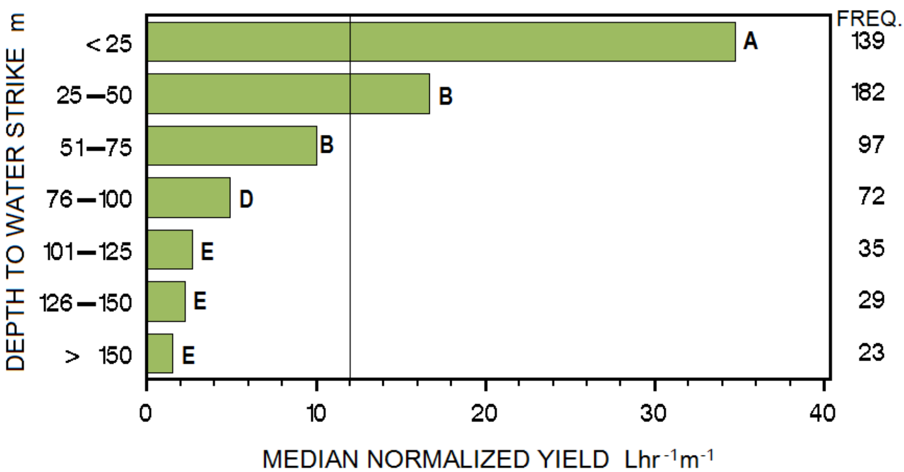


Fig. 60. Bar chart showing the median normalized yield (Q/d_s $Lhr^{-1}m^{-1}$) of private drilled wells in the CF database in different groups of STR (n=577). The overall median normalized yield of private wells (12 $Lhr^{-1}m^{-1}$) is marked with a vertical line. Comparisons significant at the α≤0.05 level between the groups tested with the nonparametric median one-way analysis for group medians are indicated by different letters A, B, etc. The number of wells per bar is indicated in the rightmost column.

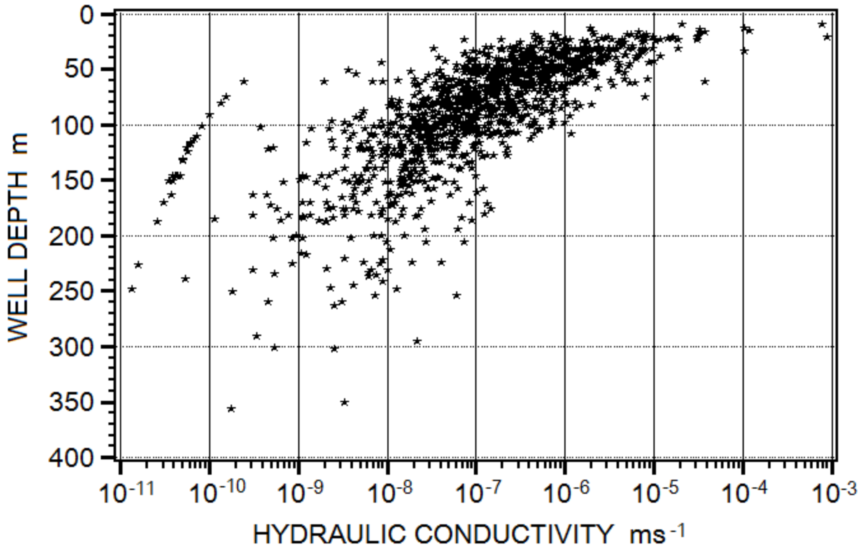


Fig. 61. Hydraulic conductivity (ms^{-1}) vs. total well depth (m) for private drilled wells in the CF database (n=1227).The separate group in the left part of the graph represents for most part dry wells with a nominal yield of 1 Lhr^{-1}.

Stober (1997) has given the following regression equation for drilled wells in the Black Forest crystalline basement area in Germany

$$\log D = 0{,}946 - 0{,}149 \log K \qquad (61)$$

where D is the well depth (m) and K is the hydraulic conductivity (ms^{-1}); the coefficient of determination was R^2=0,186. In Central Finland the corresponding regression equation for private drilled wells (n=1227, R^2=0,586) is

$$\log D = 0{,}581 - 0{,}184 \log K \qquad (62)$$

In Central Finland, the hydraulic conductivity shows very large variance near the surface (< 25 m; > 2x10^{-8}) compared to the deeper parts of the bedrock (> 150 m; <1x10^{-15}). Stober and Bucher (2007) have made similar observations in the Black Forest region in Germany.

Henriksen (2006b) argues that when the computation of hydraulic properties is based on yield and drilled rock depth, the values of hydraulic parameters may be overestimated in shallow high-yield wells and underestimated in deep low-yield wells. On the other hand, in big samples these over- and underestimates will tend to cancel one another out (Henriksen 2006b). According to Henriksen (2008), the variability in hydraulic conductivity and flow rate values is best explained by the drilled rock depth. According to Brusila (1983), decreasing hydraulic conductivity with increasing depth is a common finding of water loss measurements made in Finland.

The statistical examination of well yield and hydraulic parameters in different well depth groups is presented in Table 35 and Fig. 62.

Table 35. Statistics for well yield and hydraulic parameters in different depth groups of all drilled wells in the CF database (n=1300). Highest values are in bold. Comparisons significant at the α≤0.05 level between the groups tested with the nonparametric Kruskal-Wallis test for group means and with the median one-way analysis for group medians are indicated by different letters [A, B, etc.].

Variable	Parameter	WELL DEPTH			
		< 50 m n=314 24%	50-100 m n=559 43%	101-150 m n=279 22%	> 150 m n=148 11%
Q Lhr^{-1}	mean median	**2105** ^{A} **1200** $^{\mathbf{A}}$	1764 ^{B} 850 ^{B}	1122 ^{C} 500 ^{C}	756 ^{D} 345 ^{D}
Q/d_s $Lhr^{-1}m^{-1}$	mean median	**121** ^{A} **47** $^{\mathbf{A}}$	32 ^{B} 14 ^{B}	11 ^{C} 4 ^{C}	5 ^{D} 2 ^{D}
Q/s m^2s^{-1}	mean median	$\mathbf{1{,}35x10^{-4\,A}}$ $\mathbf{5{,}39x10^{-5\,A}}$	$3{,}66x10^{-5\,B}$ $1{,}62x10^{-5\,B}$	$1{,}27x10^{-5\,C}$ $5{,}19x10^{-6\,C}$	$5{,}25x10^{-6\,D}$ $2{,}40x10^{-6\,D}$
Q_w ms^{-1}	mean median	$\mathbf{1{,}88x10^{-5\,A}}$ $\mathbf{2{,}01x10^{-6\,A}}$	$8{,}68x10^{-7\,B}$ $2{,}78x10^{-7\,B}$	$1{,}28x10^{-7\,C}$ $4{,}50x10^{-8\,C}$	$3{,}20x10^{-8\,D}$ $1{,}30x10^{-8\,D}$
T m^2s^{-1}	mean median	$\mathbf{6{,}82x10^{-5\,A}}$ $\mathbf{2{,}80x10^{-5\,A}}$	$1{,}89x10^{-5\,B}$ $8{,}55x10^{-6\,B}$	$6{,}67x10^{-6\,C}$ $2{,}78x10^{-6\,C}$	$2{,}78x10^{-6\,D}$ $1{,}30x10^{-6\,D}$
K ms^{-1}	mean median	$\mathbf{9{,}35x10^{-6\,A}}$ $\mathbf{1{,}05x10^{-6\,A}}$	$4{,}47x10^{-7\,B}$ $1{,}47x10^{-7\,B}$	$6{,}70x10^{-8\,C}$ $2{,}40x10^{-8\,C}$	$1{,}70x10^{-8\,D}$ $7{,}00x10^{-9\,D}$

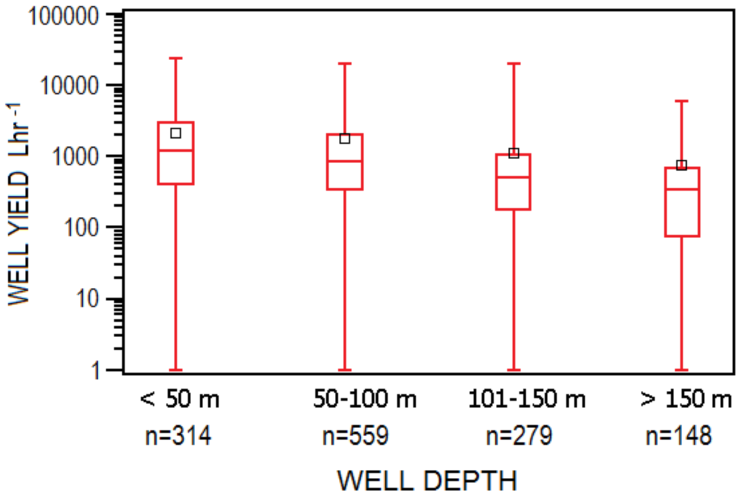


Fig. 62. Box-plot of well yield vs. well depth for all drilled wells in the CF database (n=1300). The maximum and minimum values and the 25th and 75th percentiles of the yield are presented with tops and bottoms on the whiskers and on the boxes, respectively. Median is presented with a horizontal line in the boxes and the mean with a square. The yield of a dry well has been entered into calculations as a nominal figure of 1 Lhr^{-1}.

According to Caswell (1987), large diameter wells enable better yields than small diameter wells. This is because of increase in borehole circumference and surface area exposure of the fractures intersected by the well bore. Entrance velocity and turbulence are reduced, as is the difficulty of moving water along the well bore to the pump intake, all of which allow more water to be pumped from the well with less drawdown (Caswell 1987). Also Daniel

(1987, 1989) has found that well diameter could have a significant influence on yield: for a given depth, yield is directly proportional to well diameter. Similar results come from this study at the depth intervals of 76-150 m and > 200 m (Table 36). However, the diameter of a well does not affect its yield in general (Table 34). Actually, in the CF database, the yield of a well decreases with depth while at the same time the well diameter increases (Table 36).

According to Morland (1997), the majority of bedrock entities in Norway showed a clear decrease in the median normalized yield of deep boreholes compared to shallow boreholes; for the rest no clear relationship could be identified. Morland (1997) suggests that the reason for this general decrease might be the diminishing frequency of horizontal joints within the bedrock due to the increasing vertical stress. Partly this may be only an artifact of the way in which water wells are drilled: where fractured rocks are encountered drilling stops at a shallow depth, and where rocks of low permeability exist wells are drilled deeper and the apparent overall permeability of the rock is low. In other words, the drilled depth represents the minimum depth required to obtain sufficient yield for the user, i.e. the yield controls the depth of the well. Because of this there is a bias in the depth/yield data (Davis 1981, Knopman 1990, Mabee 1992, Knopman & Hollyday 1993, Morland 1997, Moore et al. 2002a, Henriksen 2003a, Kenny et al. 2006, Courtois et al. 2010). Also the results of Wladis et al. (1997) suggest that wells in less conductive parts of the rock are drilled to greater depths, and further, if the yield is low near the surface, it remains low at greater depths. Huntley et al. (1992) put this as follows:"What is generally not taken into account in well depth/yield analysis is that a well that has a poor yield at say 200 m also had a poor yield at 30 m". Moore et al. (2002a) have noticed that, at a potentially high-yield site, drilling typically is stopped at a shallower depth than at an average site, which effectively reduces the number of deep high-yielding wells in that population. At low-yield sites, drilling is continued to a greater depth to obtain a desired yield, which consequently reduces the number of low-yielding shallow wells.

Table 36. Spearman rank correlations between well yield and diameter in different depth groups of private drilled wells in the CF database (n=1017). Pairs having statistically significant correlations are in bold.

DEPTH GROUP m	n	YIELD Md Lhr^{-1}	DIA Md mm	Spearman correlation coefficient	Statistical significance p
< 25	40	1580	110	0,24180	0,1328
25-50	228	1200	110	0,06072	0,3615
51-75	203	1000	110	0,08414	0,2326
76-100	**193**	**800**	**115**	**0,35072**	**<0,0001**
101-125	**142**	**500**	**140**	**0,36707**	**<0,0001**
126-150	**81**	**420**	**150**	**0,22551**	**0,0429**
151-175	59	300	160	0,03236	0,8078
176-200	31	200	157	0,11003	0,5557
> 200	**40**	**315**	**160**	**0,38701**	**0,0136**

Deepening the wells beyond the optimal depth has little or no effect on their ultimate specific capacities and yields (e.g. Acheampong & Hess 1998). However, specific capacities of deep wells (> 150 m) are not necessarily low. On the contrary, they can range from 1 to more than 1000 $Lhr^{-1}m^{-1}$ (Neves & Morales 2007a, 2007b). Daniel (1987, 1989) found that maximum well yields are obtained from much greater depths (150-250 m) than previously believed. According to Ramasesha et al. (2003), in Karnataka region in India deep drillings (> 100 m) in lineaments of the Precambrian basement have brought out encouraging results: water-yielding fractures have been encountered up to a depth of 187 m bgl and 60% of wells have yielded more than 10,800 Lhr^{-1}. Also in Ghana a realization that deeper (> 100 m) boreholes may well have higher yields (> 7200 Lhr^{-1}) has increased borehole success rates (Cobbing & Davies 2008).

In the CF database, there are eight high-yield wells ($Q \geq 4{,}000$ Lhr^{-1}), which are deeper than 150 m (Table 37). These deep high-yield wells consist only 5% of all high-yield wells in Central Finland while the proportion of deep wells is around 10% of all drilled wells in Central Finland. Most high-yield wells have their place in the shallowest well groups, whereas low-yield wells ($Q \leq 100$ Lhr^{-1}) dominate in the deepest well groups (Fig. 63).

Table 37. Deep high-yield drilled wells in the CF database (n=8).

WELL ID	DEPTH m	STR m	YIELD Lhr^{-1}
077026	205	35	4710
077087	180	7	6000
179126	171	112	5000
182169	160	--	4000
410145	157	--	4000
592097	185	--	4980
850002	253	16	5000
931009	175	165	6000

The fracturing degree of bedrock at well site (FRAC) significantly affects the well yield and hydraulic parameters (Table 38, Fig. 64). The more fractured the bedrock is, the higher are well yield and hydraulic values. The median well yield is 14-fold and the hydraulic values from 24- to 50-fold in the group ‘high’ compared to the group ‘low’. The FRAC affects the well depths, too, but in an opposite way. It should still be noted that there are low-yield wells in highly fractured bedrock and vice versa (Table 39).

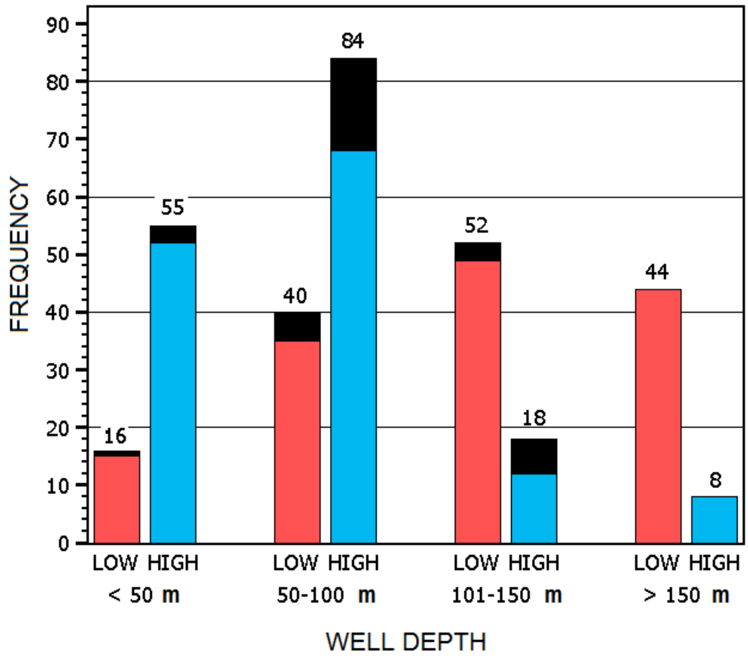


Fig. 63. Bar chart showing the distribution of low-yield ($Q \leq 100$ Lhr^{-1}) and high-yield ($Q \geq 4{,}000$ Lhr^{-1}) drilled wells in the CF database in different depth groups. The number of wells per bar is indicated above each bar (n=317, test wells marked in black).

Table 38. Statistics for well depth, yield and hydraulic parameters in different groups of fracturing degree of bedrock of all drilled wells in the CF database (n=444). Highest values are in bold. Comparisons significant at the $\alpha \leq 0.05$ level between the groups tested with the nonparametric Kruskal-Wallis test for group means and with the median one-way analysis for group medians are indicated by different letters $^{A, B, etc.}$.

Variable	Parameter	Fracturing degree of the bedrock at well site (FRAC)		
		LOW n=253 57%	MEDIUM n=128 29%	HIGH n=63 14%
DEPTH m	mean median	**105** ^{A} **92** ^{A}	80 ^{B} 69 ^{B}	58 ^{C} 52 ^{C}
Q Lhr^{-1}	mean median	769 ^{C} 350 ^{C}	1823 ^{B} 1100 ^{B}	**5202** ^{A} **4800** ^{A}
Q/d_s $Lhr^{-1}m^{-1}$	mean median	16 ^{C} 4 ^{C}	45 ^{B} 18 ^{B}	**210** ^{A} **100** ^{A}
Q/s m^2s^{-1}	mean median	$1{,}82 \times 10^{-5}$ ^{C} $4{,}55 \times 10^{-6}$ ^{C}	$5{,}11 \times 10^{-5}$ ^{B} $2{,}03 \times 10^{-5}$ ^{B}	**$2{,}31 \times 10^{-4}$** ^{A} **$1{,}13 \times 10^{-4}$** ^{A}
Q_w ms^{-1}	mean median	$5{,}73 \times 10^{-7}$ ^{C} $5{,}30 \times 10^{-8}$ ^{C}	$1{,}95 \times 10^{-6}$ ^{B} $3{,}42 \times 10^{-7}$ ^{B}	**$3{,}87 \times 10^{-5}$** ^{A} **$2{,}63 \times 10^{-6}$** ^{A}
T m^2s^{-1}	mean median	$9{,}44 \times 10^{-6}$ ^{C} $2{,}44 \times 10^{-6}$ ^{C}	$2{,}63 \times 10^{-5}$ ^{B} $1{,}07 \times 10^{-5}$ ^{B}	**$1{,}16 \times 10^{-4}$** ^{A} **$5{,}80 \times 10^{-5}$** ^{A}
K ms^{-1}	mean median	$2{,}95 \times 10^{-7}$ ^{C} $2{,}80 \times 10^{-8}$ ^{C}	$9{,}98 \times 10^{-7}$ ^{B} $1{,}80 \times 10^{-7}$ ^{B}	**$1{,}91 \times 10^{-5}$** ^{A} **$1{,}34 \times 10^{-6}$** ^{A}

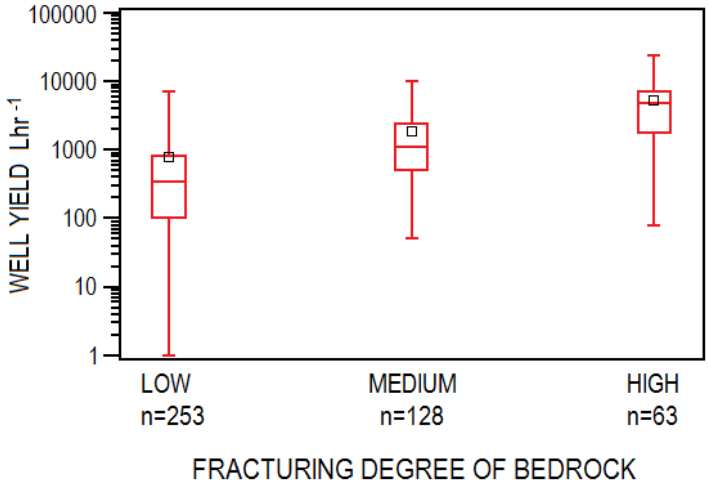


Fig. 64. Box-plot of well yield vs. fracturing degree of bedrock for all drilled wells in the CF database (n=444). The maximum and minimum values and the 25th and 75th percentiles of the yield are presented with tops and bottoms on the whiskers and on the boxes, respectively. Median is presented with a horizontal line in the boxes and the mean with a square. The yield of a dry well has been entered into calculations as a nominal figure of 1 Lhr^{-1}.

Table 39. Proportion of low-yield (Q≤ 100 Lhr^{-1}) and high-yield (Q≥ 4,000 Lhr^{-1}) drilled wells in different FRAC groups of the CF database (n=444, test wells included).

FRAC	n	LOW-YIELD WELLS Q ≤ 100 Lhr^{-1} %	HIGH-YIELD WELLS Q ≥ 4000 Lhr^{-1} %
LOW	253	26	5
MEDIUM	128	1	14
HIGH	63	2	55

Many studies in crystalline rock areas show that almost all of the flow conducted into or out of boreholes during aquifer tests enters or exits through a very small subpopulation of the fractures that intersect the borehole (Ellis 1906, Carlsson & Olsson 1977c, Rand 1978, Magnusson & Duran 1984, Long & Billaux 1987, Paillet et al. 1987, Cohen 1993, Paillet 1995, Anon 1996, Banks et al. 1996, Hsieh & Shapiro 1996, Barton et al. 1998, Taylor et al. 1999, Gudmundsson 2000b, Lyslo 2000, Mazurek 2000, Sharp et al. 2000, Caine & Tomusiak 2003, Rogers 2003, Rutqvist & Stephansson 2003, Min et al. 2004, Martínez-Landa & Carrera 2005, Lachassagne 2008, Boutt et al. 2010). Often one fracture dominates the production of a well in hard rock aquifers (Williamson & Woolley 1980, Briz-Kishore & Bhimasankaram 1982, Howard et al. 1992, Atkinson et al. 1994, Rebouças 1999b).

Results from test well drillings in Central Finland confirm these observations: most often groundwater enters a borehole from a few fractures while 1-3 of these fractures dominate the well yield (Figs. 65 and 66).

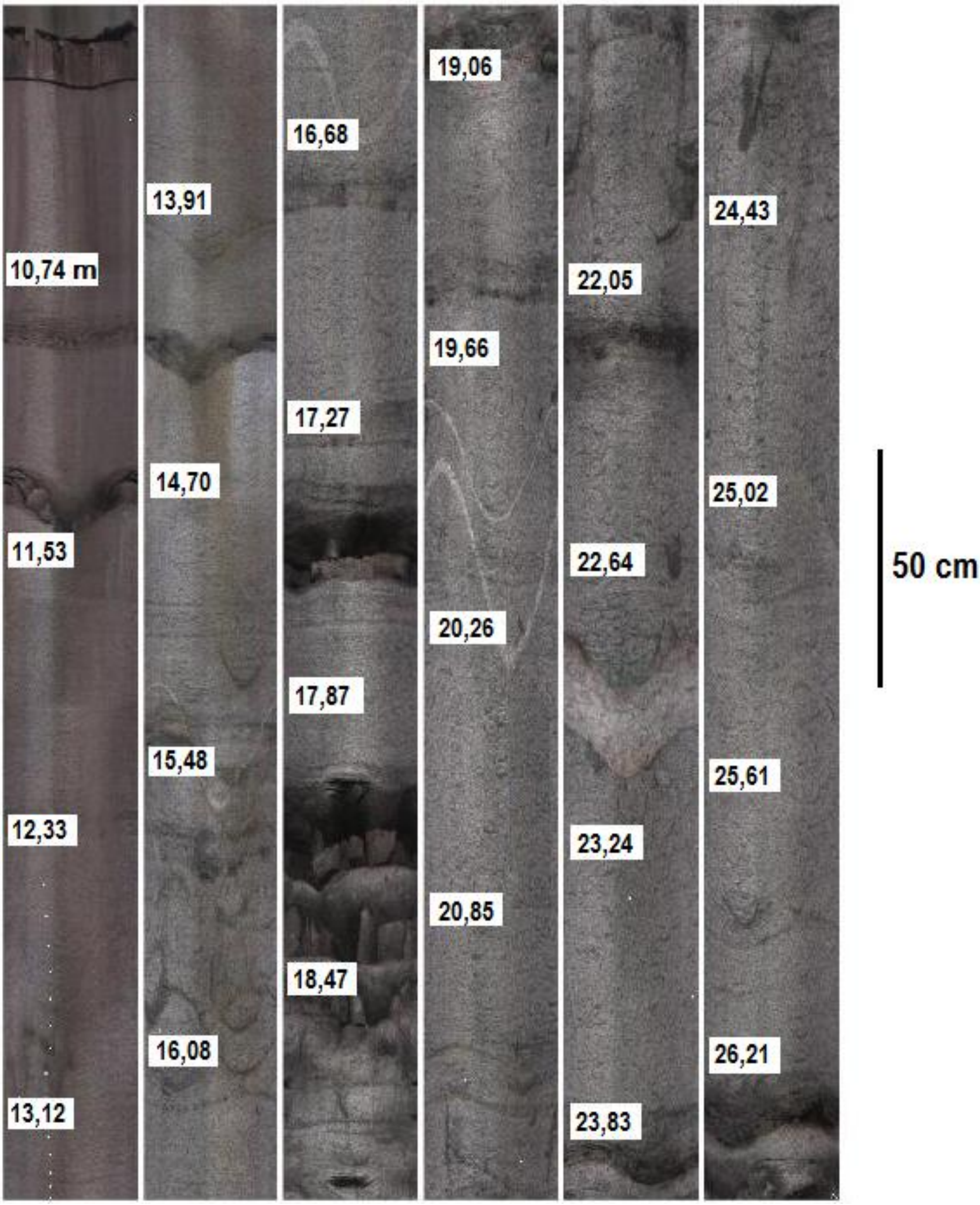


Fig. 65. The upper portion of an oriented digital borehole camera image from the Luotolansaari test well no. 931164 in the municipality of Viitasaari, Central Finland. The camera generates a 360° image of the borehole wall below groundwater table. The image covers borehole wall between 10,20 and 26,50 meters below ground level. Well depth in meters is indicated along the images; the total well depth is 91 m. The two fractures at 17,5 and 18,1 m bgl dominate the production of the well (Q=9,000 Lhr^{-1}). The bottom of the casing is seen in the top left corner of the image. Dipping planar features cutting through the wellbore appear as sinusoids in the unwrapped image. Modified from Ikävalko (2006).

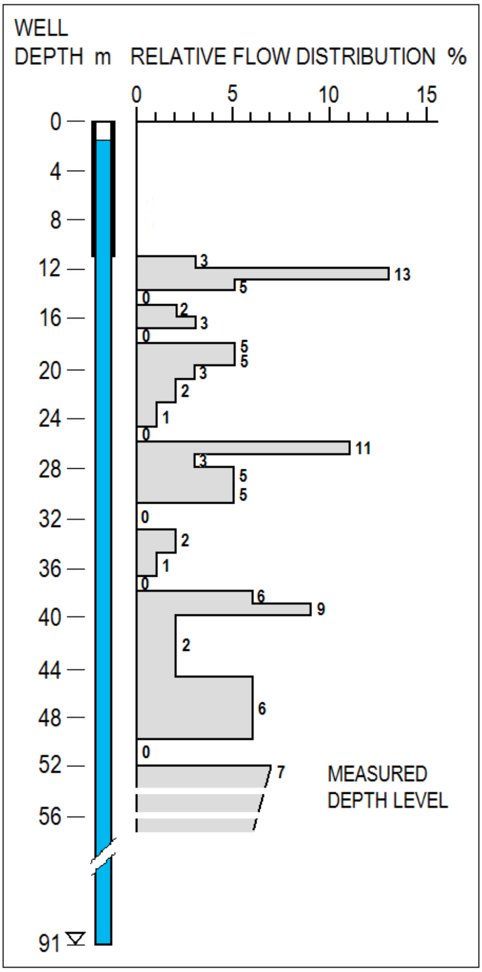


Fig. 66. Vertical distribution of groundwater flow in the Käräjämäki test well no. 216037 in the municipality of Kannonkoski, Central Finland. Drilling date 11/12/1996, total well depth 91 m, Q/s_m=1,050 $Lhr^{-1}m^{-1}$. Flow measurements by Finnish Groundwater Technics Ltd. 13/5/1997.

7.3.2 Geologic factors

The distribution of the wells in the CF database is in relation to different soil types in Central Finland. The dominant soil type at well sites is till (87%), whereas coarse-grained sediments (sand, gravel) and fine-grained sediments (silt, clay) cover only 5% and 3% of the sites, respectively (Table 40). Some 5% of the wells have been drilled on bedrock outcrops. The median overburden thickness at well sites is 3 meters; the mean is 6 meters. About one quarter of the wells in the CF database has overburden thickness of one meter or less.

In general, the mean and median well yield and hydraulic values are significantly higher in wells drilled in fine-grained areas compared to till and outcrop wells (Table 40, Fig. 67). Well yields are not statistically different between fine-grained and coarse-grained areas. The lowest mean and median values of yield and hydraulic parameters are detected in outcrop wells. Mean and median well depths between soil type groups do not differ statistically from each other.

Table 40. Statistics for well depth, yield and hydraulic parameters in different soil type groups of all drilled wells in the CF database (n=1288). Highest values are in bold. Comparisons significant at the α≤0.05 level between the groups tested with the nonparametric Kruskal-Wallis test for group means and with the median one-way analysis for group medians are indicated by different letters [A, B, etc.].

Variable	Parameter	Soil type at well site (SOIL)			
		OUTCROP n=56 5%	**TILL n=1124 87%**	**COARSE-GR n=68 5%**	**FINE-GR n=40 3%**
DEPTH m	mean median	**99** 91	88 79	95 **96**	82 86
Q Lhr^{-1}	mean median	1118 ^{C} 490 ^{B}	1574 ^{B} 700 ^{B}	1974 ^{AB} 590 ^{AB}	**2400** ^{A} **1200** ^{A}
Q/d_s $Lhr^{-1}m^{-1}$	mean median	19 ^{C} 5 ^{B}	44 ^{B} 12 ^{AB}	49 ^{B} 9 ^{B}	**143** ^{A} **28** ^{A}
Q/s m^2s^{-1}	mean median	$2{,}16\times10^{-5}$ ^{C} $5{,}91\times10^{-6}$ ^{B}	$4{,}94\times10^{-5}$ ^{B} $1{,}40\times10^{-5}$ ^{AB}	$5{,}53\times10^{-5}$ ^{B} $1{,}05\times10^{-5}$ ^{B}	$\mathbf{1{,}56\times10^{-4}}$ ^{A} $\mathbf{3{,}25\times10^{-5}}$ ^{A}
Q_w ms^{-1}	mean median	$4{,}49\times10^{-7}$ ^{C} $7{,}00\times10^{-8}$ ^{C}	$3{,}80\times10^{-6}$ ^{B} $2{,}08\times10^{-7}$ ^{B}	$1{,}84\times10^{-6}$ ^{AB} $1{,}41\times10^{-7}$ ^{BC}	$\mathbf{4{,}99\times10^{-5}}$ ^{A} $\mathbf{5{,}75\times10^{-7}}$ ^{A}
T m^2s^{-1}	mean median	$1{,}12\times10^{-5}$ ^{C} $3{,}16\times10^{-6}$ ^{B}	$2{,}53\times10^{-5}$ ^{B} $7{,}40\times10^{-6}$ ^{AB}	$2{,}83\times10^{-5}$ ^{B} $5{,}59\times10^{-6}$ ^{B}	$\mathbf{7{,}79\times10^{-5}}$ ^{A} $\mathbf{1{,}70\times10^{-5}}$ ^{A}
K ms^{-1}	mean median	$2{,}33\times10^{-7}$ ^{C} $3{,}70\times10^{-8}$ ^{C}	$1{,}91\times10^{-6}$ ^{B} $1{,}10\times10^{-7}$ ^{B}	$9{,}38\times10^{-7}$ ^{AB} $7{,}50\times10^{-8}$ ^{BC}	$\mathbf{2{,}45\times10^{-5}}$ ^{A} $\mathbf{3{,}01\times10^{-7}}$ ^{A}

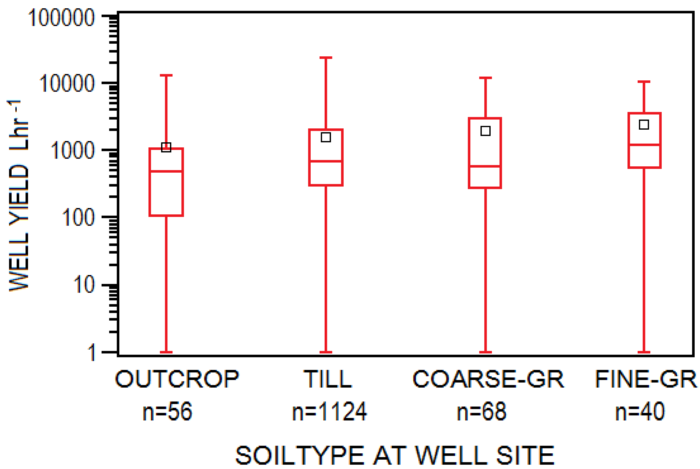


Fig. 67. Box-plot of well yield vs. soil type for all drilled wells in the CF database (n=1288). The maximum and minimum values and the 25th and 75th percentiles of the yield are presented with tops and bottoms on the whiskers and on the boxes, respectively. Median is presented with a horizontal line in the boxes and the mean with a square. The yield of a dry well has been entered into calculations as a nominal figure of 1 Lhr^{-1}.

The thickness of overburden has highly significant, though low, positive correlations with well yield and hydraulic parameters (Table 41).

The overburden is at its thickest in coarse-grained (Md 6,5 m) and fine-grained well sites (Md 6,0 m). Till areas have the median overburden thickness of 3 meters. In outcrops the soil mantle is at its thinnest (Md 1,0 m). This may partly explain the higher well yields and hydraulic values in fine-grained well sites.

Table 41. Spearman rank correlation coefficients (r) and significance levels (p) between overburden thickness and well depth, yield and hydraulic parameters for all drilled wells in the CF database (n=1058).

		DEPTH	Q	Q/d$_S$	Q/s	Q$_W$	T	K
OVER	r	0,04633	0,10334	0,10649	0,10649	0,09829	0,10649	0,09824
	p	0,1320	0,0008	0,0005	0,0005	0,0014	0,0005	0,0014

The wells were divided into four groups according to the thickness of the overburden at the well site (OVER; 0-1, 2-5, 6-10, >10 m) for nonparametric Kruskal-Wallis and median tests. Well yields and hydraulic parameters are at their highest when the thickness of the overburden is more than 10 meters. At their lowest they are when the overburden layer is thin or absent (Table 42, Fig. 68).

Table 42. Statistics for well depth, yield and hydraulic parameters in different groups of overburden thickness of all drilled wells in the CF database (n=1058). Highest values are in bold. Comparisons significant at the α≤0.05 level between the groups tested with the nonparametric Kruskal-Wallis test for group means and with the median one-way analysis for group medians are indicated by different letters $^{A, B, etc.}$.

Variable	Parameter	Overburden thickness at well site (OVER) 0-1 m n=270 26%	2-5 m n=435 41%	6-10 m n=179 17%	> 10 m n=174 16%
DEPTH m	mean median	95 ^{AB} 85 ^{A}	81 ^{C} 74 ^{B}	92 ^{B} 83 ^{AB}	**103** ^{A} **91** $^{\mathbf{A}}$
Q Lhr^{-1}	mean median	1324 ^{B} 600 ^{B}	1467 ^{AB} 625 ^{AB}	1848 ^{A} 700 ^{AB}	**2269** $^{\mathbf{A}}$ **1000** $^{\mathbf{A}}$
Q/d$_S$ $Lhr^{-1}m^{-1}$	mean median	32 ^{B} 8 ^{B}	38 ^{A} 12 ^{A}	50 ^{A} 10 ^{B}	**80** ^{A} **16** $^{\mathbf{A}}$
Q/s m^2s^{-1}	mean median	$3,64x10^{-5}$ ^{B} $9,07x10^{-6}$ ^{C}	$4,31x10^{-5}$ ^{A} $1,41x10^{-5}$ ^{AB}	$5,57x10^{-5}$ ^{A} $1,19x10^{-5}$ ^{BC}	$\mathbf{8,87x10^{-5}}$ ^{A} $\mathbf{1,84x10^{-5}}$ $^{\mathbf{A}}$
Q$_W$ ms^{-1}	mean median	$1,56x10^{-6}$ ^{B} $1,35x10^{-7}$ ^{B}	$2,04x10^{-6}$ ^{A} $2,05x10^{-7}$ ^{A}	$3,59x10^{-6}$ ^{AB} $1,96x10^{-7}$ ^{AB}	$\mathbf{1,43x10^{-5}}$ $^{\mathbf{A}}$ $\mathbf{2,64x10^{-7}}$ $^{\mathbf{AB}}$
T m^2s^{-1}	mean median	$1,87x10^{-5}$ ^{B} $4,83x10^{-6}$ ^{C}	$2,22x10^{-5}$ ^{A} $7,46x10^{-6}$ ^{AB}	$2,85x10^{-5}$ ^{A} $6,33x10^{-6}$ ^{BC}	$\mathbf{4,48x10^{-5}}$ ^{A} $\mathbf{9,69x10^{-5}}$ $^{\mathbf{A}}$
K ms^{-1}	mean median	$7,98x10^{-7}$ ^{B} $7,10x10^{-8}$ ^{B}	$1,04x10^{-6}$ ^{A} $1,08x10^{-7}$ ^{A}	$1,81x10^{-6}$ ^{AB} $1,04x10^{-7}$ ^{AB}	$\mathbf{7,04x10^{-6}}$ $^{\mathbf{A}}$ $\mathbf{1,38x10^{-7}}$ $^{\mathbf{AB}}$

According to Krásný (2002), the presence of Quaternary deposits at well sites results in higher transmissivities in bedrock wells. In western Australia the most successful borehole sites have been found where there is a thick eluvial soil profile (Allen & Davidson 1982). In Sweden, Olsson (1980) found that borehole yields increase rapidly with increasing thickness of the overburden. Regression analysis for 1,045 drilled wells gave a log-linear relation between the well yield and soil thickness with a high coefficient of determination (R^2=0,792). In Central Finland, these parameters correlate positively highly significantly with each other (Table 41). However, the coefficient of determination of the corresponding log-linear regression is very low (R^2=0,013, n=1058).

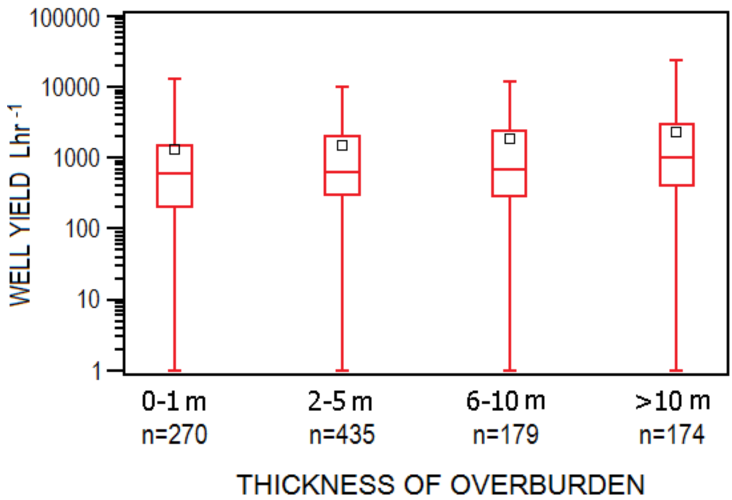


Fig. 68. Box-plot of well yield vs. thickness of overburden for all drilled wells in the CF database (n=1058). The maximum and minimum values and the 25th and 75th percentiles of the yield are presented with tops and bottoms on the whiskers and on the boxes, respectively. Median is presented with a horizontal line in the boxes and the mean with a square. The yield of a dry well has been entered into calculations as a nominal figure of 1 Lhr^{-1}.

Uhl et al. (1979) and Johansson (2005) found no correlation between overburden thickness and well yield in crystalline rock areas of India and Sri Lanka, respectively. Neither found Neves and Morales (2007a) direct influence of overburden thickness (sedimentary coverings or weathered layer) on well productivity in crystalline terrains of southeastern Brazil. Tennakoon (1990) found a negative correlation between the thickness of the overburden and the well yield in Sri Lanka. Lewis (1990) has observed that there is a negative correlation between the saturated thickness of the regolith and the specific capacity of 145 boreholes in basement aquifers of Malawi. According to Mailu (1990), boreholes that penetrate overburden aquifers have lower yields than those, which penetrate the fractured aquifers in metamorphic rocks of Kenya. Holland and Witthüser (2009, 2011) state that the thickness of weathering does not appear to be a major controlling factor on a regional scale, especially in areas of high borehole productivity. According to Poth (1968) and Henry (1992), the depth of weathering does not affect the well yield, but the weathered zone is an important storage reservoir.

In order to find out if there are any differences in well production properties between the wells located in recharge and discharge areas the wells were roughly divided into two groups as follows: recharge area wells (GWT>OVER), discharge area wells (GWT≤OVER). According to this division, two thirds of the private wells in the CF database are situated in recharge areas (n=467) and the rest in discharge areas (n=235). The mean and median well yields and hydraulic values are significantly higher in the discharge well group than in the recharge group (Table 43). Jammallo (1984) has made similar observations in Richmond area in Vermont, United States.

Table 43. Statistics for well depth, yield and hydraulic parameters of the recharge and discharge area wells in the CF database (private drilled wells n=460). Highest values are in bold. Comparisons significant at the α≤0.05 level between the groups tested with the nonparametric Kruskal-Wallis test for group means and with the median one-way analysis for group medians are indicated by different letters [A, B].

Variable	Parameter	RECHARGE AREA WELLS n=281 61%	DISCHARGE AREA WELLS n=179 39%
DEPTH m	mean median	**87** 75	82 **76**
Q Lhr^{-1}	mean median	1437 ^{B} 600 ^{B}	**2187** ^{A} **1000** ^{A}
Q/d_s $Lhr^{-1}m^{-1}$	mean median	41 ^{B} 11 ^{B}	**84** ^{A} **16** ^{A}
Q/s m^2s^{-1}	mean median	$4{,}58x10^{-5\ B}$ $1{,}32x10^{-5\ B}$	$\mathbf{9{,}25x10^{-5}}\ ^{A}$ $\mathbf{1{,}84x10^{-5}}\ ^{A}$
Q_w ms^{-1}	mean median	$2{,}79x10^{-6\ B}$ $1{,}96x10^{-7\ B}$	$\mathbf{1{,}39x10^{-5}}\ ^{A}$ $\mathbf{3{,}43x10^{-7}}\ ^{A}$
T m^2s^{-1}	mean median	$2{,}35x10^{-5\ B}$ $7{,}00x10^{-6\ B}$	$\mathbf{4{,}67x10^{-5}}\ ^{A}$ $\mathbf{9{,}70x10^{-6}}\ ^{A}$
K ms^{-1}	mean median	$1{,}42x10^{-6\ B}$ $1{,}04x10^{-7\ B}$	$\mathbf{6{,}88x10^{-6}}\ ^{A}$ $\mathbf{1{,}82x10^{-7}}\ ^{A}$

Drilled wells in the CF database are evenly distributed in relation to different rock types in the study area (Table 44). This indicates that the database wells are a representative sample of drilled wells of Central Finland. Most wells (74%) are situated in felsic intrusive rocks in granite (32%) and granodiorite (42%) areas. Nearly 8% of the wells are located in bedrock areas composed of mica gneiss and schist. Some 5% of the wells are drilled into quartz monzonite rocks. Mafic intrusive rocks and subvolcanic rocks are rare in the study area and only 1,3% of the database wells are found in these rock types. Around 17% of the wells in Central Finland have been drilled into postkinematic rocks.

For statistical examinations the database wells were divided into seven groups of rock types (Table 8): granites GR (code 1), porphyritic granites PGR (2-3), granodiorites GRDR (4), porphyritic granodiorites PGRDR (5), other intrusive rocks OIR (6-14), volcanic and subvolcanic rocks VR (15-20), and metasediments MS (21-22). The median yield and hydraulic values are at their highest in wells drilled in metasediments and at their lowest in porphyritic granodiorites (Table 45). However, none of the yield and hydraulic parameter differences between rock types are statistically significant. Actually, the within-group variation of the well production properties is several orders of magnitude larger than the between-group variation (Fig. 69), which has been confirmed in other crystalline rock areas, too (e.g. Uhl & Sharma 1978, Tennakoon 1987, Brook 1988, Hazell et al. 1992, Wright 1992, Banks et al. 1993, 1994, 1996, Henriksen & Kyrkjeeide 1993, Krásný 1993b, 2002, Kastrinos & Wilkinson 1994, Filho & Rebouças 1995, Henriksen 1995, Morland 1997, Havlík & Krásný 1998, Lloyd 1999b, Sander 1999, Johansson 2005, Neves & Morales 2007a, 2007b). The well production properties of postkinematic rocks do not differ statistically from other rocks.

Table 44. The distribution of all drilled wells in the CF database in different rock types (n=2479).

Code	Rock type	Abbre-viation	Area km²	Area %	Area Cum. %	Number of wells Freq.	Number of wells %	Number of wells Cum. %
1	Granite	Gr	2555,4	12,8	12,8	279	11,3	11,3
2	Porphyritic granite	PGr	3823,8	19,2	32,0	501	20,2	31,5
3	Pyroxene-bearing granite	PxGr	42,2	0,2	32,2	9	0,4	31,8
4	Granodiorite	GrDr	6511,8	32,6	64,8	768	31,0	62,8
5	Porphyritic granodiorite	PGrDr	2782,2	13,9	78,8	266	10,7	73,5
6	Tonalite	Ton	687,1	3,4	82,2	85	3,4	77,0
7	Quartz monzonite	QMZ	21,3	0,1	82,3	4	0,2	77,1
8	Porphyritic quartz monzonite	PQMZ	526,2	2,6	85,0	123	5,0	82,1
9	Quartz monzodiorite	QMDr	79,8	0,4	85,4	11	0,4	82,5
10	Quartz diorite	QDr	247,9	1,2	86,6	60	2,4	85,0
11	Monzodiorite	MDr	5,7	0,03	86,6	0	0,0	85,0
12	Diorite	Dr	70,7	0,4	87,0	16	0,7	85,6
13	Gabbro	Gb	76,7	0,4	87,4	12	0,5	86,1
14	Ultramafic intrusive rock	UM	1,5	0,007	87,4	0	0,0	86,1
15	Intermediate subvolcanic rock	ISv	142,9	0,7	88,1	8	0,3	86,4
16	Felsic subvolcanic rock	FSv	108,2	0,5	88,6	13	0,5	86,9
17	Mafic volcanic rock	MV	135,5	0,7	89,3	14	0,6	87,5
18	Intermediate volcanic rock	IV	515,0	2,6	91,9	91	3,7	91,2
19	Felsic volcanic rock	FV	135,3	0,7	92,6	19	0,8	91,9
20	Quartz-feldspar schist/gneiss	QF	147,3	0,7	93,3	11	0,4	92,4
21	Mica gneiss	MG	1218,2	6,1	99,4	153	6,2	98,6
22	Mica schist	MS	116,2	0,6	100,0	36	1,5	100,0
∑ 22			**19951,7**			**2479**		

Table 45. Statistics for well depth, yield and hydraulic parameters in different rock type groups of all drilled wells in the CF database (n=1288). Highest values are in bold. Differences between the groups, which were tested with the nonparametric Kruskal-Wallis test for group means and with the median one-way analysis for group medians, are statistically not significant (α≤0.05).

Variable	Para-meter	GR n=145 11%	PGR n=247 19%	GRDR n=426 33%	PGRDR n=139 11%	OIR n=144 11%	VR n=87 7%	MS n=100 8%
DEPTH m	mean	88	87	91	89	**92**	79	88
	median	80	80	79	82	**89**	69	82
Q Lhr^{-1}	mean	1644	1552	1480	1723	**1851**	1477	1754
	median	700	800	670	500	650	700	**832**
Q/d_s $Lhr^{-1}m^{-1}$	mean	**56**	46	44	39	39	32	35
	median	12	12	12	9	11	13	**17**
Q/s m^2s^{-1}	mean	$6,19x10^{-5}$	$5,21x10^{-5}$	$4,97x10^{-5}$	**$7,37x10^{-5}$**	$4,42x10^{-5}$	$3,58x10^{-5}$	$3,99x10^{-5}$
	median	$1,38x10^{-5}$	$1,37x10^{-5}$	$1,39x10^{-5}$	$1,02x10^{-5}$	$1,25x10^{-5}$	$1,56x10^{-5}$	**$1,91x10^{-5}$**
Q_w ms^{-1}	mean	$3,78x10^{-6}$	$3,38x10^{-6}$	$6,41x10^{-6}$	**$1,37x10^{-5}$**	$1,44x10^{-6}$	$9,65x10^{-7}$	$1,10x10^{-6}$
	median	$2,01x10^{-7}$	$2,02x10^{-7}$	$2,02x10^{-7}$	$1,53x10^{-7}$	$1,87x10^{-7}$	$2,68x10^{-7}$	**$3,18x10^{-7}$**
T m^2s^{-1}	mean	$3,14x10^{-5}$	$2,67x10^{-5}$	$2,54x10^{-5}$	**$3,72x10^{-5}$**	$2,27x10^{-5}$	$1,85x10^{-5}$	$2,06x10^{-5}$
	median	$7,29x10^{-6}$	$7,27x10^{-6}$	$7,37x10^{-6}$	$5,39x10^{-6}$	$6,60x10^{-6}$	$8,23x10^{-6}$	**$1,01x10^{-5}$**
K ms^{-1}	mean	$1,90x10^{-6}$	$1,71x10^{-6}$	$3,18x10^{-6}$	**$6,76x10^{-6}$**	$7,39x10^{-7}$	$4,98x10^{-7}$	$5,67x10^{-7}$
	median	$1,06x10^{-7}$	$1,08x10^{-7}$	$1,07x10^{-7}$	$8,10x10^{-8}$	$9,90x10^{-8}$	$1,42x10^{-7}$	**$1,68x10^{-7}$**

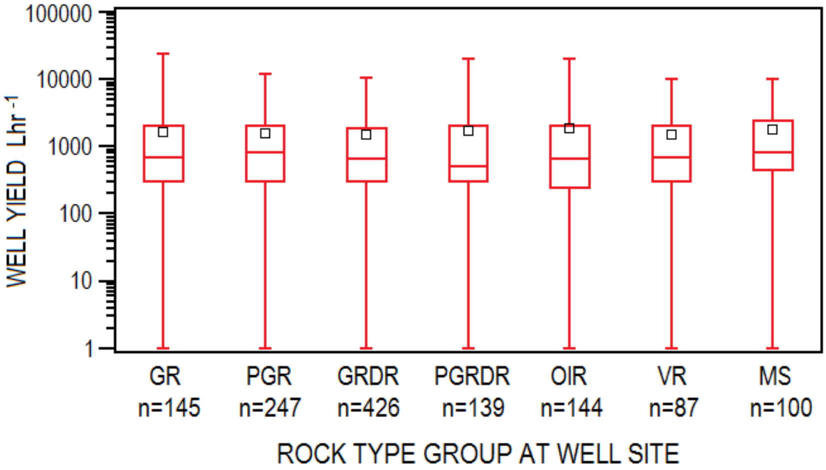


Fig. 69. Box-plot of well yield vs. rock type group for all drilled wells in the CF database (n=1288). The maximum and minimum values and the 25th and 75th percentiles of the yield are presented with tops and bottoms on the whiskers and on the boxes, respectively. Median is presented with a horizontal line in the boxes and the mean with a square. The yield of a dry well has been entered into calculations as a nominal figure of 1 Lhr^{-1}.

The distribution of low-yield and high-yield drilled wells in different rock type groups is shown in Fig. 70.

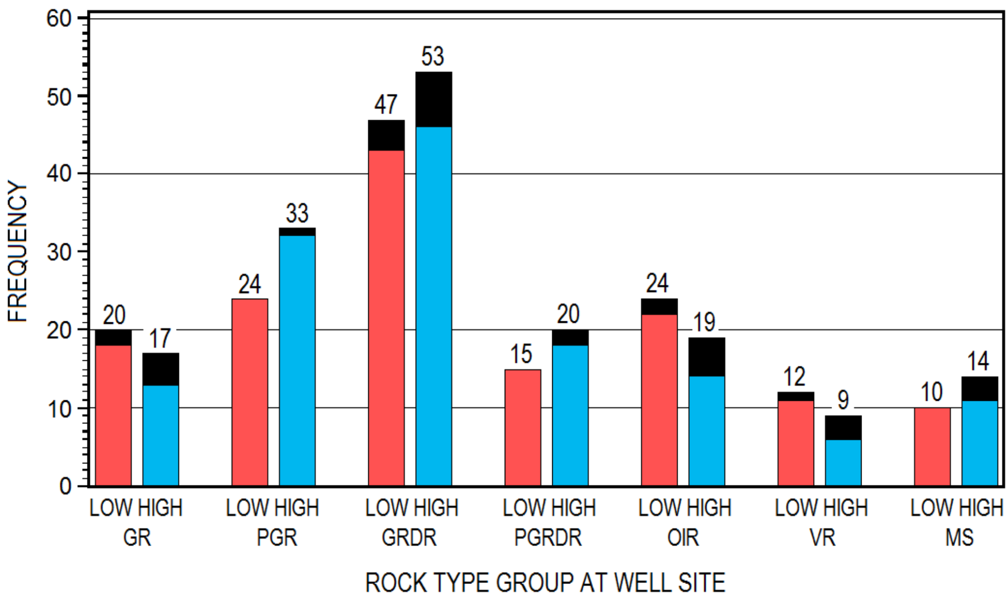


Fig. 70. Bar chart showing the distribution of low-yield ($Q \leq 100$ Lhr^{-1}) and high-yield ($Q \geq 4{,}000$ Lhr^{-1}) drilled wells in the CF database in different rock type groups. The number of wells per bar is indicated above each bar (n=317, test wells marked in black).

In Sweden, the yield of drilled wells is around 60-80% higher in granitic rock types than in metamorphosed rock types. However, the variation within one rock type can be much larger (Larsson 1987, Olofsson et al. 2001). Wladis (1995) found no clear evidence about the influence of lithology on well yield in southwestern Sweden. Yet a strong correlation between the silica content of rocks and well capacity was established (Wladis 1995).

A case study carried out by Rohr-Torp (1987, 2003) in a hard rock area south of Oslo, Norway, revealed differences in yield of wells drilled into different rock types. Amphibolic and gabbroic rocks had the lowest mean yield per drilled meters (8,9 $Lhr^{-1}m^{-1}$); granitic to quartz-dioritic biotite gneiss had the highest yield (28,1 $Lhr^{-1}m^{-1}$). Also Henriksen (1995, 2006b) and Morland (1997) found that well yields are dependent on lithology of the crystalline rocks in Norway. The lithologies yielding most groundwater per drilled meter were the metamorphosed rhyolites and the Permian volcanic, sedimentary and plutonic rocks (20-70 $Lhr^{-1}m^{-1}$). The lithologies having the lowest groundwater yield per drilled meter were the gabbroic and ultramafic rocks and the quartz diorite (2-3 $Lhr^{-1}m^{-1}$; Morland 1997). However, as Henriksen (1995) states, the within-group variation of yield was several orders of magnitude larger than the between-group variation.

According to Florquist et al. (1973), experience has shown that pegmatite dikes are by far the most reliable and productive geologic features in crystalline rocks of Colorado, USA. The next most productive features in granitic rock are faults. In Ghana, boreholes that intercept Precambrian quartz-veins and pegmatites have significant yields (> 8,000 Lhr^{-1}; Agyekum & Dapaah-Siakwan 2008). Sutphin et al. (2001) found that granite and metagranite rocks had significantly higher yield than other metaigneous rock types in the Pinardville area of Loudoun County, Virginia, USA. In crystalline rocks of the Piedmont and Blue Ridge Provinces of North Carolina, USA, wells in the most productive bedrock/hydrogeologic units had average yields twice those of wells in least-productive units (Daniel 1987, 1989). In New Hampshire, USA, amphibolite gneiss, layered migmatite and tonalite are zones of high well productivity. Instead, foliated plutons tend to have lower than average yields (Moore et al. 2002a). According to Knopman (1990), lithology is an important factor in explaining the well yield in the Piedmont and Valley and Ridge provinces of Pennsylvania, USA.

According to Gustafson and Krásný (1994), the general rule seems to be that acid brittle rocks have a higher hydraulic conductivity with a larger variance than basic rocks. In the crystalline basement of Black Forest area, Germany, Stober (1996) identified slightly higher hydraulic conductivities in granites (9,6x10^{-7} ms^{-1}) than in gneisses (5,0x10^{-8} ms^{-1}). On the other hand, Krásný (1996b) and Havlík and Krásný (1998) state that greater than average transmissivity was expected in areas of basic igneous rocks and some types of granites in the Bohemian Massif of the Czech Republic (see also Chambel et al. 2003). In addition, the decrease of permeability in granitic rocks and orthogneisses with depth was evidently faster than in metasedimentary rocks in general (Havlík & Krásný 1998). However, the importance of petrology on different transmissivity distributions of hard rocks, with crystalline limestones as well known exceptions, generally proved to be insignificant on a regional scale (Havlík & Krásný 1998, Krásný 2002).

The land uplift rate (UPLIFT) of the postglacial rebound is 4,8 mm a^{-1} in Central Finland on the average; the range is only 2,7 mm a^{-1} (Table 46). The highest shore level (HSL) at well sites varies between 130 and 201 m a.s.l.; the average HSL is slightly over 150 m a.s.l. The UPLIFT and HSL correlate positively very highly significantly with each other (Fig. 71).

Table 46. Statistics for UPLIFT and HSL of all drilled wells in the CF database. The explanations of variable abbreviations are given in Table 3 and in Chapter 6.3.3.

Variable	n	Mean	Std dev	Median	Minimum	Maximum	Range
UPLIFT mm a^{-1}	2395	4,8	0,4	4,8	3,7	6,4	2,7
HSL m a.s.l.	2395	154	13	151	130	201	71

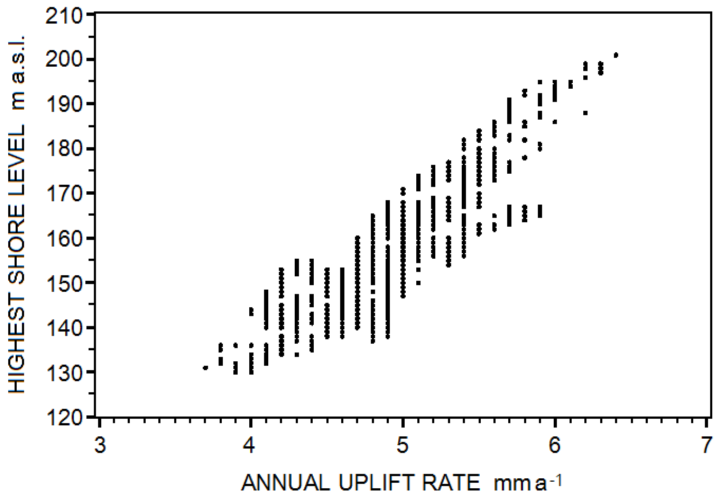


Fig. 71. Highest shore level (m a.s.l.) vs. annual uplift rate (mm a^{-1}) at drilled well sites in the CF database (n=2395).

The uplift rate does not correlate statistically significantly with the well yield and hydraulic parameters (Table 47). The HSL has statistically significant, though low, correlations with the well yield and hydraulic parameters. The well depth correlates negatively significantly with UPLIFT and HSL.

Table 47. Spearman rank correlation coefficients (r) and their significance levels (p) between UPLIFT and HSL and well depth, yield and hydraulic parameters for all drilled wells in the CF database (n=1288).

		DEPTH	Q	Q/d_s	Q/s	Q_w	T	K
UPLIFT	r	-0,08571	-0,01268	0,02393	0,02393	0,04990	0,02393	0,05009
	p	0,0021	0,6492	0,3908	0,3908	0,0734	0,3908	0,0723
HSL	r	-0,16931	0,06341	0,11448	0,11448	0,14450	0,11448	0,14462
	p	<0,0001	0,0229	<0,0001	<0,0001	<0,0001	<0,0001	<0,0001

The wells were divided into six groups according to the rate of annual uplift (UPLIFT; <4,0, 4,0-4,4, 4,5-4,9, 5,0-5,4, 5,5-5,9 and ≥6,0 mm a^{-1}) for nonparametric Kruskal-Wallis and median tests. The uplift rate does not indicate any clear trend in well yields and hydraulic properties (Table 48, Fig. 72).

Table 48. Statistics for well depth, yield and hydraulic parameters in different groups of uplift rate of all drilled wells in the CF database (n=1288). Highest values are in bold. Comparisons significant at the α≤0.05 level between the groups tested with the nonparametric Kruskal-Wallis test for group means and with the median one-way analysis for group medians are indicated by different letters $^{A, B, etc.}$.

Variable	Para meter	Uplift rate at well site (UPLIFT) < 4,0 mma^{-1} n=16 1%	4,0-4,4 mma^{-1} n=192 15%	4,5-4,9 mma^{-1} n=703 55%	5,0-5,4 mma^{-1} n=275 21%	5,5-5,9 mma^{-1} n=89 7%	≥6,0 mma^{-1} n=13 1%
DEPTH m	mean median	**97** ^{A} **99** ^{A}	84 ^{C} 73 ^{C}	95 ^{AB} 89 ^{AB}	79 ^{C} 70 ^{C}	78 ^{C} 62 ^{D}	73 ^{BC} 65 ^{BC}
Q Lhr^{-1}	mean median	1330 ^{AB} 525 ^{AB}	1761 ^{A} **1000** ^{A}	1431 ^{B} 600 ^{B}	**1872** ^{A} 800 ^{A}	1837 ^{A} **1000** ^{A}	1387 ^{AB} **1000** ^{AB}
Q/d_s Lhr^{-1}m^{-1}	mean median	18 ^{BC} 10 ^{ABC}	42 ^{AC} **18** ^{A}	42 ^{B} 8 ^{C}	**59** ^{AC} 13 ^{B}	58 ^{A} **18** ^{A}	22 ^{ABC} **18** ^{AC}
Q/s m^{2}s^{-1}	mean median	2,06x10^{-5} ^{BC} 1,20x10^{-5} ^{ABC}	4,80x10^{-5} ^{AC} 2,04x10^{-5} ^{A}	4,67x10^{-5} ^{B} 9,75x10^{-6} ^{C}	**6,63x10^{-5}** ^{AC} 1,53x10^{-5} ^{B}	6,50x10^{-5} ^{A} **2,10x10^{-5}** ^{A}	2,57x10^{-5} ^{ABC} 2,04x10^{-5} ^{AC}
Q_w ms^{-1}	mean median	5,57x10^{-7} ^{CD} 1,26x10^{-7} ^{CD}	2,36x10^{-6} ^{A} 3,25x10^{-7} ^{AC}	**6,41x10^{-6}** ^{BC} 1,47x10^{-7} ^{D}	4,24x10^{-6} ^{AD} 2,68x10^{-7} ^{C}	3,09x10^{-6} ^{A} **4,84x10^{-7}** ^{A}	5,56x10^{-7} ^{AB} 4,44x10^{-7} ^{A}
T m^{2}s^{-1}	mean median	1,07x10^{-5} ^{BC} 6,34x10^{-6} ^{ABC}	2,47x10^{-5} ^{AC} 1,07x10^{-5} ^{A}	2,38x10^{-5} ^{B} 5,18x10^{-6} ^{C}	**3,38x10^{-5}** ^{AC} 8,08x10^{-6} ^{B}	3,33x10^{-5} ^{A} **1,11x10^{-5}** ^{A}	1,34x10^{-5} ^{ABC} 1,07x10^{-5} ^{AC}
K ms^{-1}	mean median	2,90x10^{-7} ^{CD} 6,70x10^{-8} ^{CD}	1,21x10^{-6} ^{A} 1,71x10^{-7} ^{AC}	**3,18x10^{-6}** ^{BC} 7,80x10^{-8} ^{D}	2,13x10^{-6} ^{AD} 1,41x10^{-7} ^{C}	1,57x10^{-6} ^{A} **2,56x10^{-7}** ^{A}	2,91x10^{-7} ^{AB} 2,32x10^{-7} ^{A}

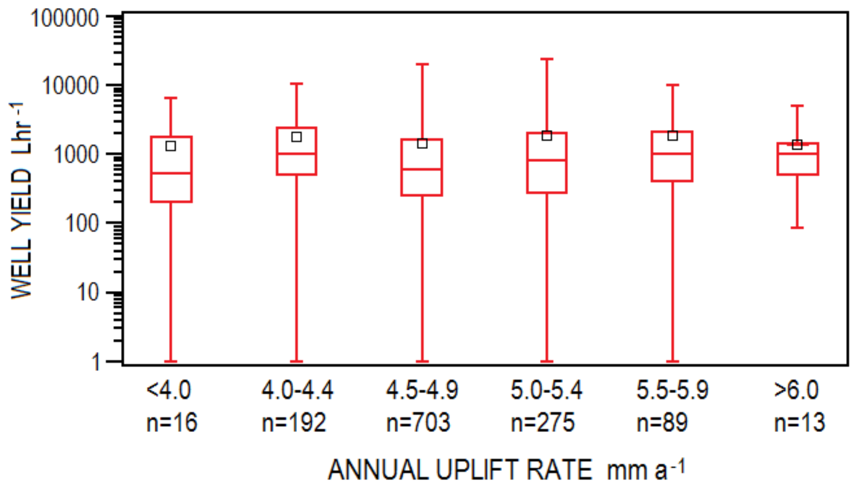


Fig. 72. Box-plot of well yield vs. uplift rate for all drilled wells in the CF database (n=1288). The maximum and minimum values and the 25th and 75th percentiles of the yield are presented with tops and bottoms on the whiskers and on the boxes, respectively. Median is presented with a horizontal line in the boxes and the mean with a square. The yield of a dry well has been entered into calculations as a nominal figure of 1 Lhr^{-1}.

In northwestern Central Finland, where the uplift rate is over 5,4 mm a^{-1} (Fig. 29), the relative number of cable tool wells is twice that of the rest of Central Finland. As it has been described earlier, the yield of cable tool wells is higher than in DTH wells, which in turn are deeper than cable tool wells. This explains, at least for the most part, the higher yield and hydraulic values in the uplift groups 5,5-5,9 and >5,9 mm a^{-1}.

In Norway, Rohr-Torp (1994a, 1995) and Morland (1997) have suggested a positive relation between well yield and land uplift rate. According to them, for each mm increase in annual uplift in any rock type, 100 –130 Lhr^{-1} can be added to the total yield at the same time as the depth required to achieve this yield is reduced by 6,0-6,5 meters.

The CF database wells were divided into six groups according to the highest shore level of the Baltic Sea at well site (HSL; ≤ 140, 141-150, 151-160, 161-170, 171-180 and >180 m a.s.l.) for nonparametric Kruskal-Wallis and median tests. The results are similar with the UPLIFT results (Table 49 and Fig. 73).

In Norway, Morland (1997) was not able to statistically confirm or reject a relationship between groundwater yield and location of bedrock boreholes in relation to the postglacial Marine Limit (ML).

Table 49. Statistics for well depth, yield and hydraulic parameters in different groups of HSL of all drilled wells in the CF database (n=1288). Highest values are in bold. Comparisons significant at the α≤0.05 level between the groups tested with the nonparametric Kruskal-Wallis test for group means and with the median one-way analysis for group medians are indicated by different letters $^{A, B, etc.}$.

Variable	Parameter	Highest shore level at well site (HSL)					
		≤ 140 m a.s.l. n=108 8%	**141-150 m a.s.l. n=522 40%**	**151-160 m a.s.l. n=343 27%**	**161-170 m a.s.l. n=129 10%**	**171-180 m a.s.l. n=150 12%**	**> 180 m a.s.l. n=36 3%**
DEPTH m	mean median	**112** ^{A} **103** ^{A}	93 ^{B} 85 ^{B}	87 ^{B} 81 ^{B}	78 ^{C} 66 ^{C}	73 ^{C} 63 ^{C}	79 ^{BC} 66 ^{BC}
Q Lhr^{-1}	mean median	1083 ^{C} 500 ^{BC}	1523 ^{B} 704 ^{B}	1507 ^{BC} 520 ^{C}	1754 ^{AB} 800 ^{AB}	**2400** ^{A} **1000** ^{A}	1301 ^{ABC} **1000** ^{ABC}
Q/d_s $Lhr^{-1}m^{-1}$	mean median	19 ^{E} 7 ^{C}	38 ^{BD} 12 ^{BC}	51 ^{BC} 10 ^{C}	61 ^{AD} 14 ^{AB}	**75** ^{A} **19** ^{A}	35 ^{ABCD} 17 ^{A}
Q/s m^2s^{-1}	mean median	$2{,}13x10^{-5}$ ^{E} $7{,}75x10^{-6}$ ^{C}	$4{,}28x10^{-5}$ ^{BD} $1{,}35x10^{-5}$ ^{BC}	$5{,}62x10^{-5}$ ^{BC} $1{,}11x10^{-5}$ ^{C}	$6{,}85x10^{-5}$ ^{AD} $1{,}60x10^{-5}$ ^{AB}	**$8{,}35x10^{-5}$** ^{A} **$2{,}20x10^{-5}$** ^{A}	$4{,}00x10^{-5}$ ^{ABCD} $2{,}02x10^{-5}$ ^{A}
Q_w ms^{-1}	mean median	$1{,}24x10^{-6}$ ^{D} $8{,}50x10^{-8}$ ^{D}	$1{,}90x10^{-6}$ ^{C} $1{,}96x10^{-7}$ ^{C}	**$1{,}14x10^{-5}$** ^{BC} $1{,}64x10^{-7}$ ^{C}	$5{,}50x10^{-6}$ ^{A} $2{,}90x10^{-7}$ ^{BC}	$4{,}13x10^{-6}$ ^{A} $5{,}41x10^{-7}$ ^{A}	$1{,}83x10^{-6}$ ^{ABC} $3{,}59x10^{-7}$ ^{AB}
T m^2s^{-1}	mean median	$1{,}11x10^{-5}$ ^{E} $4{,}13x10^{-6}$ ^{C}	$2{,}20x10^{-5}$ ^{BD} $7{,}15x10^{-6}$ ^{BC}	$2{,}85x10^{-5}$ ^{BC} $5{,}90x10^{-6}$ ^{C}	$3{,}49x10^{-5}$ ^{AD} $8{,}46x10^{-6}$ ^{AB}	**$4{,}24x10^{-5}$** ^{A} **$1{,}16x10^{-5}$** ^{A}	$2{,}06x10^{-5}$ ^{ABCD} $1{,}06x10^{-5}$ ^{A}
K ms^{-1}	mean median	$6{,}37x10^{-7}$ ^{D} $4{,}50x10^{-8}$ ^{D}	$9{,}70x10^{-7}$ ^{C} $1{,}04x10^{-7}$ ^{C}	**$5{,}60x10^{-6}$** ^{BC} $8{,}70x10^{-8}$ ^{C}	$2{,}77x10^{-6}$ ^{A} $1{,}53x10^{-7}$ ^{BC}	$2{,}08x10^{-6}$ ^{A} **$2{,}84x10^{-7}$** ^{A}	$9{,}33x10^{-7}$ ^{ABC} $1{,}88x10^{-7}$ ^{AB}

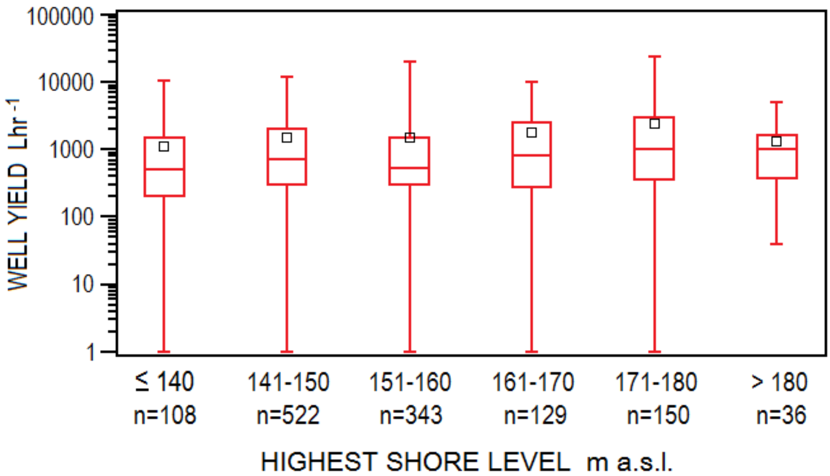


Fig. 73. Box-plot of well yield vs. the highest shore level for all drilled wells in the CF database (n=1288). The maximum and minimum values and the 25th and 75th percentiles of the yield are presented with tops and bottoms on the whiskers and on the boxes, respectively. Median is presented with a horizontal line in the boxes and the mean with a square. The yield of a dry well has been entered into calculations as a nominal figure of 1 Lhr^{-1}.

7.3.3 Topographic factors

Most of the drilled wells (43%) in the CF database (n=2395) are situated on flatlands, while only 11% of them are located on hilltops. Slope wells consist of 30% and valley wells of 16% of the wells.

Topographic setting has an important influence on drilled well depths. Deepest wells are situated on hilltops, second deepest on slopes and the lowest wells in flatlands and valleys (Fig. 74). The average hilltop wells are 24 meters deeper than the valley wells. The median height of the well sites decreases from hilltops to valleys (Fig. 74).

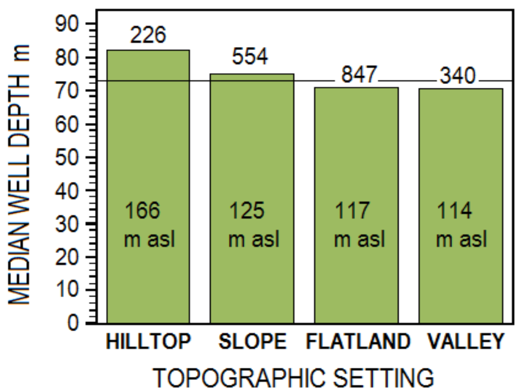


Fig. 74. Bar chart showing the median drilled well depth in different topographic settings in the CF database (n=1967, test wells included). The overall median depth of drilled wells (73 m) is marked with a horizontal line. The median height of the well sites (ASL m a.s.l.) is marked within each bar. The number of wells per bar is indicated above each bar.

Median well yields and hydraulic values, which vary from 2,3- to more than 5-fold between different topographic settings, are at their highest in valley wells and at their lowest in hilltop wells (Figs. 75 and 76, Table 50). For example, the median yield and transmissivity of the valley wells are 1,135 Lhr^{-1} and 1,14x10^{-5} m^{2}s^{-1}, respectively. In hilltops wells the corresponding Q- and T-values are 500 Lhr^{-1} and 3,38x10^{-6} m^{2}s^{-1}, respectively. All the differences between these two topographic settings are statistically significant (Table 50).

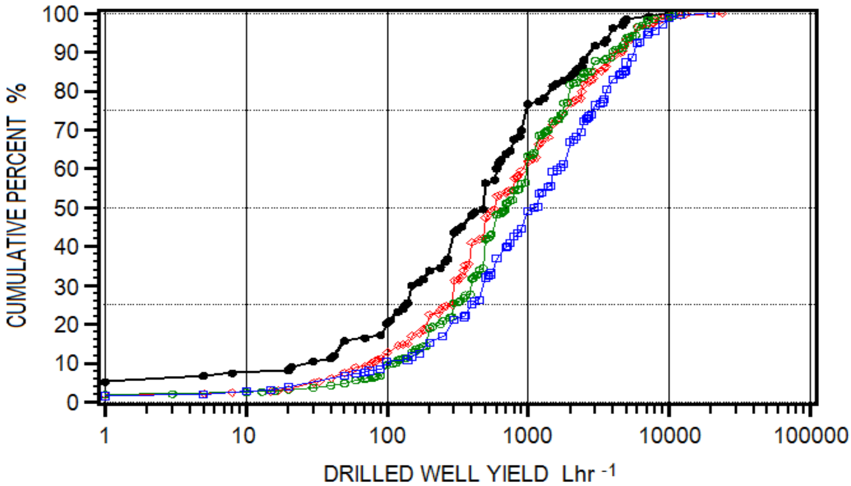


Fig. 75. Cumulative percent frequency of drilled well yields in the CF database in different topographic settings (n=1296, test wells included). Legend: ● = hilltop, ◊ = slope, ○ = flatland, □ = valley.

Table 50. Statistics for well depth, yield and hydraulic parameters in different topographic settings of all drilled wells in the CF database (n=1288). Highest values are in bold. Comparisons significant at the α≤0.05 level between the groups tested with the nonparametric Kruskal-Wallis test for group means and with the median one-way analysis for group medians are indicated by different letters $^{A,\ B,\ etc.}$.

Variable	Parameter	Topographic setting (TOPO)			
		HILLTOP n=133 10%	SLOPE n=358 28%	FLATLAND n=551 43%	VALLEY n=246 19%
DEPTH m	mean	**103** ^{A}	92 ^{B}	87 ^{C}	79 ^{C}
	median	**97** ^{A}	83 ^{B}	79 ^{B}	73 ^{B}
Q Lhr^{-1}	mean	1012 ^{C}	1571 ^{B}	1495 ^{B}	**2200** ^{A}
	median	500 ^{B}	600 ^{B}	700 ^{B}	**1135** ^{A}
Q/d$_{s}$ Lhr^{-1}m^{-1}	mean	25 ^{C}	51 ^{B}	44 ^{B}	**58** ^{A}
	median	5 ^{C}	10 ^{B}	12 ^{B}	**18** ^{A}
Q/s m^{2}s^{-1}	mean	2,80x10$^{-5\ C}$	5,63x10$^{-5\ B}$	4,89x10$^{-5\ B}$	**6,47x10^{-5}** ^{A}
	median	6,01x10$^{-6\ C}$	1,16x10$^{-5\ B}$	1,39x10$^{-5\ B}$	**2,10x10^{-5}** ^{A}
Q$_{w}$ ms^{-1}	mean	1,09x10$^{-6\ C}$	**7,26x10^{-6}** ^{A}	5,08x10$^{-6\ A}$	3,55x10$^{-6\ B}$
	median	6,50x10$^{-8\ C}$	1,74x10$^{-7\ B}$	2,12x10$^{-7\ B}$	**3,65x10^{-7}** ^{A}
T m^{2}s^{-1}	mean	1,44x10$^{-5\ C}$	2,86x10$^{-5\ B}$	2,50x10$^{-5\ B}$	**3,31x10^{-5}** ^{A}
	median	3,22x10$^{-6\ C}$	6,14x10$^{-6\ B}$	7,33x10$^{-6\ B}$	**1,11x10^{-5}** ^{A}
K ms^{-1}	mean	5,60x10$^{-7\ C}$	**3,60x10^{-6}** ^{A}	2,53x10$^{-6\ A}$	1,80x10$^{-6\ B}$
	median	3,50x10$^{-8\ C}$	9,30x10$^{-8\ B}$	1,12x10$^{-7\ B}$	**1,92x10^{-7}** ^{A}

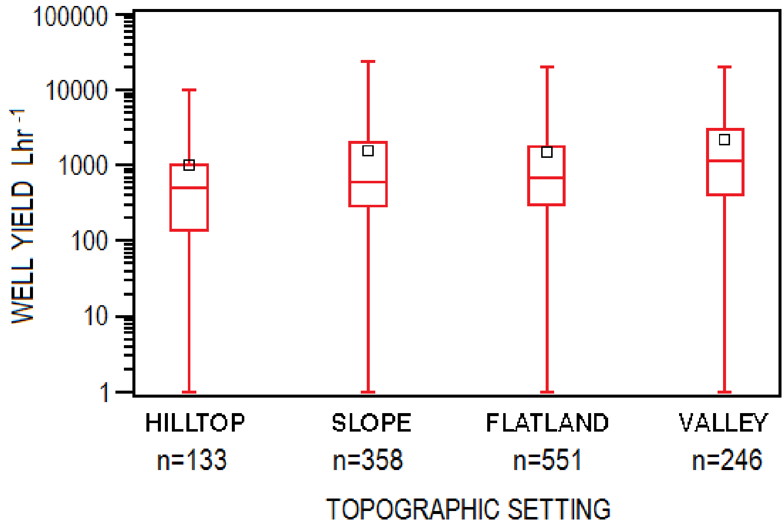


Fig. 76. Box-plot of well yield vs. topographic setting for all drilled wells in the CF database (n=1288). The maximum and minimum values and the 25th and 75th percentiles of the yield are presented with tops and bottoms on the whiskers and on the boxes, respectively. Median is presented with a horizontal line in the boxes and the mean with a square. The yield of a dry well has been entered into calculations as a nominal figure of 1 Lhr^{-1}.

Depth to the first (main) water strike (STR) is at its greatest in hilltop wells, where the median STR (60 m) is almost twice that of the valley wells (35 m; Table 51). The median groundwater table from ground surface is at its deepest in hilltop wells (9 m), whereas in valley wells the median GWT is 4 meters. Half of the 51 flowing artesian water wells in the CF database are situated in valleys whereas on hilltops they are totally missing.

Table 51. Depth of the first (main) water strike (STR/SASL) in different topographic settings (TOPO) for all drilled wells in the CF database (n=807). Highest values are in bold.

TOPO	STR m / SASL m a.s.l.						
	n	Mean	Std dev	Median	Minimum	Maximum	Range
HILLTOP	88	**68 / 94**	**52 / 63**	**60 / 96**	**3 / -51**	237 / **238**	234 / 289
SLOPE	220	58 / 79	46 / 59	45 / 78	1 / -124	**250** / 235	**249 / 359**
FLATLAND	301	55 / 67	40 / 50	42 / 73	2 / -117	200 / 181	198 / 298
VALLEY	198	47 / 72	39 / 49	35 / 79	**3** / -110	230 / 198	227 / 308

The relative proportion of high-yield wells ($Q \geq 4{,}000$ Lhr^{-1}) increases and that of low-yield wells ($Q \leq 100$ Lhr^{-1}) decreases as one moves from hilltops to valleys (Fig. 77). Most of the test wells (70%) have been drilled in valleys; none of them are situated on hilltops.

The median thickness of the overburden is 4,0 meters at flatland well sites, whereas at hilltops the median soil mantle is only half of that. Slope wells and valley wells have median overburden thicknesses of 3 meters. Hence, the thickness of the overburden does not explain the higher yield of valley wells. The relative number of fine-grained well sites is at its highest in flatlands and valleys, whereas at hilltops fine-grained soils are totally missing.

This partly explains the higher yield and hydraulic values of fine-grained well sites (Chapter 7.3.2).

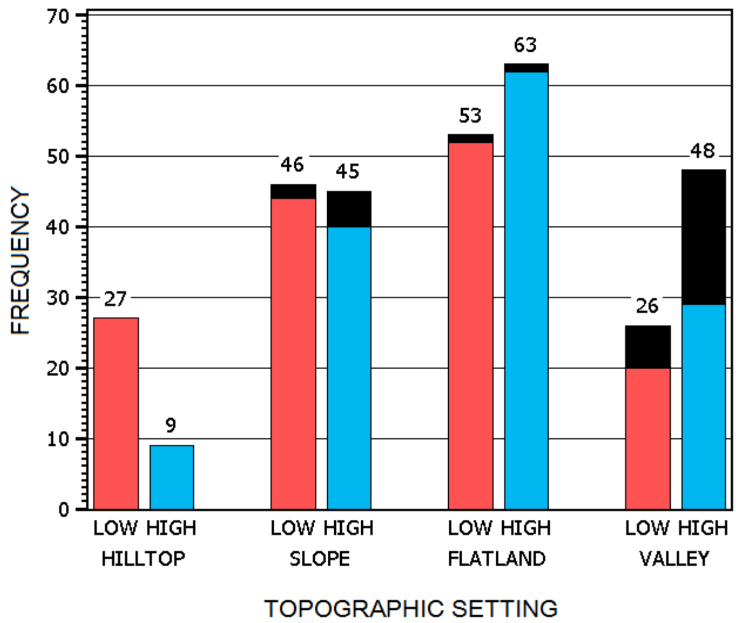


Fig. 77. Bar chart showing the distribution of low-yield ($Q \leq 100$ Lhr^{-1}) and high-yield ($Q \geq 4000$ Lhr^{-1}) drilled wells in the CF database in different topographic settings. The number of wells per bar is indicated above each bar (n=317, test wells marked in black).

The relative height differences (RELA=ASLH-ASLL) in wells´ catchment areas (1 km^2) are typically around 35 meters in Central Finland (Table 52). At their maximum they are a little over 100 meters; in flatland areas they may be less than five meters. The median distance from a well to the nearest (bedrock) hilltop (HDIS) is 290 meters. The peak of the hilltop is typically about 20 meters above the well site. The average rate of change in elevation (SLOH °) between the well site and the hilltop is less than five degrees and at its maximum less than 20 degrees (Table 52).

Table 52. Statistics for topographic factors of all drilled wells in the CF database. Median values are in bold. The explanations of variable abbreviations are given in Table 3 and in Chapter 6.3.4.

Variable		n	Mean	Std dev	Median	Minimum	Maximum	Range
RELA	m	2395	38	19	**35**	4	117	113
RELAH	m	2395	23	17	**19**	0	104	104
RELAL	m	2395	16	11	**13**	0	71	71
RERE	--	2382	2,9	5,0	**1,4**	0	97	97
HDIS	m	957	294	144	**290**	5	560	555
HASL	m a.s.l.	957	160	33	**155**	102	256	154
HAASL	m	957	19	15	**16**	0	94	94
SLOH	°	957	4	2	**3**	0	17	17

The topographic factors RELA, RELAH, HASL and HAASL have only weak correlations with the well yield and hydraulic parameters (Table 53). On the other hand, RELAL, RERE and HDIS correlate highly significantly with well yield and hydraulic parameters. The SLOH does not correlate with well yield and hydraulic values.

Table 53. Spearman rank correlation coefficients (r) with significance levels (p) and number of observations (n) between topographic factors and well depth, yield and hydraulic parameters for all drilled wells in the CF database. The explanations of variable abbreviations are given in Table 3 and in Chapter 6.3.4.

		DEPTH	Q	Q/d_s	Q/s	Q_w	T	K
RELA	r	0,04611	-0,04678	-0,05502	-0,05502	-0,05771	-0,05502	-0,05774
n=1288	p	0,0981	0,0933	0,0483	0,0483	0,0384	0,0483	0,0383
RELAH	r	-0,00528	0,06395	0,05135	0,05135	0,04005	0,05135	0,03997
n=1288	p	0,8500	0,0217	0,0655	0,0655	0,1508	0,0655	0,1517
RELAL	r	0,04340	-0,15026	-0,13782	-0,13782	-0,11990	-0,13782	-0,11976
n=1288	p	0,1195	<0,0001	<0,0001	<0,0001	<0,0001	<0,0001	<0,0001
RERE	r	-0,03023	0,13241	0,11699	0,11699	0,09795	0,11699	0,09781
n=1280	p	0,2799	<0,0001	<0,0001	<0,0001	0,0004	<0,0001	0,0005
HDIS	r	-0,09347	0,13847	0,13960	0,13960	0,13634	0,13960	0,13643
n=506	p	0,0356	0,0018	0,0016	0,0016	0,0021	0,0016	0,0021
HASL	r	0,10501	-0,05564	-0,07694	-0,07694	-0,09379	-0,07694	-0,09431
n=506	p	0,0181	0,2115	0,0838	0,0838	0,0349	0,0838	0,0339
HAASL	r	-0,02728	0,12637	0,11131	0,11131	0,09947	0,11131	0,09935
n=506	p	0,5403	0,0044	0,0122	0,0122	0,0252	0,0122	0,0254
SLOH	r	0,07016	0,03475	0,00497	0,00497	-0,01197	0,00497	-0,01211
n=506	p	0,1150	0,4354	0,9113	0,9113	0,7882	0,9113	0,7859

At a distance below 200 meters from bedrock hilltops the mean and median well yields and hydraulic values are at their lowest. At their highest they are beyond 400 meters from hilltops (Table 54, Fig. 78). The relative number of low-yield wells is at its highest at a distance below 200 meters from hilltops (Fig. 79). Hilltop well sites are mainly composed of intact bedrock whereas in valley well sites the bedrock is often (highly) fractured (Fig. 80). The number and relative proportion of high-yield wells increase when the value of RERE goes over 1.50 (Fig. 81).

Table 54. Statistics for well depth, yield and hydraulic parameters in different HDIS groups of all drilled wells in the CF database (n=506). Highest values are in bold. Comparisons significant at the α≤0.05 level between the groups tested with the nonparametric Kruskal-Wallis test for group means and with the median one-way analysis for group medians are indicated by different letters [A, B, etc.].

Variable	Parameter	Distance to the nearest hilltop (HDIS)			
		< 200 m n=136 27%	200-300 m n=126 25%	301-400 m n=115 23%	> 400 m n=129 25%
DEPTH m	mean median	**102** [A] **91**	93 [AB] **91**	91 [AB] 88	86 [B] 79
Q Lhr^{-1}	mean median	995 [C] 500	1485 [BC] 550	1605 [AB] 708	**1711** [AB] **900**
Q/d_s $Lhr^{-1}m^{-1}$	mean median	22 [B] 7	33 [A] 9	36 [A] **12**	**44** [A] 11
Q/s m^2s^{-1}	mean median	$2{,}49x10^{-5}$ [B] $8{,}05x10^{-6}$	$3{,}78x10^{-5}$ [A] $1{,}06x10^{-5}$	$4{,}03x10^{-5}$ [A] $\mathbf{1{,}42x10^{-5}}$	$\mathbf{4{,}92x10^{-5}}$ [A] $1{,}24x10^{-5}$
Q_w ms^{-1}	mean median	$8{,}84x10^{-7}$ [B] $9{,}00x10^{-8}$	$1{,}43x10^{-6}$ [A] $1{,}77x10^{-7}$	$1{,}66x10^{-6}$ [A] $\mathbf{2{,}39x10^{-7}}$	$\mathbf{2{,}79x10^{-6}}$ [A] $2{,}05x10^{-7}$
T m^2s^{-1}	mean median	$1{,}29x10^{-5}$ [B] $4{,}29x10^{-6}$	$1{,}95x10^{-5}$ [A] $5{,}63x10^{-6}$	$2{,}08x10^{-5}$ [A] $\mathbf{7{,}50x10^{-6}}$	$\mathbf{2{,}52x10^{-5}}$ [A] $6{,}59x10^{-6}$
K ms^{-1}	mean median	$4{,}54x10^{-7}$ [B] $4{,}80x10^{-8}$	$7{,}32x10^{-7}$ [A] $9{,}50x10^{-8}$	$8{,}48x10^{-7}$ [A] $\mathbf{1{,}26x10^{-7}}$	$\mathbf{1{,}41x10^{-6}}$ [A] $1{,}08x10^{-7}$

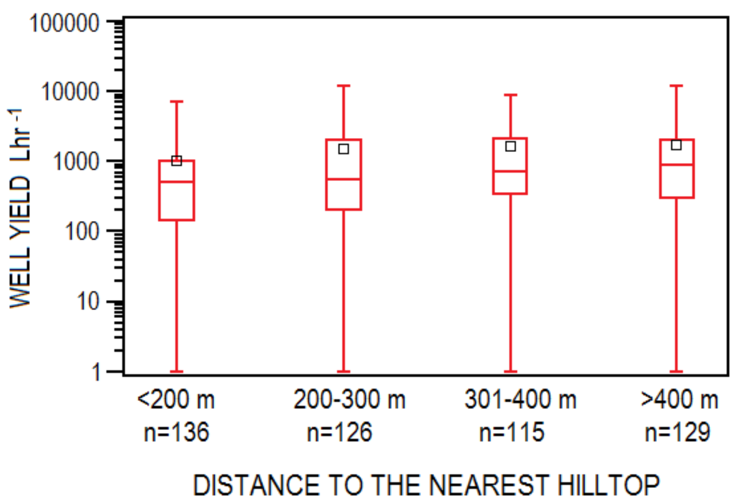


Fig. 78. Box-plot of well yield vs. distance to the nearest hilltop for all drilled wells in the CF database (n=506). The maximum and minimum values and the 25th and 75th percentiles of the yield are presented with tops and bottoms on the whiskers and on the boxes, respectively. Median is presented with a horizontal line in the boxes and the mean with a square. The yield of a dry well has been entered into calculations as a nominal figure of 1 Lhr^{-1}.

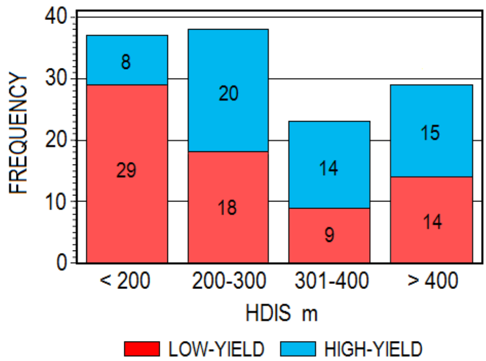


Fig. 79. Bar chart showing the distribution of low-yield (Q≤100 Lhr^{-1}) and high-yield (Q≥4000 Lhr^{-1}) drilled wells in the CF database in different HDIS groups (n=127, test wells included). The number of wells per group is indicated within each bar.

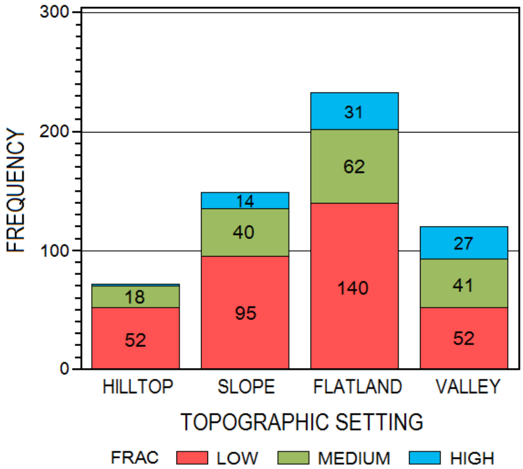


Fig. 80. Bar chart showing the distribution of FRAC in different topographic settings of all drilled wells in the CF database (n=574). The number of wells per FRAC groups is indicated within each bar.

Krásný (1975, 1993b, 1996c, 1997, 2000) has done statistical works on a few hundreds of boreholes drilled in the Bohemian massif, which show that boreholes in drainage areas close to the valley axis are of obviously higher capacity than those on infiltration areas. Transmissivity was estimated from 2 to 4,5 times higher in zones of discharge (valleys) than in recharge zones (slopes and elevations; see also Uhl & Sharma 1978, Chambel et al. 2003). Henriksen (1995) found a positive correlation between high well yields and well sites in low valley sides and valley bottoms in western Norway. Eftimi (2003) suggests that the topographic and geomorphologic position of drilled wells is the controlling factor of the transmissivity and yield of wells. Many drilled wells located in the centre of small valleys at foothills in Albania yield at least 2 to 3 times more water than the wells located at higher elevations (Eftimi 2003).

In the Piedmont and Blue Ridge crystalline rocks of North Carolina, USA, Daniel (1987, 1989) found that wells in draws or valleys have average yields three times those of wells on hills and ridges. According to Zewe and Rauch (1993), valley wells produce a median yield more than four times greater than hilltop wells in the carbonate area of West

Virginia, USA. In a study of Helvey and Rauch (1993) of sedimentary rock wells (n=469) in Pennsylvania, USA, median yields were much higher in valley bottom settings (1,140 Lhr^{-1}) versus either hillside locations (570 Lhr^{-1}) or hilltop locations (340 Lhr^{-1}). Steep slopes tend to have low yields in the fractured-bedrock aquifer of New Hampshire, USA (Moore et al. 2002a). In crystalline rock area of Satpura Hills in India, the yield of valley wells is more than five times better than that of flat upland wells; the median depth of both well groups is 46 m (Uhl et al. 1979). In the Limpopo Province basement aquifers, South Africa, Holland and Witthüser (2009) found that, in addition to valley settings, flat surfaces also had higher than average transmissivities suggesting that there are controlling factors other than topography on the productivity of the boreholes.

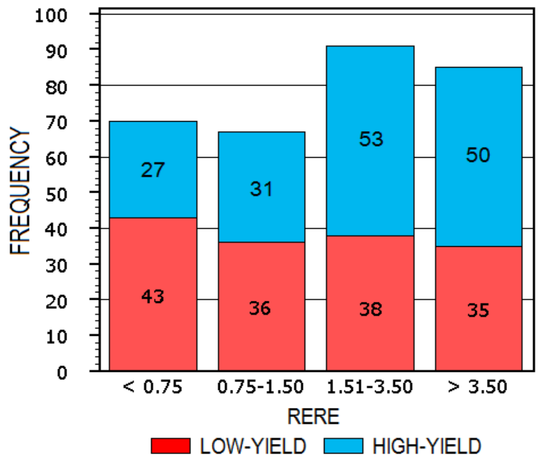


Fig. 81. Bar chart showing the distribution of low-yield (Q≤100 Lhr^{-1}) and high-yield (Q≥4000 Lhr^{-1}) drilled wells in the CF database in different RERE groups (n=313, test wells included). The number of wells per group is indicated within each bar.

Some authors regard topography (statistically) as a minor feature affecting the well yield (e.g. Lattman & Parizek 1964, Cederstrom 1972, Yin & Brook 1992a, 1992b, Knopman & Hollyday 1993). Owen et al. (2003) found no significant correlation between borehole yields and the macro-scale topography in crystalline basement aquifers of southern Zimbabwe. Johnson (1999) found that a lower average-fracture density than the wells sited on hill slopes and terraces characterized wells sited in valley settings in New Hampshire, USA.

Topography induces differences in hydraulic head (potential) leading to hydraulic gradients. Topography related hydraulic gradients induce groundwater circulation, reaching a depth of several kilometers, and might cause, for instance, the upwelling of deep-seated fluids (Stober et al. 1999). Topography-related hydraulic gradients for the Black Forest crystalline basement area in Germany are typically 0,05 m m^{-1} with a range 0,002-0,2 m m^{-1} (Stober & Bucher 2007).

In Central Finland, the height of the well site and that of the groundwater level correlate very highly significantly with each other (Table 13, Fig. 48). As the median SLOH value is three degrees (Table 52), the median hydraulic gradient in bedrock aquifers may then be estimated to be around 0,05 m m^{-1} (tan 3°). The median bulk hydraulic conductivity K is $9{,}5 \times 10^{-3}$ md^{-1} (Table 28). Hence, assuming that the typical range of effective porosity (n_e) of the crystalline bedrock up to typical drilled well depths lies between 0,0005-0,005 (Chapter 6.2.4), one can roughly quantify the groundwater bulk flow velocity (v) to range in bedrock aquifers of Central Finland between 0,1-1 md^{-1}. It must still be admitted that in individual fractures the flow velocities may well be several orders of magnitude higher.

7.3.4 Lineament factors

Around 30% of the drilled wells in the CF database (n=740) are situated so that at least one extracted lineament is running across their catchment area (1 km^2). Around 60% of those wells have depth and yield information in the CF database. In that what follows, these wells are called lineament wells whereas those wells without lineament extraction are called non-lineament wells.

The median distance from a well to the nearest lineament (LDIS) is 140 m; the mean is 168 m (Table 55). At its maximum there are six lineaments per a well´s catchment area (LFRE). Distance to the nearest lineament intersection (LIND) could be determined for about 200 drilled wells; the median LIND is 300 m. A typical well site is located about 10 meters above the bottom of the nearest lineament (ASLA).

Table 55. Statistics for lineament factors for all lineament wells in the CF drilled well database. Median values are in bold. The explanations of variable abbreviations are given in Table 3 and in Chapter 6.3.5.

Variable		n	Mean	Std dev	Median	Minimum	Maximum	Range
LDIS	m	740	168	112	**140**	1	540	539
LPRO	--	740	2	0,7	**2**	1	3	2
LFRE	--	740	1,6	0,9	**1**	1	6	5
LLEN	m	740	1381	575	**1120**	200	3700	3500
LDEN	--	740	1,38	0,58	**1,12**	0,2	3,7	3,5
LLLF	--	740	895	194	**920**	200	1220	1020
LINF	--	740	0,4	0,7	**0**	0	5	5
LIND	m	198	305	143	**300**	20	560	540
LASL	m a.s.l.	740	117	30	**111**	79	236	157
ASLA	m	740	10	10	**8**	-12	65	77
SLOL	°	740	3	3	**3**	-13	19	32
LAZI	°	740	107	56	**130**	5	180	175
LOSI	°	740	164	101	**150**	5	360	355
HLDIS	--	382	5,6	25,1	**1,6**	0,01	450	450
HALA	--	382	1,3	0,2	**1,3**	1,1	2,1	1,0
HLASL	m	382	35	16	**32**	8	97	89

Lineament wells (LDIS 1-564 m) are more productive than non-lineament wells. However, the differences in well depth, yield and hydraulic properties are statistically not significant (Table 56).

Table 56. Mean and median well depth, yield and hydraulic parameters for lineament and non-lineament wells in the CF drilled well database (n=1300, test wells included).

WELL GROUP	n		DEPTH m	Q Lhr^{-1}	Q/d_s $Lhr^{-1}m^{-1}$	Q/s m^2s^{-1}	Q_w ms^{-1}	T m^2s^{-1}	K ms^{-1}
LINEAMENT WELLS	449	Mean	86	1837	46	$5,15x10^{-5}$	$2,58x10^{-6}$	$2,64x10^{-5}$	$1,31x10^{-6}$
		Md	79	800	13	$1,55x10^{-5}$	$2,54x10^{-7}$	$8,18x10^{-6}$	$1,34x10^{-7}$
NON-LINEAMENT WELLS	851	Mean	89	1466	46	$5,17x10^{-5}$	$6,20x10^{-6}$	$2,63x10^{-5}$	$3,08x10^{-6}$
		Md	80	600	11	$1,31x10^{-5}$	$1,95x10^{-7}$	$6,94x10^{-6}$	$1,03x10^{-7}$

The distance to the nearest lineament (LDIS) has a very highly significant negative correlation with well yield and hydraulic parameters (Table 57). In other words, when the LDIS decreases the well yield and hydraulic values increase and vice versa. The ASLA correlates similarly with well yield and hydraulic parameters, but when partialled with LDIS the correlations fade away. This is because the ASLA-values near lineaments are typically smaller than farther away. In Central Finland, there exist both high-yield and low-yield wells along lineaments (LDIS ≤ 150 m; valley wells in Fig. 77). Valley wells contain 55% of lineament wells.

Table 57. Spearman rank correlation coefficients (r) with significance levels (p) and number of observations (n) between lineament factors and well depth, yield and hydraulic parameters for all lineament wells in the CF drilled well database. The explanations of variable abbreviations are given in Table 3 and in Chapter 6.3.5.

		DEPTH	Q	Q/d_S	Q/s	Q_W	T	K
LDIS	r	0,10403	-0,25339	-0,25931	-0,25931	-0,23833	-0,25931	-0,23776
n=449	p	0,0275	<0,0001	<0,0001	<0,0001	<0,0001	<0,0001	<0,0001
LPRO	r	-0,08442	0,08522	0,08966	0,08966	0,09223	0,08966	0,09195
n=449	p	0,0739	0,0712	0,0576	0,0576	0,0508	0,0576	0,0515
LFRE	r	-0,00362	-0,00489	0,01047	0,01047	0,01194	0,01047	0,01207
n=449	p	0,9391	0,9177	0,8249	0,8249	0,8008	0,8249	0,7986
LLEN	r	-0,00885	0,09337	0,09866	0,09866	0,08371	0,09866	0,08367
n=449	p	0,8517	0,0480	0,0366	0,0366	0,0764	0,0366	0,0766
LDEN	r	-0,00885	0,09337	0,09866	0,09866	0,08371	0,09866	0,08367
n=449	p	0,8517	0,0480	0,0366	0,0366	0,0764	0,0366	0,0766
LLLF	r	-0,02974	0,14693	0,13291	0,13291	0,11223	0,13291	0,11205
n=449	p	0,5296	0,0018	0,0048	0,0048	0,0174	0,0048	0,0175
LINF	r	-0,00276	0,05250	0,04936	0,04936	0,04203	0,04936	0,04188
n=449	p	0,9536	0,2670	0,2966	0,2966	0,3743	0,2966	0,3760
LIND	r	-0,02587	0,00071	-0,00798	-0,00798	0,00507	-0,00798	0,00485
n=125	p	0,7746	0,9938	0,9296	0,9296	0,9553	0,9296	0,9572
LASL	r	0,09113	-0,00087	-0,02588	-0,02588	-0,04349	-0,02588	-0,04405
n=449	p	0,0537	0,9854	0,5844	0,5844	0,3579	0,5844	0,3517
ASLA	r	0,12642	-0,19956	-0,22124	-0,22124	-0,21452	-0,22124	-0,21451
n=449	p	0,0073	<0,0001	<0,0001	<0,0001	<0,0001	<0,0001	<0,0001
SLOL	r	0,08256	-0,01474	-0,04988	-0,04988	-0,06509	-0,04988	-0,06567
n=449	p	0,0806	0,7555	0,2916	0,2916	0,1686	0,2916	0,1648
HLDIS	r	-0,21401	0,33683	0,35757	0,35757	0,34621	0,35757	0,34598
n=230	p	0,0011	<0,0001	<0,0001	<0,0001	<0,0001	<0,0001	<0,0001
HALA	r	0,01956	0,00349	-0,00553	-0,00553	-0,00733	-0,00553	-0,00699
n=230	p	0,7680	0,9581	0,9335	0,9335	0,9120	0,9335	0,9160
HLASL	r	0,14841	-0,00966	-0,05721	-0,05721	-0,08629	-0,05721	-0,08704
n=230	p	0,0244	0,8842	0,3878	0,3878	0,1922	0,3878	0,1884

The proximity of lineaments has proven to be of great importance to well yield and hydraulic properties in crystalline rock areas around the world. For instance, in the crystalline metamorphic environment of Maine, USA, wells located on lineaments are generally more productive than non-lineament wells (Mabee 1992). Also transmissivity, normalized by well depth, is positively related to well proximity to lineaments (Mabee et al. 1994). In the Precambrian bedrock of Burkina Faso, the results of 182 boreholes and wells clearly showed that proximity to a fault increases the well yields (Astier & Paterson 1989). In São Paulo district, Brazil, all except one high-capacity wells (Q/s >$1,4x10^{-4}$ m^2s^{-1}) are close to or on lineaments (Fernandes & Rudolph 2001). In addition, wells close to more than one lineament tend to have higher median Q/s values, though the range of Q/s values is the same as for the wells associated with a single lineament.

The hydraulic conductivity of a fault zone in fractured rock of the Lange Bramke basin, Germany, obtained from the tracer tests by Maloszewski et al. (1999) was about $1{,}5\times10^{-2}$ ms^{-1}, whereas the regional hydraulic conductivity of the fractured rock mass was about $3{,}0\times10^{-7}$ ms^{-1}. These values show that the hydraulic conductivity along the fault may be several orders of magnitude larger than that of the remaining fractured part of the aquifer, which, according to Maloszewski et al. (1999), confirms the dominant role of the fault zones as collectors of water and conductors of fast flow.

On the other hand, some wells in crystalline rock areas have low yields despite being located on or close to a lineament and conversely some wells at great distance away from a lineament have relatively high yields. Waters (1989) has made such observations in the Masvingo metamorphic rock area in Zimbabwe. In São Paulo district, Brazil, many wells along lineaments have low specific capacities (Fernandes & Rudolph 2001). In North East Brazil, Tröger et al. (2003) noted that wells located on clear lineament structures or even on intersected lineaments gave unexpectedly little amounts of water.

In Central Finland, the total length of lineaments (LLEN) and the lineament length density (LDEN) in a well´s catchment area correlate significantly with Q/d_s, Q/s and T (Table 57). The LLEN has some influence on well yield, too. With the exception of the 1001-1250 m group, the median well yield increases with the LLEN (Fig. 82). Hydraulic parameters do not differ statistically from each other in different LLEN groups. The number of lineaments (LFRE) does not have statistically significant relations with the well yield and hydraulic values in Central Finland. However, the combination variable LLLF=LLEN/LFRE has highly significant correlations with these parameters (Table 57). When only those wells drilled in granite and granodiorite areas (n=326) are considered, LFRE has weak positive correlations with hydraulic parameters. Interestingly, the median well yields seem to decrease when there are more than two lineaments in a well's catchment area (Fig. 83).

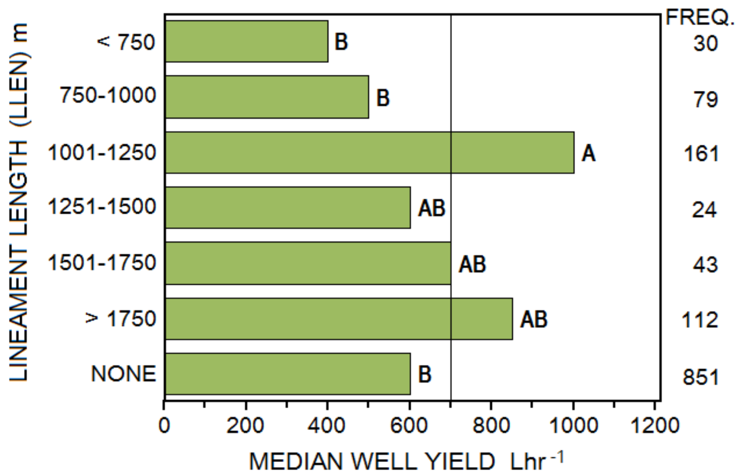


Fig. 82. Bar chart showing the median well yield in different LLEN groups for all drilled wells in the CF database (n=1300). The group 'none' contains only non-lineament wells. Comparisons significant at the $\alpha \leq 0.05$ level between the groups tested with the nonparametric median one-way analysis for group medians are indicated by different letters A and B. The overall median yield of wells (700 Lhr^{-1}, n=1308) is marked with a vertical line. The number of wells per bar is indicated in the rightmost column.

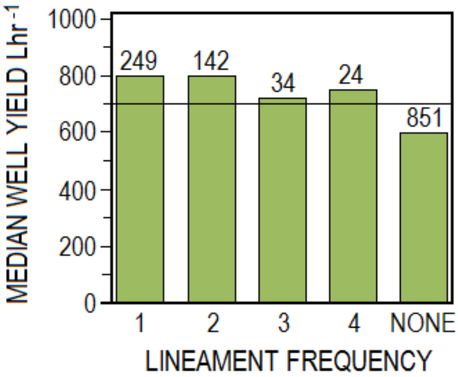


Fig. 83. Bar chart showing the median well yield in different lineament frequency (LFRE) groups for all drilled wells in the CF database (n=1300). The group 'none' represents non-lineament wells. The number of wells per bar is indicated above each bar. The overall median yield of wells in the CF database (700 Lhr^{-1}, n=1308) is marked with a horizontal line.

Tossavainen (1992) found a positive correlation between lineament density and well yield in crystalline rocks of southeastern Finland. Also in the study of Henriksen (2006a), the lineament density was considered an important factor influencing well yield in the bedrock of Sunnfjord, Western Norway. According to Edet et al. (1998), lineament-length density values show a moderate positive correlation with well yield and transmissivity in southeastern Nigeria. On the other hand, Lewis (1990) found no correlation between the density of lineaments and the specific capacity of 145 boreholes in basement aquifers of Malawi.

The number of lineament intersections (LINF) and the distance to the nearest lineament intersection (LIND) do not correlate with well yield and hydraulic parameters in Central Finland (Table 57). In other hard rock areas lineament intersections have shown to be of major importance to well yield. In the carbonate rocks of Pennsylvania, USA, Lattman and Parizek (1964) found that wells located at the intersection of two fractures exhibited yields 10 to 100 times greater than wells drilled in zones between fractures. Woodruff et al. (1974) noted that in the crystalline Piedmont area of Delaware, USA, wells drilled on fracture intersections showed more than 10 times greater yield than average wells in the area. In crystalline rock area of Vancouver Island in British Columbia, Canada, Kenny et al. (2006) found that well yields decrease with increasing distance from lineament intersections. In the Precambrian bedrock of Burkina Faso, drilling close to a fault intersection increased the probability of obtaining a yield greater than 1,000 Lhr^{-1} between 50 and 80 percent (Astier & Paterson 1989). In Ghana, intersections between some weakly expressed lineaments and drainages proved to be good well locations (Sander 1996). According to Dash (2003), the areas of intersecting lineaments mark hydropotential zones in Orissa region in India. In Cypros, the highest-yield wells were situated with a lineation intersection less than 700 m away up-hill from the wells (Afrodisis 1990). In crystalline rock regions of Georgia, USA, Brook (1988) found that well depths increased with increasing distance from a lineament intersection.

Well yields in Central Finland are at their highest in central parts of the lineaments (LDIS <50 m) where the median well yield is 1800 Lhr^{-1} (Fig. 84, Table 58). This is nearly three times that of beyond 100 meters (700 Lhr^{-1}) and six times that of beyond 300 meters (320 Lhr^{-1}) from a lineament centre. Also the Q/s- and T-values are at their highest in lineament centers (LDIS < 50 m), where the median transmissivity ($1{,}7 \times 10^{-5}$ m^2s^{-1}) is seven times that of the T-value beyond 300 meters. The Q_w- and K-values are at their highest in the LDIS-group 50-100 m, where the median K-value ($2{,}8\times10^{-7}$ ms^{-1}) is nearly one magnitude higher than that of beyond 300 meters. In distances beyond 100 m hydraulic values are lower than in central parts of lineaments; most differences are statistically significant. However, none of the parameters between LDIS-groups <50m and 50-100 m differ statistically from each other (Table 58). The fracturing degree of bedrock (FRAC) is at its highest in the LDIS groups <50 and 50-100 m.

Interestingly, the median yield and hydraulic values of non-lineament wells (group NONE, n=851) are similar to those of the LDIS group 151-300 m and statistically higher than in the wells of the LDIS group > 300 m (Fig. 84 and 85, Table 58). However, the lineament centre wells (LDIS ≤ 100 m) have significantly higher yields and hydraulic values than the group 'none' wells.

In Central Finland, the mean and median LDIS-distances are shortest in well sites consisting of coarse- and fine-grained soils. This partly explains the higher yield and hydraulic values of fine-grained well sites compared to outcrop and till well sites (Chapter 7.3.2).

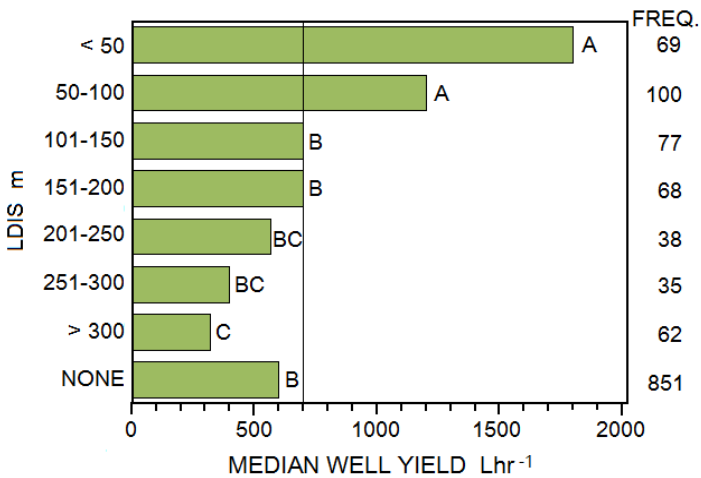


Fig. 84. Bar chart showing the median well yield in different LDIS groups for all drilled wells in the CF database (n=1300). The group 'none' represents non-lineament wells. Comparisons significant at the α≤0.05 level between the groups tested with the nonparametric median one-way analysis for group medians are indicated by different letters A, B, etc. The overall median yield of wells (700 Lhr^{-1}, n=1308) is marked with a vertical line. The number of wells per bar is indicated in the rightmost column.

Table 58. Statistics for well depth, yield and hydraulic parameters in different LDIS groups of all drilled wells in the CF database (n=1300). The group ‘none’ represents non-lineament wells. Highest values are in bold. Comparisons significant at the α≤0.05 level between the groups tested with the nonparametric Kruskal-Wallis test for group means and with the median one-way analysis for group medians are indicated by different letters $^{A, B, etc.}$.

Variable	Para-meter	Distance to the nearest lineament (LDIS)					
		< 50 m n=69 5%	50-100 m n=100 8%	101-150 m n=77 6%	151-300 m n=141 11%	> 300 m n=62 5%	NONE n=851 65%
DEPTH m	mean median	83 ^{ABC} 91 ^{ABCDE}	74 ^{C} 71 ^{CDE}	83 ^{BC} 70 ^{DE}	92 ^{AB} 82 ^{BC}	**98 ^{A}** **97 ^{AB}**	89 ^{AB} 80 ^{CD}
Q Lhr^{-1}	mean median	**2752 ^{A}** **1800 ^{A}**	2301 ^{A} 1200 ^{A}	1576 ^{B} 700 ^{B}	1592 ^{B} 600 ^{BC}	953 ^{C} 320 ^{C}	1466 ^{B} 600 ^{B}
Q/d_s $Lhr^{-1}m^{-1}$	mean median	66 ^{A} **29 ^{A}**	**72 ^{A}** 22 ^{A}	31 ^{B} 13 ^{B}	33 ^{B} 9 ^{B}	28 ^{C} 4 ^{C}	46 ^{B} 11 ^{B}
Q/s m^2s^{-1}	mean median	$7,45\times10^{-5}$ ^{A} **$3,28\times10^{-5}$ ^{A}**	**$8,04\times10^{-5}$ ^{A}** $2,54\times10^{-5}$ ^{A}	$3,54\times10^{-5}$ ^{B} $1,53\times10^{-5}$ ^{B}	$3,74\times10^{-5}$ ^{B} $1,10\times10^{-5}$ ^{B}	$3,11\times10^{-5}$ ^{C} $4,40\times10^{-6}$ ^{C}	$5,17\times10^{-5}$ ^{B} $1,31\times10^{-5}$ ^{B}
Q_w ms^{-1}	mean median	$3,53\times10^{-6}$ ^{AB} $4,81\times10^{-7}$ ^{AB}	$5,59\times10^{-6}$ ^{A} **$5,28\times10^{-7}$ ^{AB}**	$9,37\times10^{-7}$ ^{BC} $2,73\times10^{-7}$ ^{BC}	$1,32\times10^{-6}$ ^{C} $1,67\times10^{-7}$ ^{C}	$1,62\times10^{-6}$ ^{D} $5,90\times10^{-8}$ ^{D}	**$6,20\times10^{-6}$ ^{C}** $1,95\times10^{-7}$ ^{C}
T m^2s^{-1}	mean median	$3,81\times10^{-5}$ ^{A} **$1,72\times10^{-5}$ ^{A}**	**$4,11\times10^{-5}$ ^{A}** $1,33\times10^{-5}$ ^{A}	$1,83\times10^{-5}$ ^{B} $8,08\times10^{-6}$ ^{B}	$1,93\times10^{-5}$ ^{B} $5,85\times10^{-6}$ ^{B}	$1,59\times10^{-5}$ ^{C} $2,37\times10^{-6}$ ^{C}	$2,63\times10^{-5}$ ^{B} $6,94\times10^{-6}$ ^{B}
K ms^{-1}	mean median	$1,79\times10^{-6}$ ^{AB} $2,53\times10^{-7}$ ^{AB}	$2,82\times10^{-6}$ ^{A} **$2,77\times10^{-7}$ ^{AB}**	$4,84\times10^{-7}$ ^{BC} $1,43\times10^{-7}$ ^{BC}	$6,73\times10^{-7}$ ^{C} $8,90\times10^{-8}$ ^{C}	$8,18\times10^{-7}$ ^{D} $3,20\times10^{-8}$ ^{D}	**$3,08\times10^{-6}$ ^{C}** $1,03\times10^{-7}$ ^{C}

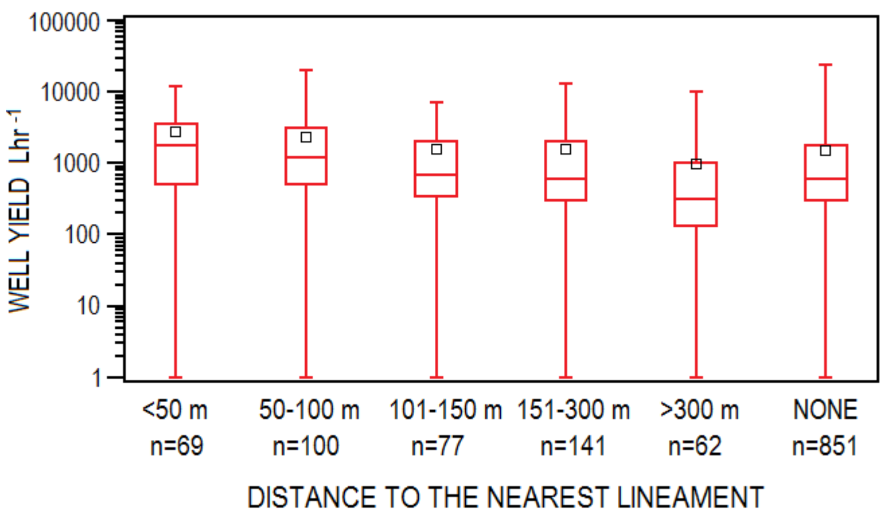


Fig. 85. Box-plot of well yield vs. distance to the nearest lineament for all drilled wells in the CF database (n=1300). The group ‘none’ contains only non-lineament wells. The maximum and minimum values and the 25th and 75th percentiles of the yield are presented with tops and bottoms on the whiskers and on the boxes, respectively. Median is presented with a horizontal line in the boxes and the mean with a square. The yield of a dry well has been entered into calculations as a nominal figure of 1 Lhr^{-1}.

The relative number of high-yield wells is at its highest in the LDIS group 50-100 m (74%) and the second highest in the group <50 m (65%). In the LDIS group >300 m most wells (80%) are low-yield wells. In the LDIS groups between 101 and 300 m the numbers are even (Fig. 86).

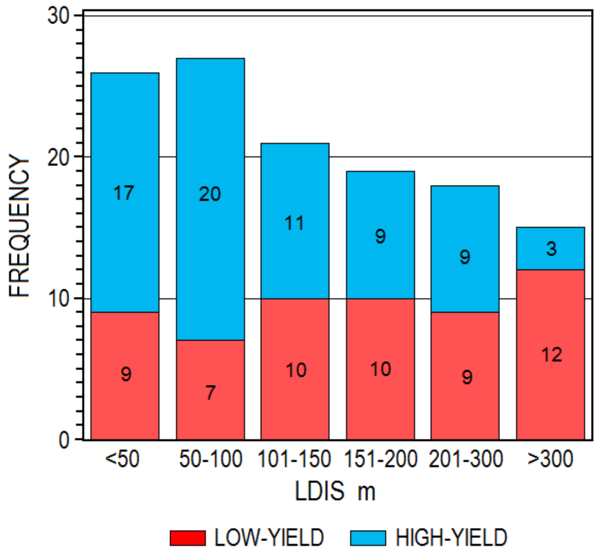


Fig. 86. Bar chart showing the distribution of low-yield (Q≤100 Lhr^{-1}) and high-yield (Q≥4,000 Lhr^{-1}) drilled wells in the CF database in different LDIS groups (n=126, test wells included). The number of wells per group is indicated within each bar.

In the Gothenburg crystalline rock region, southwestern Sweden, the Q/d values of drilled wells decrease as the distance to nearest lineament increases from 0 to 150 m (Wladis & Rosenbaum 1994, 1995, Wladis 1995). In the Upper Austrian Bohemian Massif, median borehole yield correlates negatively with distance to nearest lineament: 0-200 m/3,460 Lhr^{-1}, 200-400 m/1,580 Lhr^{-1}, >400 m/360-720 Lhr^{-1} (Stibitz 1998). Also Owen et al. (2003) found a clear negative correlation between distance to lineament and well yield in Archean crystalline basement aquifers of southern Zimbabwe.

In Robinson´s (2002) study of carbonate rocks in West Virginia, USA, the wells within 400 meters of a strike-slip fault had 47 times greater median well yield than wells >400 meters away. Wells <100 meters to a thrust fault had a median yield seven times greater than wells >100 meters away. In the Precambrian bedrock of Burkina Faso, proximity to a fault was noticeable in well yields to a distance of at least 600 m (Astier & Paterson 1989). In a rural groundwater project in Ghana, West Africa, Chesley et al. (1995) considered 300 m as the maximum lateral distance a lineament would affect hydrologic properties. In São Paulo district, Brazil, Madrucci et al. (2003) determined a high well yield area beyond 200 m from lineaments and beyond 600 m from lineament intersections. It was also evidenced that high well yields are related to low and intermediary morphostructure features.

In some crystalline rock areas, lineament centers have turned out to be rather poor sites for drilled wells. In crystalline rocks of Norway, Henriksen (2006a) found that the well yields increased about 50 m from the lineament centre, while at the center they were clearly lower. Solomon and Ghebreab (2008) confirmed this in the central highlands of Eritrea: the yield of wells was improved at a distance of 90-150 m away from the shear fractures. In crystalline rock area of Vancouver Island in British Columbia, Canada, Kenny et al. (2006) noted that the median yield was higher for wells between 50 and 300 m from a lineament in comparison to wells <50 m from a lineament. In Johansson´s (2005) study in Kandy area, Sri Lanka, the median yield of wells situated between 0 and 50 m from a lineament was 0 Lhr^{-1} (!), whereas the median well yield between 50 and 100 m from a lineament was about 750 Lhr^{-1}. Morland (1997), Braathen et al. (1999) and Kellgren (2002) did not found any clear relationship between distance to regional lineaments and borehole yields. Cho et al. (2003) found no evidence supporting a relationship between well yield and distance to lineament or lineament density in hard rocks of the Korean Peninsula.

In Central Finland, short lineaments (LPRO <2 km) seem to be more productive than long lineaments (LPRO >10 km, Table 59). This is highlighted in test valley wells, where the median yield difference between short and long lineaments is more than 3,000 Lhr^{-1}; the corresponding statistically significant difference in the K-values is more than one-order of magnitude (Table 59). However, a closer examination reveals that the distance to the nearest lineament (LDIS) is affecting more importantly on well yield and hydraulic values than LPRO (Table 60).

In short lineaments there are both low-yield and high-yield wells nearly in proportion to their overall prevalence in this LPRO group (Fig. 87). In medium length lineaments the number of high-yield wells is higher and that of low-yield wells lower than expected whereas in long lineaments the situation is opposite.

Table 59. Statistics for well depth, yield and hydraulic parameters in different LPRO groups of drilled valley wells (LDIS≤150 m) in the CF database. Highest values are in bold. Comparisons significant at the α≤0.05 level between the groups tested with the nonparametric Kruskal-Wallis test for group means and with the median one-way analysis for group medians are indicated by different letters $^{A\ and\ B}$.

Variable	Parameter	Length of the nearest fracture zone (LPRO)					
		All valley wells n=246			Test valley wells n=51		
		< 2 km n=78 32%	2-10 km n=119 48%	> 10 km n=49 20%	< 2 km n=15 29%	2-10 km n=26 51%	> 10 km n=10 20%
DEPTH m	mean median	81 79	76 67	**86** **80**	80 ^{B} 79	85 ^{AB} 91	**102** ^{A} **111**
Q Lhr^{-1}	mean median	**2460** **1400**	2200 1100	1788 850	**4319** **4000**	3926 2830	1957 885
Q/d_s $Lhr^{-1}m^{-1}$	mean median	**66** 18	55 **20**	50 13	**79** **71**	59 49	30 11
Q/s m^2s^{-1}	mean median	**$7,42x10^{-5}$** $2,12x10^{-5}$	$6,19x10^{-5}$ **$2,36x10^{-5}$**	$5,63x10^{-5}$ $1,55x10^{-5}$	**$8,88x10^{-5}$** **$8,00x10^{-5}$**	$6,65x10^{-5}$ $5,56x10^{-5}$	$3,40x10^{-5}$ $1,22x10^{-5}$
Q_w ms^{-1}	mean median	**$4,97x10^{-6}$** $3,36x10^{-7}$	$3,00x10^{-6}$ **$4,56x10^{-7}$**	$2,66x10^{-6}$ $2,07x10^{-7}$	**$1,88x10^{-6}$** **$1,31x10^{-6}$** ^{A}	$1,31x10^{-6}$ $9,64x10^{-7}$ ^{AB}	$7,39x10^{-7}$ $1,02x10^{-7}$ ^{B}
T m^2s^{-1}	mean median	**$3,79x10^{-5}$** $1,12x10^{-5}$	$3,18x10^{-5}$ **$1,24x10^{-5}$**	$2,88x10^{-5}$ $8,18x10^{-6}$	**$4,55x10^{-5}$** **$4,13x10^{-5}$**	$3,42x10^{-5}$ $2,88x10^{-5}$	$1,76x10^{-5}$ $6,47x10^{-6}$
K ms^{-1}	mean median	**$2,50x10^{-6}$** $1,77x10^{-7}$	$1,53x10^{-6}$ **$2,39x10^{-7}$**	$1,35x10^{-6}$ $1,10x10^{-7}$	**$9,65x10^{-7}$** **$6,79x10^{-7}$** ^{A}	$6,77x10^{-7}$ $4,99x10^{-7}$ ^{AB}	$3,81x10^{-7}$ $5,40x10^{-8}$ ^{B}

Table 60. Median well depth, yield and hydraulic parameters for different LPRO vs. LDIS groups of all drilled wells in the CF database (n=449). Highest values are in bold. Comparisons significant at the α≤0.05 level between the groups tested with the nonparametric median one-way analysis are indicated by different letters $^{A\ and\ B}$ or by different numbers $^{1\ and\ 2}$.

LPRO km	Variable	Dimension	Distance to the nearest lineament (LDIS) ≤150 m n *%*		151-300 m n *%*		> 300 m n *%*	
			n *%*	median	n *%*	median	n *%*	median
< 2	Depth	m	78 *17*	79^{B}	29 *6*	68^{B}	16 *4*	$99^{\mathbf{A}}$
	Q	Lhr^{-1}		$\mathbf{1400}^{\mathbf{A}}$		600^{AB}		250^{B}
	Q/d_s	$Lhr^{-1}m^{-1}$		18^{A}		9^{AB}		3^{B}
	Q/s	m^2s^{-1}		$2{,}12\text{x}10^{-5\ A}$		$1{,}10\text{x}10^{-5\ AB}$		$3{,}21\text{x}10^{-6\ B}$
	Q_w	ms^{-1}		$3{,}36\text{x}10^{-7\ A}$		$2{,}90\text{x}10^{-7\ A}$		$4{,}30\text{x}10^{-8\ B}$
	T	m^2s^{-1}		$1{,}12\text{x}10^{-5\ A}$		$5{,}85\text{x}10^{-6\ AB}$		$1{,}73\text{x}10^{-6\ B}$
	K	ms^{-1}		$1{,}77\text{x}10^{-7\ A}$		$1{,}53\text{x}10^{-7\ A}$		$2{,}30\text{x}10^{-8\ B}$
2-10	Depth	m	119 *27*	67^{2}	79 *18*	82^{1}	28 *6*	82^{12}
	Q	Lhr^{-1}		1100^{1}		600^{12}		420^{2}
	Q/d_s	$Lhr^{-1}m^{-1}$		$\mathbf{20}^{\mathbf{1}}$		11^{12}		6^{2}
	Q/s	m^2s^{-1}		$\mathbf{2{,}36\text{x}10^{-5\ 1}}$		$1{,}29\text{x}10^{-5\ 12}$		$7{,}18\text{x}10^{-6\ 2}$
	Q_w	ms^{-1}		$\mathbf{4{,}56\text{x}10^{-7\ 1}}$		$1{,}62\text{x}10^{-7\ 2}$		$1{,}25\text{x}10^{-7\ 2}$
	T	m^2s^{-1}		$\mathbf{1{,}24\text{x}10^{-5\ 1}}$		$6{,}85\text{x}10^{-6\ 12}$		$3{,}83\text{x}10^{-6\ 2}$
	K	ms^{-1}		$\mathbf{2{,}39\text{x}10^{-7\ 1}}$		$8{,}60\text{x}10^{-8\ 2}$		$6{,}60\text{x}10^{-8\ 2}$
> 10	Depth	m	49 *11*	80	33 *7*	**101**	18 *4*	98
	Q	Lhr^{-1}		850		480		380
	Q/d_s	$Lhr^{-1}m^{-1}$		13		8		3
	Q/s	m^2s^{-1}		$1{,}55\text{x}10^{-5}$		$9{,}01\text{x}10^{-6}$		$4{,}13\text{x}10^{-6}$
	Q_w	ms^{-1}		$2{,}07\text{x}10^{-7}$		$7{,}80\text{x}10^{-8}$		$5{,}30\text{x}10^{-8}$
	T	m^2s^{-1}		$8{,}18\text{x}10^{-6}$		$4{,}79\text{x}10^{-6}$		$2{,}22\text{x}10^{-6}$
	K	ms^{-1}		$1{,}10\text{x}10^{-7}$		$4{,}20\text{x}10^{-8}$		$2{,}90\text{x}10^{-8}$

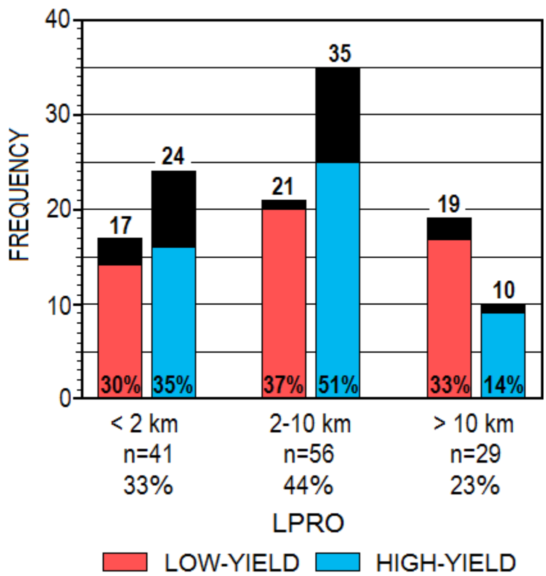


Fig. 87. Bar chart showing the distribution of low-yield (Q≤100 Lhr^{-1}) and high-yield (Q≥4,000 Lhr^{-1}) drilled wells in the CF database in different LPRO groups (n=126, test wells marked in black). The number of wells per bar is indicated above each bar. The percentages within each bar have been calculated separately for low-yield (n=57) and high-yield wells (n=69). The expected percentages for wells in different LPRO groups are shown below the bars.

In Central Finland, the median well yield in the innermost lineament center (LDIS <50 m) is nearly the same (1,600-1,800 Lhr^{-1}) in all LPRO groups (Fig. 88). The well yields also tend to decrease with increasing distance from the lineaments. However, the rate of decrease is different in each LPRO group. In short lineaments the median well yields decline steadily, whereas in other lineament groups the decrease is sharper being perhaps most abrupt in long lineaments. If the lineament width is defined on the basis of median well yield of 1,000 Lhr^{-1} or more, the following lineament widths are to be obtained: LPRO <2km 300 m, LPRO 2-10 km 200 m, and LPRO >10 km 100 m (Fig. 88). That is, the shorter the lineament the wider its permeable central part.

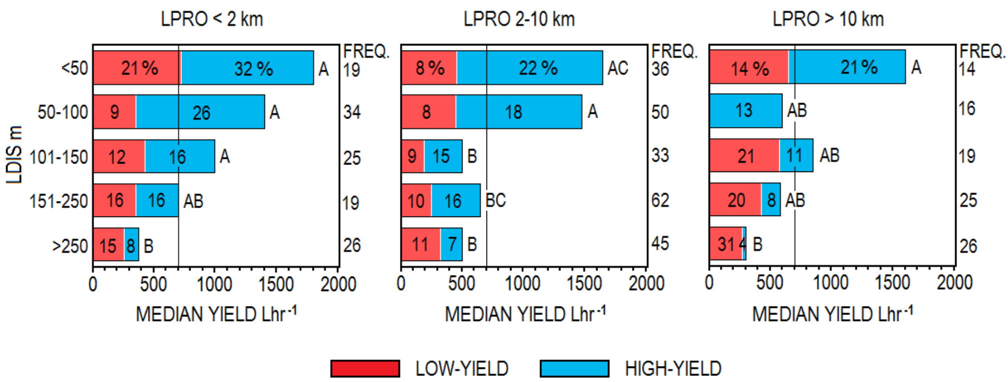


Fig. 88. Bar charts showing the median well yield and the relative distribution (in percentage) of low-yield ($Q \leq 100$ Lhr^{-1}) and high-yield ($Q \geq 4{,}000$ Lhr^{-1}) drilled wells in different LDIS vs. LPRO groups in the CF database (n=449, test wells included). Comparisons significant at the $\alpha \leq 0.05$ level between the LDIS groups in each LPRO group tested with the nonparametric median one-way analysis for group medians are indicated by different letters A, B, etc. The total number of wells per bar is indicated in the FREQ-columns. The overall median yield of wells in the CF database (700 Lhr^{-1}, n=1308) is marked with a vertical line.

In the United States, a strong correlation between increased water-well productivity and decreased distances to short lineaments (<2 km) has been observed in carbonate and shale aquifers (LaRiccia & Rauch 1977, Beebe & Rauch 1979, De La Garza & Slade 1986, Zewe & Rauch 1991, Helvey & Rauch 1993, Robinson 2002). However, no correlation is evident between well productivity and distances to long lineaments (Helvey & Rauch 1993). According to Buckley and Zeil (1984), there is some evidence from Botswana that the main groundwater circulation is not conducted through the most obvious shear lineaments but through a system of less obvious joints and fractures of tension origin related to the shear lineaments.

Opposite results have been presented from crystalline rocks of Nigeria, where Odeyemi et al. (1985) discovered a positive correlation between lineament length and groundwater yield and Malomo (1990) noted that longer (> 5 km) lineaments have better yields than shorter (<2,5 km) lineaments. Based on observations of the Beech Grove lineament of Tennessee, USA, Moore and Hollyday (1977) suggested that the stronger the surface expression and the longer the feature, the more significant the lineament is to groundwater exploration.

In Central Finland, the relative proportion of high-yield wells vs. low-yield wells is at its greatest (82%) when the relation HLDIS=HDIS/LDIS goes over 4,50; the corresponding proportion is at its lowest (28%) with HLDIS-values < 1,00 (Fig. 89).

Median well yields (650 Lhr^{-1}) and hydraulic values are somewhat lower when there is a water course in the nearest fracture zone (LWAT=1, n=94) compared to the situation when there is none (800 Lhr^{-1}; LWAT=0, n=355). However, the differences in well yields and hydraulic values are statistically not significant.

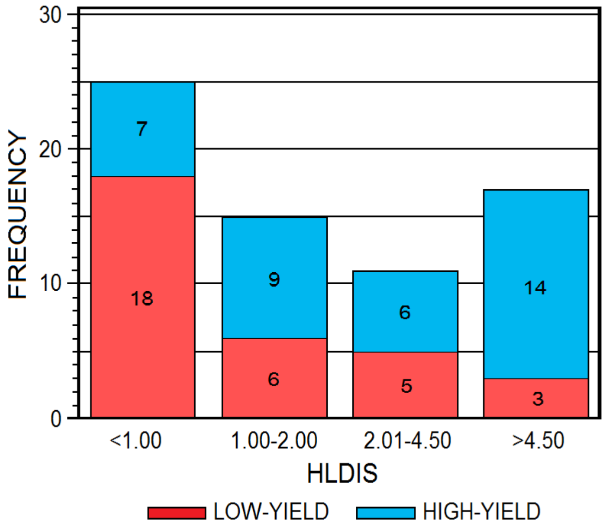


Fig. 89. Bar chart showing the distribution of low-yield (Q≤100 Lhr^{-1}) and high-yield (Q≥4,000 Lhr^{-1}) drilled wells in the CF database in different HLDIS groups (n=68, test wells included). The number of wells per group is indicated within each bar.

Median well yield is at its highest in LAZI groups 21-40°, 41-60° and 121-140°, where it varies from 1,125 to 1,200 Lhr^{-1}. The median yields in other LAZI groups are clearly lower varying from 300 to 760 Lhr^{-1} (Table 61, Fig. 90). When examining valley wells only (Table 62), the corresponding well yields in LAZI groups 21-40° and 121-140° are approximately twice that of all wells (Table 61). There are no statistically significant differences in the median LDIS values in different LAZI groups.

Table 61. Median depth, yield and hydraulic parameters for all drilled wells in the CF database in different LAZI groups (n=449). Highest values are in bold. Comparisons significant at the α≤0.05 level between the groups tested with the nonparametric median one-way analysis are indicated by different letters $^{A, B, etc.}$ or by different numbers $^{1 \text{ and } 2}$.

LAZI °	n	DEPTH m	Q Lhr^{-1}	Q/d$_s$ Lhr^{-1}m^{-1}	Q/s m^{2}s^{-1}	Q$_w$ ms^{-1}	T m^{2}s^{-1}	K ms^{-1}
1-20	50	77 ^{ABC}	700 ^{12}	12 ^{AB}	1,39x10^{-5} ^{12}	2,53x10^{-7} ^{AB}	7,34x10^{-6} ^{12}	1,34x10^{-7} ^{AB}
21-40	50	73 ^{BC}	**1200** ^{1}	**23** ^{AB}	**2,62x10^{-5}** ^{12}	3,04x10^{-7} ^{A}	**1,37x10^{-5}** ^{12}	1,60x10^{-7} ^{A}
41-60	32	85 ^{ABC}	1185 ^{12}	16 ^{AB}	1,84x10^{-5} ^{12}	2,68x10^{-7} ^{AB}	9,71x10^{-6} ^{12}	1,41x10^{-7} ^{AB}
61-80	16	**97** ^{AB}	300 ^{12}	2 ^{AB}	2,20x10^{-6} ^{12}	1,50x10^{-8} ^{AB}	1,19x10^{-6} ^{12}	8,00x10^{-9} ^{AB}
81-100	15	45 ^{D}	600 ^{12}	16 ^{AB}	1,89x10^{-5} ^{12}	**6,89x10^{-7}** ^{A}	9,95x10^{-6} ^{12}	**3,60x10^{-7}** ^{A}
101-120	42	81 ^{BC}	750 ^{12}	12 ^{AB}	1,43x10^{-5} ^{12}	2,61x10^{-7} ^{AB}	7,56x10^{-6} ^{12}	1,39x10^{-7} ^{AB}
121-140	90	73 ^{C}	1125 ^{1}	18 ^{A}	2,07x10^{-5} ^{1}	3,73x10^{-7} ^{AB}	1,09x10^{-5} ^{1}	1,96x10^{-7} ^{AB}
141-160	84	80 ^{BC}	760 ^{12}	14 ^{AB}	1,58x10^{-5} ^{12}	2,23x10^{-7} ^{AB}	8,34x10^{-6} ^{12}	1,17x10^{-7} ^{AB}
161-180	70	**97** ^{A}	500 ^{2}	9 ^{B}	1,05x10^{-5} ^{2}	1,07x10^{-7} ^{B}	5,57x10^{-6} ^{2}	5,70x10^{-8} ^{B}

Table 62. Median depth, yield and hydraulic parameters for all drilled valley wells in the CF database in different LAZI groups (n=246). Highest values are in bold. Comparisons significant at the α≤0.05 level between the groups tested with the nonparametric median one-way analysis are indicated by different letters [A, B, etc.] or by different numbers [1, 2, etc.].

LAZI °	n	DEPTH m	Q Lhr^{-1}	Q/d_s $Lhr^{-1}m^{-1}$	Q/s m^2s^{-1}	Q_w ms^{-1}	T m^2s^{-1}	K ms^{-1}
1-20	26	74 [AB]	1200 [123]	16 [AB]	$1,83x10^{-5}$ [12]	$4,58x10^{-7}$ [AB]	$9,66x10^{-6}$ [12]	$2,41x10^{-7}$ [AB]
21-40	30	75 [AB]	**2500** [12]	**34** [A]	**$3,87x10^{-5}$** [1]	$8,16x10^{-7}$ [AB]	**$2,02x10^{-5}$** [1]	$4,22x10^{-7}$ [AB]
41-60	19	84 [ABC]	1500 [123]	22 [AB]	$2,51x10^{-5}$ [12]	**$8,44x10^{-7}$** [AB]	$1,32x10^{-5}$ [12]	**$4,44x10^{-7}$** [AB]
61-80	13	**100** [A]	350 [3]	2 [AB]	$2,59x10^{-6}$ [12]	$2,30x10^{-8}$ [B]	$1,40x10^{-6}$ [12]	$1,20x10^{-8}$ [B]
81-100	11	43 [C]	700 [3]	21 [AB]	$2,38x10^{-5}$ [12]	$6,99x10^{-7}$ [AB]	$1,25x10^{-5}$ [12]	$3,67x10^{-7}$ [AB]
101-120	22	62 [ABC]	800 [23]	13 [AB]	$1,53x10^{-5}$ [12]	$3,22x10^{-7}$ [AB]	$8,07x10^{-6}$ [12]	$1,70x10^{-7}$ [AB]
121-140	50	72 [B]	2000 [1]	29 [A]	$3,28x10^{-5}$ [1]	$5,67x10^{-7}$ [A]	$1,71x10^{-5}$ [1]	$2,96x10^{-7}$ [A]
141-160	40	70 [AB]	825 [23]	17 [AB]	$2,02x10^{-5}$ [12]	$3,09x10^{-7}$ [AB]	$1,06x10^{-5}$ [12]	$1,62x10^{-7}$ [AB]
161-180	35	91 [ABC]	500 [3]	11 [B]	$1,31x10^{-5}$ [2]	$1,77x10^{-7}$ [AB]	$6,93x10^{-6}$ [2]	$9,20x10^{-8}$ [AB]

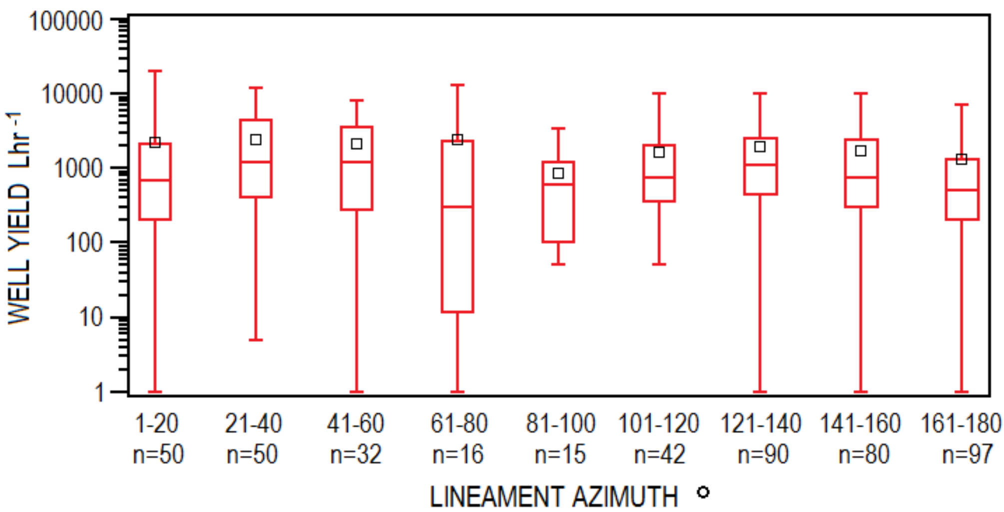


Fig. 90. Box-plot of well yield vs. lineament azimuth for all drilled wells in the CF database (n=449). The maximum and minimum values and the 25th and 75th percentiles of the yield are presented with tops and bottoms on the whiskers and on the boxes, respectively. Median is presented with a horizontal line in the boxes and the mean with a square. The yield of a dry well has been entered into calculations as a nominal figure of 1 Lhr^{-1}.

Approximately half of the nearest lineaments traversing wells´ catchment areas (n=449) in Central Finland run in the (N)NW-(S)SE direction. The lineament rose diagrams (Fig. 91a and 91b) are similar to that of all lineaments in Central Finland (Fig. 24). The wells are at their deepest nearly 100 meters in LAZI groups 61-80º and 161-180º; the shallowest group is LAZI 81-100º where the median well depth is only 45 m (Table 61, Fig. 92a and 92b).

When the median well yields are illustrated in a lineament azimuth rose diagram the most productive lineament directions (N40ºE and N50ºW with the longest horns) take up a position perpendicular to each other (Fig. 93a). In a corresponding way, the least productive lineaments (N80ºE and N10ºW with the shortest horns) cross at right angle with each other and they lie between the most productive lineaments. The situation is the same when examining the valley wells only (Fig. 93b) or the private drilled wells (Fig. 94a and 94b). Also hydraulic parameters behave in a similar manner (Figs 95-97). The pronounced appearance of the E-W direction in connection with hydraulic parameters is due to shallow well depths in LAZI group 81-100º (Tables 61-62). It should also be noted that the lineament frequency rose diagrams (Fig. 91a and 91b) are different to the productivity rose diagrams.

Rather unexpected results showed up when the LPRO factor was considered in examining the differences in well yields and hydraulic values between different LAZI groups. These examinations indicated that in long and medium length lineaments the most productive lineament direction is N40ºE, whereas in short lineaments the situation is totally different: the most productive direction is around N60ºW (Table 63, Figs 98 and 100-101). Long NE-SW faults (> 10 km) produce twice that much water than the corresponding NW-SE faults; with the short faults (< 2 km) the situation is opposite. There are no clear production differences between the corresponding fault directions in the medium length faults (Fig. 99).

Based on the LPOS directions, the median well yields seem to be at their highest on the NW side of the NE-SW orientated lineaments (Fig. 102a and 102b). Correspondingly, the SW-side of the NW-SE orientated lineaments have higher median yields. As was the case with LAZI, the most productive LPOS directions with the longest horns take up a position approximately perpendicular to each other. The number of observations in some LPOS directions is low and the differences between median yields and hydraulic values in different LPOS groups are statistically not significant.

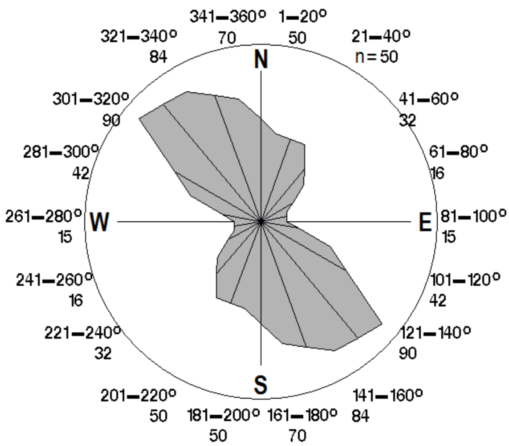


Fig. 91a. Number of lineaments in different LAZI groups (all drilled wells n=449). The radius of the circle is n=100.

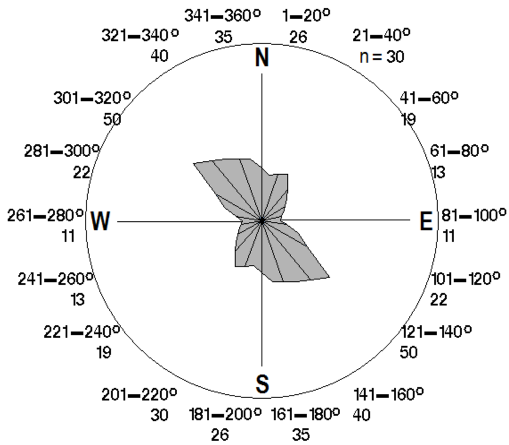


Fig. 91b. Number of lineaments in different LAZI groups (drilled valley wells n=246).The radius of the circle is n=100.

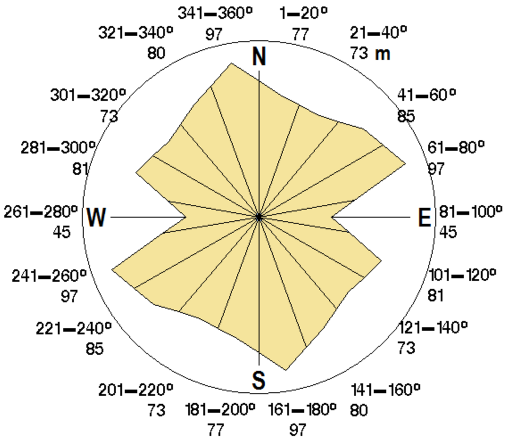


Fig. 92a. Median depth (m) of all drilled wells in different LAZI groups (n=449). The radius of the circle is 110 m.

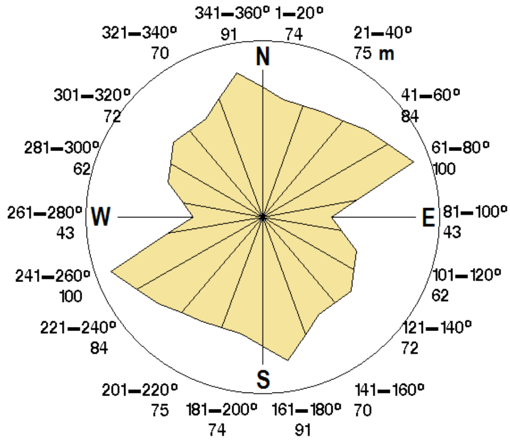


Fig. 92b. Median depth (m) of drilled valley wells in different LAZI groups (n=246). The radius of the circle is 110 m.

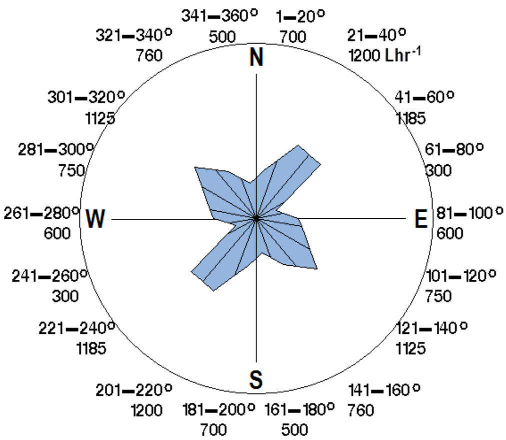


Fig. 93a. Median yield (Lhr^{-1}) of all drilled wells in different LAZI groups (n=449). The radius of the circle is 2,500 Lhr^{-1}.

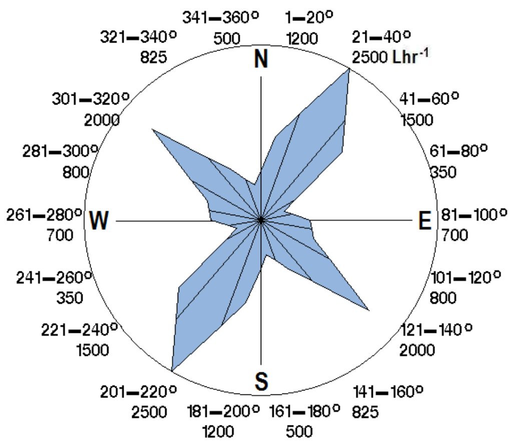


Fig. 93b. Median yield (Lhr^{-1}) of drilled valley wells in different LAZI groups (n=246). The radius of the circle is 2,500 Lhr^{-1}.

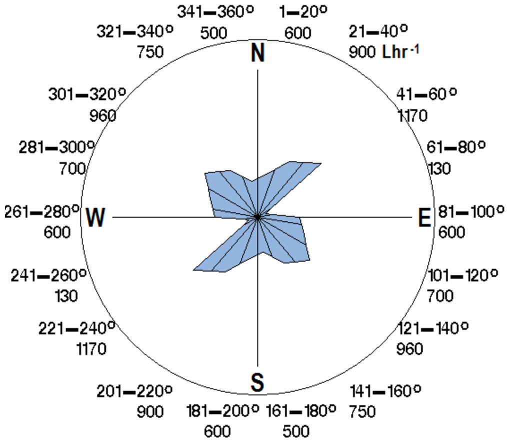


Fig. 94a. Median yield (Lhr^{-1}) of private drilled wells in different LAZI groups (n=396). The radius of the circle is 2,500 Lhr^{-1}.

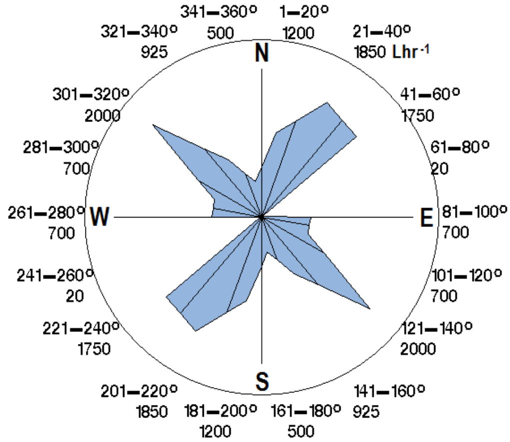


Fig. 94b. Median yield (Lhr^{-1}) of private valley wells in different LAZI groups (n=195). The radius of the circle is 2,500 Lhr^{-1}.

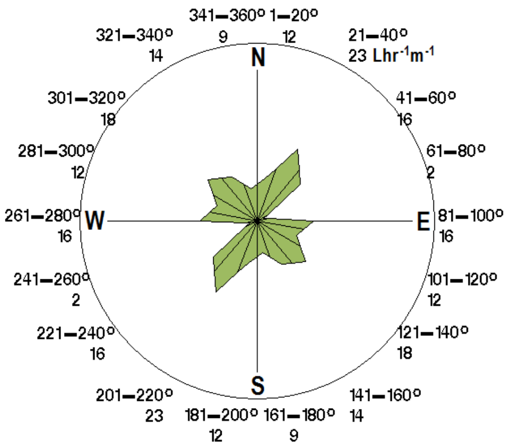


Fig. 95a. Median normalized yield (Lhr^{-1}m^{-1}) of all drilled wells in different LAZI groups (n=449). The radius of the circle is 50 Lhr^{-1}m^{-1}.

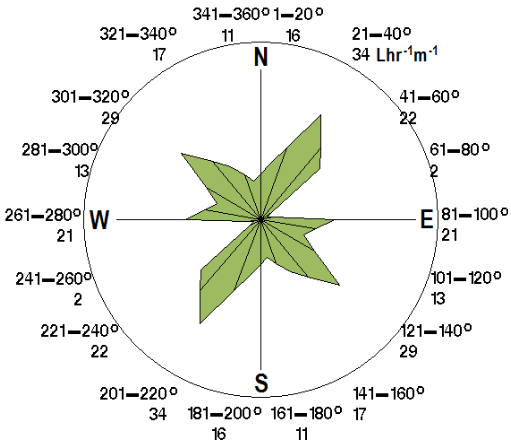


Fig. 95b. Median normalized yield (Lhr^{-1}m^{-1}) of drilled valley wells in different LAZI groups (n=246). The radius of the circle is 50 Lhr^{-1}m^{-1}.

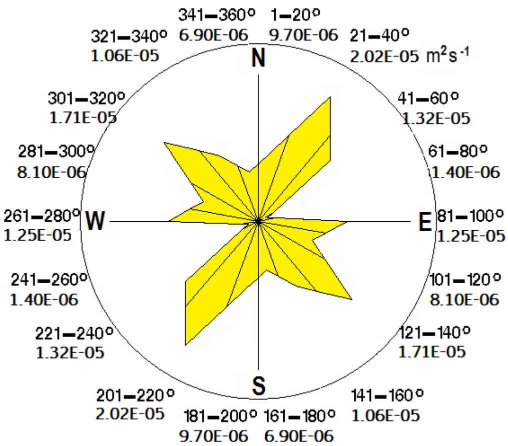


Fig. 96. Median transmissivity (m^2s^{-1}) of drilled valley wells in different LAZI groups (n=246). The radius of the circle is $2{,}5 \times 10^{-5}$ m^2s^{-1}.

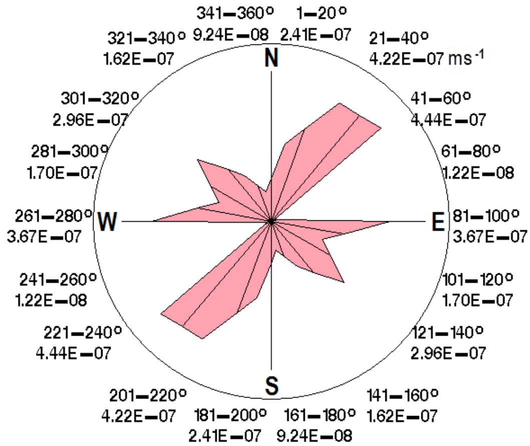


Fig. 97. Median hydraulic conductivity (ms^{-1}) of drilled valley wells in different LAZI groups (n=246).The radius of the circle is $5{,}5 \times 10^{-7}$ ms^{-1}.

Table 63. Median yield (Q) and normalized yield (Q/d_s) of all drilled wells in the CF database in different LAZI vs. LPRO groups (n=449). Highest values are in bold. Comparisons significant at the α≤0.05 level between the groups tested with the nonparametric median one-way analysis for group medians are indicated by different letters $^{A, B, etc.}$ or by different numbers $^{1, 2, etc.}$.

LAZI °	LPRO < 2 km n=123 28%			LPRO 2-10 km n=226 50%			LPRO > 10 km n=100 22%		
	n	Q Lhr^{-1} Median	Q/d_s $Lhr^{-1}m^{-1}$ Median	n	Q Lhr^{-1} Median	Q/d_s $Lhr^{-1}m^{-1}$ Median	n	Q Lhr^{-1} Median	Q/d_s $Lhr^{-1}m^{-1}$ Median
1-20	18	1100^{12}	13	22	650	9^{BD}	10	540	11
21-40	15	800^{12}	14	29	1500	29^{BD}	6	1700	21
41-60	18	900^{123}	10	10	**1750**	**32^{AD}**	4	**2100**	**24**
61-80	5	350^{123}	2	10	245	2^{ACD}	1	600	7
81-100	8	750^{23}	13	3	700	21^{ACD}	4	350	14
101-120	6	**2450^{12}**	41	25	500	10^{CB}	11	1000	17
121-140	17	2000^{1}	**49**	45	1400	26^{AB}	28	825	10
141-160	20	900^{12}	14	42	900	15^{BD}	22	510	6
161-180	16	300^{3}	3	40	650	12^{CD}	14	450	6

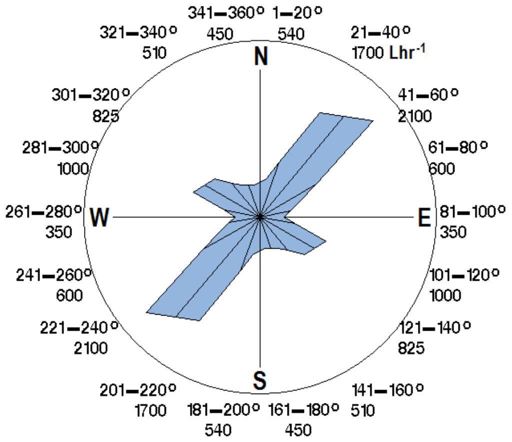


Fig. 98. Median yield (Lhr^{-1}) of long lineament (> 10 km) wells in different LAZI groups (n=100). The radius of the circle is 2,500 Lhr^{-1}.

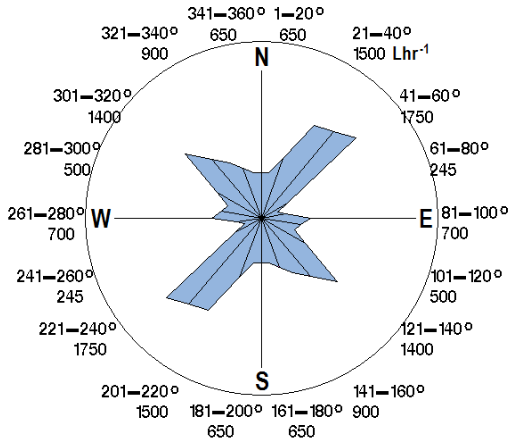


Fig. 99. Median yield (Lhr^{-1}) of medium length lineament (2-10 km) wells in different LAZI groups (n=226). The radius of the circle is 2,500 Lhr^{-1}.

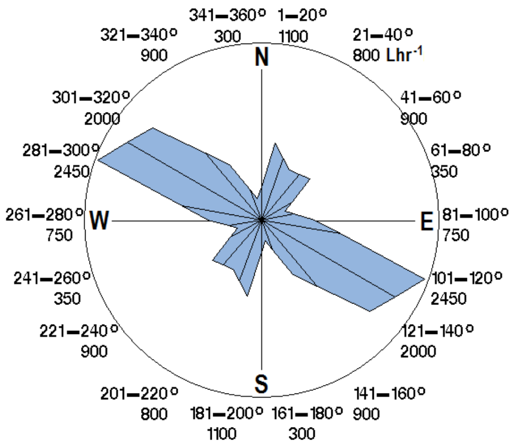


Fig. 100. Median yield (Lhr^{-1}) of short lineament (< 2 km) wells in different LAZI groups (n=123). The radius of the circle is 2,500 Lhr^{-1}.

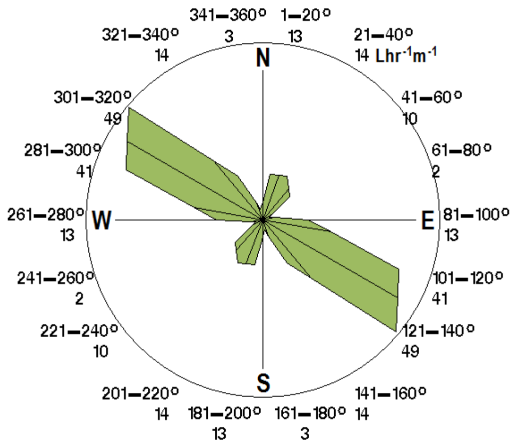


Fig. 101. Median normalized yield ($Lhr^{-1}m^{-1}$) of short lineament (< 2 km) wells in different LAZI groups (n=123). The radius of the circle is 50 $Lhr^{-1}m^{-1}$.

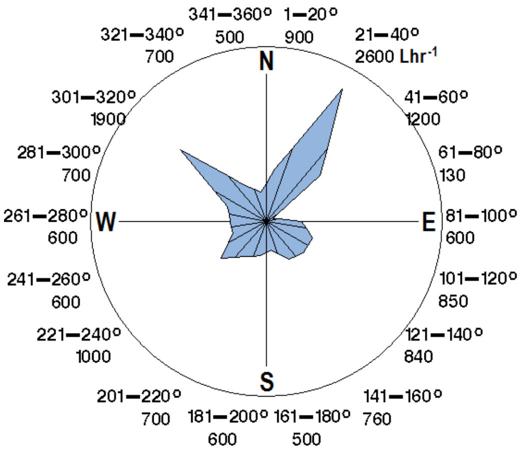


Fig. 102a. Median yield (Lhr^{-1}) of all drilled wells in different LPOS groups (n=449). The radius of the circle is 3,000 Lhr^{-1}.

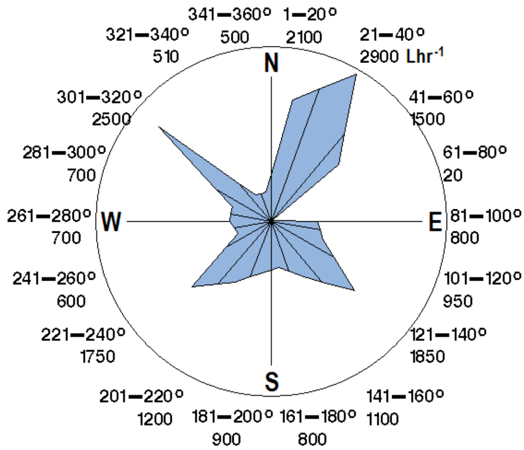


Fig. 102b. Median yield (Lhr^{-1}) of drilled valley wells in different LPOS groups (n=246). The radius of the circle is 3,000 Lhr^{-1}.

The relative frequency of high-yield wells vs. low-yield wells is at its highest in the LAZI group 21-40º and the second highest in the LAZI group 121-140º. High-yield wells are totally missing in the LAZI group 81-100º (Fig. 103).

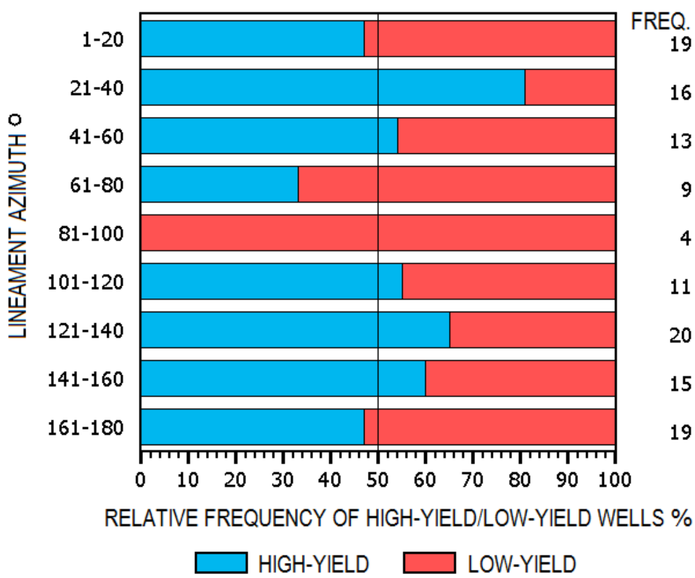


Fig. 103. Bar chart showing the relative frequency of high-yield (Q≥4,000 Lhr^{-1}) vs. low-yield (Q≤100 Lhr^{-1}) drilled wells in different LAZI groups in the CF database (n=126, test wells included). On the vertical line (50%) the number of high-yield and low-yield wells is equal. The total number of high-yield and low-yield wells per bar is indicated in the rightmost column.

Table 64 shows the distribution of low-yield and high-yield wells and the median yield of all wells in different LAZI groups for short, medium length and long lineaments. The first two LAZI groups represent the most productive lineament directions, i.e. 40±15° (NE-SW) and 130±15° (NW-SE) (Fig. 93a). The group OTHER represents the rest of the lineaments. Nearly one third of the wells in the NE-SW orientated medium length lineaments are high-yield wells. In the NW-SE orientated short lineaments high-yield wells comprise more than 1/3 of the wells. It should also be noted that there are no low-yield wells in this group. On the other hand, in long NW-SE lineaments the proportion of high-yield wells is only 5% whereas almost 20% of the wells are low-yield wells. In the long NE-SW lineaments the median well yield is at its highest; there are no low-yield wells in this group. However, the group only contains nine wells (2% of 449 lineament wells).

Table 64. Distribution of low-yield (Q≤100 Lhr^{-1}) and high-yield (Q≥4,000 Lhr^{-1}) drilled wells and the median yield of all drilled wells in different LAZI vs. LPRO groups in the CF database (total n=449, test wells included).

LAZI GROUP		LPRO < 2 km		LPRO 2-10 km		LPRO > 10 km		TOTAL	
		LOW	HIGH	LOW	HIGH	LOW	HIGH	LOW	HIGH
40°±15° NE-SW	n	3	6	3	11	0	2	6	19
	%	*11*	*21*	*8*	*31*	*0*	*22*	*8*	*26*
	Lhr^{-1}	900		1750		2400		1200	
	n	28		36		9		73	
130°±15° NW-SE	n	0	9	2	10	7	2	9	21
	%	*0*	*36*	*3*	*13*	*18*	*5*	*6*	*15*
	Lhr^{-1}	2000		1000		800		1000	
	n	25		76		39		140	
OTHER	n	14	9	16	14	12	6	42	29
	%	*20*	*13*	*14*	*12*	*23*	*12*	*18*	*12*
	Lhr^{-1}	550		650		510		565	
	n	70		114		52		236	
TOTAL	n	17	24	21	35	19	10	57	69
	%	*14*	*20*	*9*	*15*	*19*	*10*	*13*	*15*
	Lhr^{-1}	900		800		600		800	
	n	123		226		100		449	

Greenbaum (1992) examined the correlation between borehole yield and the trend of lineaments occurring within about 150 m of boreholes in crystalline basement aquifers of southeast Zimbabwe. His results showed that boreholes were drilled along lineaments of all directions with both successes and failures. The results did not convincingly demonstrate the existence of any favorable fracture direction, which was in agreement with Lewis (1990). LaRiccia and Rauch (1977) did not find any relationship between well productivity and nearest lineament orientation in carbonate aquifers of Frederick Valley, Maryland, USA. Neither was there any relationship between well yields and lineation trends in Cyprus (Afrodisis 1990).

In crystalline rocks of New Hampshire, USA, hydraulically active borehole fractures strike N30ºE and N45ºW. However, the orientation of these fractures is similar to the entire population of fracture orientations (Johnson 1999). In the same area, the wells within 30 m of lineaments trending N35ºW±5° have higher-than-average yields (Moore et al. 2002a). Mabee et al. (2002) found that 80% of the lineaments producing water to a bedrock tunnel in Massachusetts, USA, are orientated in the NW-SE direction. Sidle and Lee (1995) found that drilled wells situated among NW trending structures in granitoids of Maine, USA, have greater K values than other wells. In the carbonate terrain of West Virginia, USA, wells in the 90-135º-azimuth group of short lineaments have a significantly higher yield than wells in the other azimuth groups (Robinson 2002). Wells located near short straight lineaments that are oriented approximately perpendicular and especially parallel to local stratigraphic strike (25-40º) have much higher specific capacity and hydraulic conductivity values than wells in other settings in Greene County, Pennsylvania, USA (Helvey & Rauch 1993). Gas wells located near short photolineaments oriented N60ºW to N0ºE have significantly higher open flows than other wells in a gas field in West Virginia, USA (Beebe & Rauch 1979). In a granitic bedrock aquifer in New England, USA, Folan et al. (2004) found that the main water-bearing fracture zone (drilled well yields 10,440…22,700 Lhr^{-1}) is trending NNW-SSE with a dip of about 70º to the NE.

In foliated metamorphic rocks in the United States, major foliation parallel openings are the primary water-producing features responsible for high-yield wells (Lyford et al. 2003, Williams & Burton 2004, Williams et al. 2004, 2005, Manda et al. 2008).

In the Masvingo metamorphic rock area in Zimbabwe, SW-NE and NNW-SSE lineaments act as groundwater transmission and storage zones in preference to other trends (Waters 1989). In the Ivory Coast, Biémi et al. (1995) have noted that the NW-SE fracture zones are the most productive with 63% of the well yields being between 6,000 and 14,000 Lhr^{-1}, against 25% of those oriented N-S. These two directions account for the positioning of 62% of the wells in the region. Also in Burkina Faso higher well yields were detected in NW-SE orientated faults compared to NE-SW faults (Astier & Paterson 1989). In south-eastern Botswana northerly trending lineaments are highly productive (Zeil et al. 1991). In the crystalline rocks of the Upper West Region, Ghana, high-yielding boreholes are located along NNE to E trending fractures; yields decrease abruptly in boreholes more than 500 m from the lineaments. Boreholes along fractures oriented between ENE and E have low average yields close to lineaments but yields increase suddenly at a distance of 400 m of these zones (Banoeng-Yakubo & Skjernaa 2000).

Fernandes and Rudolph (2001) and Madrucci et al. (2003) discerned several weak trends in Q/s with respect to lineament direction in São Paulo district, Brazil. The highest specific capacities tend to be associated with lineaments ranging in N-S, NE-SW and NW-SE directions. However, very low Q/s values were also associated with the same orientations, which, according to Fernandes and Rudolph (2001), suggested that the lineaments might include both open and closed fracture features. In hard rocks of central Brazil the most important groundwater lineaments trend E-W and in part as conjugated pair NE and SE (Campos & Tröger 2000).

7.3.5 Catchment factors

A well´s average catchment area in Central Finland (1 km^2) consists of the following types of land use: forest land 51%, cultivated land 20%, settled area 12%, water courses 9%, outcrops of bedrock 5% and peat/paludified forest land 3% (Table 65, Fig. 104). The median distance from a well to the nearest water course (WDIS) is 210 meters; well sites lie on the average 10 meters above the level of the nearest water course (ASLW).

Catchment factors ASLW and ASWA have highly significant negative correlations with well yield and hydraulic parameters (Table 66). In other words, the smaller the difference or the lower the relation between the height of the well site (ASL) and that of the water level of the nearest water course (WASL), the higher the well yield and hydraulic values, and vice versa. The correlations still keep significant when partialled with LDIS. The factors WASL, CUL and partly SET have significant positive correlations and FOR significant negative correlation with well yield and hydraulic parameters. Distance to the nearest water course (WDIS) does not correlate statistically significantly with well yield and hydraulic parameters (Table 66).

Table 65. Statistics for catchment factors of all drilled wells in the CF database. Median values are in bold. The explanations of different abbreviations are given in Table 3 and in Chapter 6.3.6.

Variable		n	Mean	Std dev	Median	Minimum	Maximum	Range
WDIS	m	1587	230	151	**210**	5	560	555
WASL	m a.s.l.	1587	110	26	**106**	79	236	157
ASLW	m	1587	11	9	**9**	-28	65	93
ASWA	--	1587	1,11	0,09	**1,09**	0,78	1,74	0,96
BED	%	2395	5	8	**0**	0	68	68
FOR	%	2395	51	19	**52**	0	100	100
PEA	%	2395	3	6	**0**	0	92	92
CUL	%	2395	20	16	**16**	0	88	88
SET	%	2395	12	11	**8**	0	76	76
WAT	%	2395	9	13	**4**	0	76	76

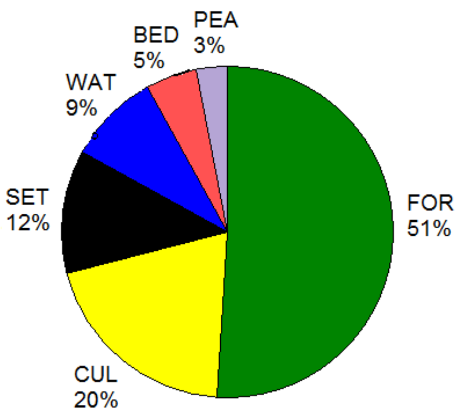


Fig. 104. Pie chart showing the average land use of a well´s catchment area (in percentage of 1 km^2) in the CF database (n=2395, test wells included).

In crystalline bedrock of New Hampshire, USA, Moore et al. (2002a) found that distance to a water body was negatively related to well yield: wells further away from water bodies tend to have lower than average yields. According to Tam et al. (2004), the closer the boreholes are to a water course, the higher is their specific capacity. On the other hand, Rosenberry and Winter (1993) and Mabee (1999) suggest that the surface water bodies have a minor influence on the yield of drilled wells. According to Johansson (2005), in Kandy district of Sri Lanka, the median well yield is 4,440 Lhr^{-1} at a distance of 0-250 m from a stream, whereas the median yield of wells further away from the stream is 240 Lhr^{-1}. In the present author´s opinion, this might be explained with streams occupying fracture zones in bedrock.

In order to statistically examine whether the ASLW affects the well depth, yield and hydraulic parameters, drilled wells in the CF database were divided into four groups according to their ASLW values as follows: ≤ 5, 6-10, 11-15 and >15 m. The results show that well yields and hydraulic values are in general significantly lower in the ASLW group >15 m compared to other groups (Table 67, Fig. 105).

Table 66. Spearman rank correlation coefficients (r) with significance levels (p) and number of observations (n) between catchment factors and well depth, yield and hydraulic parameters for all drilled wells in the CF database. The explanations of different abbreviations are given in Table 3 and in Chapter 6.3.6.

		DEPTH	Q	Q/d$_s$	Q/s	Q$_w$	T	K
WDIS	r	-0,09409	-0,06039	-0,00597	-0,00597	0,02693	-0,00597	0,02697
n=862	p	0,0057	0,0764	0,8611	0,8611	0,4298	0,8611	0,4291
WASL	r	-0,04162	0,05093	0,06755	0,06755	0,07358	0,06755	0,07353
n=862	p	0,2222	0,1352	0,0474	0,0474	0,0308	0,0474	0,0309
ASLW	r	0,03647	-0,14550	-0,13115	-0,13115	-0,11084	-0,13115	-0,11074
n=862	p	0,2849	<0,0001	0,0001	0,0001	0,0011	0,0001	0,0011
ASWA	r	0,04873	-0,15739	-0,14754	-0,14754	-0,12956	-0,14754	-0,12946
n=862	p	0,1529	<0,0001	<0,0001	<0,0001	0,0001	<0,0001	0,0001
BED	r	-0,04967	0,00201	0,00611	0,00611	0,00999	0,00611	0,01005
n=1288	p	0,0747	0,9425	0,8265	0,8265	0,7202	0,8265	0,7187
FOR	r	0,10110	-0,04133	-0,06219	-0,06219	-0,07671	-0,06219	-0,07694
n=1288	p	0,0003	0,1382	0,0256	0,0256	0,0059	0,0256	0,0057
PEA	r	0,03376	-0,03345	-0,03460	-0,03460	-0,03251	-0,03460	-0,03256
n=1288	p	0,2259	0,2302	0,2147	0,2147	0,2437	0,2147	0,2430
CUL	r	-0,05548	0,05539	0,06201	0,06201	0,06162	0,06201	0,06161
n=1288	p	0,0465	0,0469	0,0260	0,0260	0,0270	0,0260	0,0270
SET	r	-0,13583	-0,04529	0,01350	0,01350	0,05545	0,01350	0,05599
n=1288	p	<0,0001	0,1043	0,6283	0,6283	0,0466	0,6283	0,0445
WAT	r	0,07951	0,03701	-0,00016	-0,00016	-0,02296	-0,00016	-0,02306
n=1288	p	0,0043	0,1844	0,9955	0,9955	0,4103	0,9955	0,4083

Table 67. Statistics for well depth, yield and hydraulic parameters in different ASLW groups of all drilled wells in the CF database (n=862). Highest values are in bold. Comparisons significant at the α≤0.05 level between the groups tested with the nonparametric Kruskal-Wallis test for group means and with the median one-way analysis for group medians are indicated by different letters $^{A\text{ and }B}$.

Variable	Parameter	ASLW = ASL-WASL ≤ 5 m n=226 26%	6-10 m n=258 30%	11-15 m n=162 19%	> 15 m n=216 25%
DEPTH m	mean median	92 ^{AB} 82	86 ^{AB} 80	85 ^{B} 76	**96 ^{A}** **86**
Q Lhr^{-1}	mean median	**1981 ^{A}** **1000 ^{A}**	1844 ^{A} 800 ^{A}	1525 ^{A} 800 ^{A}	1224 ^{B} 500 ^{B}
Q/d$_s$ Lhr^{-1}m^{-1}	mean median	47 ^{A} 13 ^{A}	**50 ^{A}** 13 ^{A}	46 ^{A} **16 ^{AB}**	27 ^{B} 8 ^{B}
Q/s m^{2}s^{-1}	mean median	**5,33x10$^{-5\ A}$** 1,56x10$^{-5\ A}$	5,60x10$^{-5\ A}$ 1,56x10$^{-5\ A}$	5,18x10$^{-5\ A}$ **1,80x10$^{-5\ AB}$**	3,07x10$^{-5\ B}$ 9,65x10$^{-6\ B}$
Q$_w$ ms^{-1}	mean median	2,03x10$^{-6\ A}$ 2,07x10$^{-7\ AB}$	2,66x10$^{-6\ A}$ 2,67x10$^{-7\ A}$	**4,05x10$^{-6\ A}$** **2,98x10$^{-7\ A}$**	1,37x10$^{-6\ B}$ 1,34x10$^{-7\ B}$
T m^{2}s^{-1}	mean median	**2,73x10$^{-5\ A}$** 8,24x10$^{-6\ A}$	2,86x10$^{-5\ A}$ 8,26x10$^{-6\ A}$	2,65x10$^{-5\ A}$ **9,48x10$^{-6\ AB}$**	1,59x10$^{-5\ B}$ 5,13x10$^{-6\ B}$
K ms^{-1}	mean median	1,03x10$^{-6\ A}$ 1,10x10$^{-7\ AB}$	1,35x10$^{-6\ A}$ 1,40x10$^{-7\ A}$	**2,04x10$^{-6\ A}$** **1,57x10$^{-7\ A}$**	7,00x10$^{-7\ B}$ 7,20x10$^{-8\ B}$

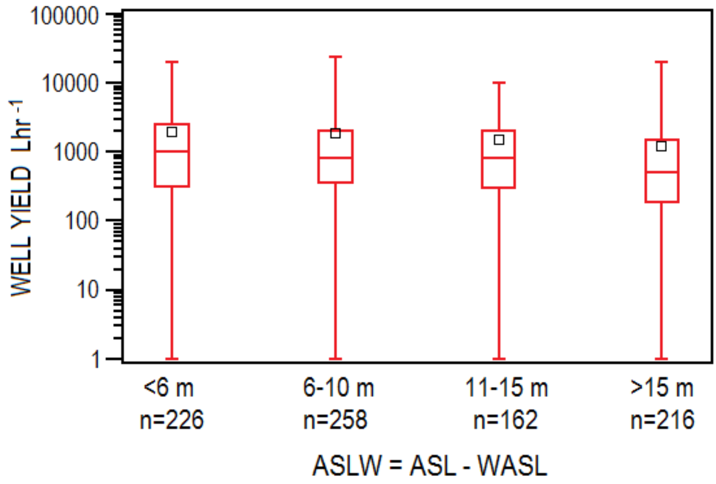


Fig. 105. Box-plot of well yield vs. ASLW for all drilled wells in the CF database (n=862). The maximum and minimum values and the 25th and 75th percentiles of the yield are presented with tops and bottoms on the whiskers and on the boxes, respectively. Median is presented with a horizontal line in the boxes and the mean with a square. The yield of a dry well has been entered into calculations as a nominal figure of 1 Lhr^{-1}.

7.3.6 Well factors in low-yield and high-yield wells

Table 68 shows statistics for well factors in low-yield and high-yield drilled wells in the CF database. Well factors are arranged in different groups beginning with construction factors.

Most well construction factors differ statistically from each other in low-yield and high-yield wells (Table 68). For instance, the first (main) water strike (STR) is 50-60 meters deeper in low-yield wells compared to high-yield wells and, in a corresponding way, low-yield wells are 50-60 meters deeper than high-yield wells. This means that the saturated open well section (SAT) is much longer in low-yield wells, though the groundwater table (GWT) is deeper, than in high-yield wells.

The overburden (OVER) is slightly thicker in high-yield well sites compared to low-yield wells. The relative height difference in a well's catchment area (RELA) and the height difference between the well site and the lowest point in the catchment area (RELAL) are smaller and the relation RERE=RELAH/RELAL is bigger in high-yield wells. Also the distance from the well site to the nearest hilltop (HDIS) is greater in high-yield wells.

The distance to the nearest lineament (LDIS) and the height difference between the well site and the bottom of the nearest lineament (ASLA) are smaller and the relation HLDIS=HDIS/LDIS is greater in high-yield wells. The height difference between the well site and the level of the nearest water course (ASLW) is smaller in high-yield wells compared to that in low-yield wells (Table 68).

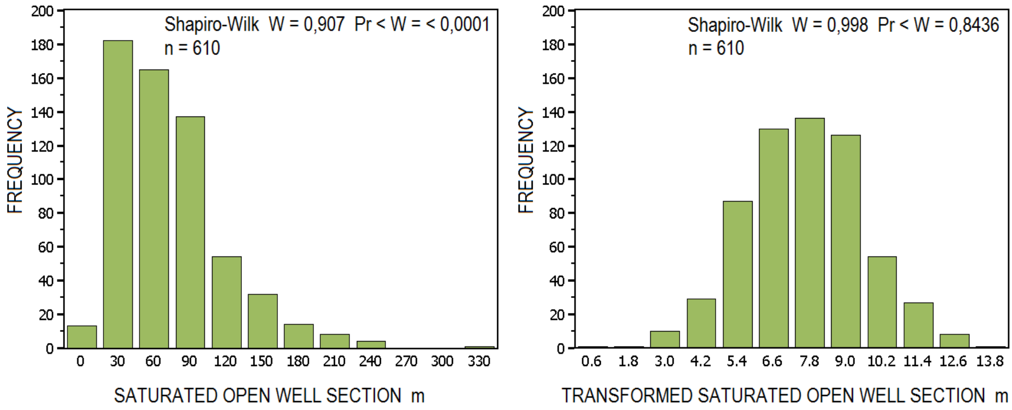
Table 68. Statistics for well factors of low-yield (L; Q≤100 Lhr^{-1}) and high-yield (H; Q≥4,000 Lhr^{-1}) drilled wells in the CF database (test wells included). Comparisons significant at the α≤0.05 level between the groups tested with the nonparametric Kruskal-Wallis test for group means and with the median one-way analysis for group medians are indicated by different letters [A and B].

Variable	Dimension	n		Mean		Std dev		Median		Minimum		Maximum	
		L	H	L	H	L	H	L	H	L	H	L	H
CY	a	139	150	**1986**[A]	**1983**[B]	12	13	**1989**[A]	**1985**[B]	1951	1953	2007	2008
ASL	m a.s.l.	152	165	131	125	37	31	122	119	81	79	246	246
CAS	m	97	94	11	12	7	8	10	10	0	1	40	49
GWT	m	58	94	**7**[A]	**5**[B]	6	5	**6**[A]	**4**[B]	0	0	33	30
GWL	m a.s.l.	58	94	124	119	35	27	114	115	79	79	221	206
SAT	m	58	94	**108**[A]	**55**[B]	55	33	**102**[A]	**49**[B]	9	2	244	223
STR	m	66	93	**84**[A]	**31**[B]	50	28	**80**[A]	**20**[B]	6	3	200	165
SASL	m a.s.l.	66	93	**45**[B]	**96**[A]	57	42	**51**[B]	**99**[A]	-102	-43	175	229
BASL	m a.s.l.	152	165	**7**[B]	**55**[A]	70	49	**9**[B]	**58**[A]	-250	-102	162	179
FRAC	--	69	66	**1**[B]	**2**[A]	0,3	0,8	**1**[B]	**3**[A]	1	1	3	3
DIA	mm	134	142	133	128	21	21	**140**[A]	**115**[B]	100	100	165	166
DEPTH	m	152	165	**124**[A]	**70**[B]	62	39	**120**[A]	**64**[B]	12	9	355	253
WDEM	m^3d^{-1}	62	79	**13**[B]	**23**[A]	39	47	**0,8**[B]	**3,0**[A]	0,1	200	0,1	250
OVER	m	127	140	**5**[B]	**8**[A]	7	9	**3**[B]	**4**[A]	0	0	36	51
UPLIFT	mma^{-1}	152	165	4,8	4,8	0,4	0,4	4,8	4,8	3,9	3,8	6,2	6,2
HSL	m a.s.l.	152	165	**152**[B]	**156**[A]	11	13	150	151	136	132	196	188
TOPO	--	152	165	**2,5**[B]	**2,9**[A]	1,0	0,9	**3,0**[B]	**3,0**[A]	1,0	1,0	4,0	4,0
ASLH	m a.s.l.	152	165	156	152	34	31	151	150	98	98	246	251
ASLL	m a.s.l.	152	165	113	112	29	28	107	107	79	79	196	226
RELA	m	152	165	43	40	19	20	**42**[A]	**36**[B]	8	10	93	104
RELAH	m	152	165	25	27	17	17	21	23	0	2	80	79
RELAL	m	152	165	**19**[A]	**13**[B]	14	10	**16**[A]	**11**[B]	1	0	65	52
RERE	--	152	161	**3,1**[B]	**3,7**[A]	4,6	5,1	**1,3**[B]	**2,3**[A]	0	0,1	27,0	48,0
HDIS	m	70	57	**255**[B]	**319**[A]	150	128	**240**[B]	**310**[A]	5	5	550	560
HASL	m a.s.l.	70	57	162	156	38	32	158	151	105	108	246	240
HAASL	m	70	57	**16**[B]	**22**[A]	13	17	15	20	0	0	58	79
SLOH	°	70	57	3,7	3,8	2,4	2,4	3,1	3,4	0	0	12	11
LDIS	m	57	69	**188**[A]	**123**[B]	132	99	**165**[A]	**90**[B]	5	1	540	430
LPRO	--	57	69	2,0	2,2	0,8	0,7	2,0	2,0	1,0	1,0	3,0	3,0
LFRE	--	57	69	1,6	1,6	0,8	1,0	1,0	1,0	1,0	1,0	4,0	6,0
LLEN	m	57	69	1376	1376	618	545	1120	1130	300	800	3000	3700
LDEN	km^{-1}	57	69	1,4	1,4	0,6	0,5	1,1	1,1	0,3	0,8	3,0	3,7
LLLF	--	57	69	865	925	212	201	880	1000	300	475	1130	1130
LINF	--	57	69	0,3	0,3	0,6	0,7	0	0	0	0	2,0	5,0
LIND	m	12	18	261	251	160	164	220	190	70	40	520	550
LASL	m a.s.l.	57	69	116	121	29	33	109	116	79	79	183	236
ASLA	m	57	69	**13**[A]	**8**[B]	14	8	**10**[A]	**6**[B]	-6	-1	65	37
SLOL	°	57	69	3,4	4,3	3,5	4,0	4,0	3,5	-13,5	-1,4	11,3	19,3
HLDIS	--	32	36	**4,8**[B]	**22,6**[A]	18,9	76,2	**0,9**[B]	**2,5**[A]	0,1	0,4	108,0	450,0
HALA	--	32	36	1,3	1,3	0,2	0,2	1,3	1,3	1,1	1,1	1,8	1,7
HLASL	m	32	36	33	33	16	14	32	30	12	10	79	79
WDIS	m	95	114	228	203	150	146	200	165	20	15	540	540
WASL	m a.s.l.	95	114	105	110	24	27	100	106	79	79	187	236
ASLW	m	95	114	**13**[A]	**8**[B]	11	8	**10**[A]	**7**[B]	-1	-28	65	44
ASWA	--	95	114	**1,1**[A]	**1,1**[B]	0,1	0,1	**1,1**[A]	**1,1**[B]	1,0	1,1	1,7	1,4
BED	%	152	165	5	6	7	8	4	4	0	0	28	40
FOR	%	152	165	54	51	19	20	56	56	8	4	96	92
PEA	%	152	165	3	3	5	6	0	0	0	0	24	40
CUL	%	152	165	18	19	14	15	16	16	0	0	64	64
SET	%	152	165	12	12	10	13	8	8	0	0	48	64
WAT	%	152	165	8	10	13	14	4	4	0	0	64	56

7.4 Multivariate data analysis

7.4.1 Transformations

The Box-Cox method was used to normalize the distributions of all well factors in the CF database (Fig. 106). The Shapiro-Wilk statistics for Box-Cox transformations showed that most variable transformations to normality were not achieved. However, the Box-Cox method could still help to regularize the data (e.g. Draper & Cox 1969).



(A) (B)

Fig. 106. Bar charts showing A) the distribution of saturated open well section and B) the distribution of transformed saturated open well section for all drilled wells in the CF database with depth and yield information (n=610). The Shapiro-Wilk statistics show that the transformed SAT values are normally distributed. The Box-Cox equation used for transformation is $SAT_{tr}=(SAT^{0,27}-1)/0,27$. The horizontal axes are displayed in midpoint values for the bars.

The linear Pearson product moment correlation analysis between the same well factors transformed both by the Box-Cox method and by logarithmic transformation gave very highly significant correlation coefficients (r=0,89639-1,00000; p<0,0001; n=230-1300). This indicated that well factors used in the study are log-normally distributed or, at least, very close to lognormal distribution. For this reason and for the sake of simplicity, logarithmic transformations of different variables were used in multivariable regression analysis (e.g. Mitzenmacher 2004). In factor and discriminant analyses, variable values were transformed with the Box-Cox method and standardized to the mean 0 and variance 1 to avoid of having variables with large variances.

The categorical variables TOPO and LPRO were treated in multivariate analyses as continuous variables. Well construction factors were not included in multivariate analyses because of their limited value in predicting the well production properties in advance, that is until the drilling is completed.

7.4.2 Multivariable regression procedure

Multivariable regression procedure was executed for the lineament wells only. After collinearity diagnostics (condition index < 30, proportion of variation < 50; e.g. Belsley 1991) the following variables were retained in the analyses: logOVER, logTOPO, logASLL, logSLOH, logHDIS, logRELAL, logRELAH, logLDIS, logLPRO, logLFRE, logLLEN, logLASL, logBED, logFOR, logPEA, logWAT, logHLDIS, logHALA, logHLASL and logRERE (Table 3). Test wells were included in the analyses.

All variables included in the models had F statistics significant at the ≤0,15 level for entry into the models with the method forward. The F statistics for the overall models were very highly significant (p<0,0001), indicating that the models explained a significant portion of the variation in the data. There were no discernible trends in scatterplots of the fitted values against the standardized residuals. The F statistics for each variable and the parameter estimates with significance probabilities of the t statistics likewise the partial and cumulative coefficients of determination R^2 and the Mallow´s C_p-values for the models are presented in Tables 69-74.

Table 69. Results from multivariable regression analysis with forward method for dependent variable logQ (n=200).

Step	Variable entered	F value	Pr > F	Parameter estimate	t value	Pr > | t |	Partial R^2	Model R^2	C_p
	Intercept			2,22556	15,78	<0,0001			
1	LogHLDIS	23,68	<0,0001	0,24399	1,92	0,0558	0,1068	0,1068	17,5154
2	LogLPRO	7,79	0,0058	0,85768	2,77	0,0062	0,0340	0,1408	11,3963
3	LogTOPO	3,16	0,0772	0,58519	1,78	0,0772	0,0136	**0,1544**	10,1414

From the parameter estimates, the fitted model for the well yield (Lhr^{-1}) can be written as logQ=2,23+0,24logHLDIS+0,86logLPRO+0,59logTOPO. The equation indicates that the value of the response variable increases with the values of the regressor variables. That is, highest well yields are met near short lineaments away from hilltops. However, the model only accounts for 15% of the variance in logarithmic well yield.

Table 70. Results from multivariable regression analysis with forward method for dependent variable $\log(Q/d_s)$ (n=200).

Step	Variable entered	F-value	Pr > F	Parameter estimate	t value	Pr > | t |	Partial R^2	Model R^2	C_p
	Intercept			1,21310	3,01	0,0030			
1	LogTOPO	30,14	<0,0001	0,87133	2,20	0,0292	0,1321	0,1321	23,1358
2	LogLPRO	8,86	0,0033	1,02525	2,91	0,0040	0,0373	0,1695	15,7088
3	LogHLDIS	3,52	0,0623	0,15576	1,05	0,2928	0,0146	0,1841	14,0138
4	LogBED	2,81	0,0952	0,18355	2,76	0,0064	0,0116	0,1957	13,0857
5	LogSLOH	4,76	0,0303	-0,55352	-2,52	0,0126	0,0193	0,2150	10,2214
6	LogWAT	3,81	0,0525	-0,18585	-2,74	0,0067	0,0152	0,2301	8,3884
7	LogRELAL	5,51	0,0200	-0,60778	-2,35	0,0200	0,0215	**0,2516**	4,9688

The multivariable regression equation for the normalized well yield ($Lhr^{-1}m^{-1}$) can be written as $\log(Q/d_s)$=1,21+0,87logTOPO+1,03logLPRO+0,16logHLDIS+0,18logBED–0,55logSLOH–0,19logWAT–0,61logRELAL. It accounts for 25% of the variance in logarithmic normalized yield. Note the negative signs in front of the parameter estimates of some regressor variables.

Table 71. Results from multivariable regression analysis with forward method for dependent variable log(Q/s) (n=200).

Step	Variable entered	F-value	Pr > F	Parameter estimate	t value	Pr > | t |	Partial R^2	Model R^2	C_p
	Intercept			-4,72305	-11,87	<0,0001			
1	LogTOPO	30,14	<0,0001	0,85902	2,20	0,0292	0,1321	0,1321	23,1358
2	LogLPRO	8,86	0,0033	1,01077	2,91	0,0040	0,0373	0,1695	15,7088
3	LogHLDIS	3,52	0,0623	0,15356	1,05	0,2928	0,0146	0,1841	14,0138
4	LogBED	2,81	0,0952	0,18096	2,76	0,0064	0,0116	0,1957	13,0857
5	LogSLOH	4,76	0,0303	-0,54570	-2,52	0,0126	0,0193	0,2150	10,2214
6	LogWAT	3,81	0,0525	-0,18322	-2,74	0,0067	0,0152	0,2301	8,3884
7	LogRELAL	5,51	0,0200	-0,59919	-2,35	0,0200	0,0215	**0,2516**	4,9688

The fitted model for the specific capacity (m^2s^{-1}) can be written as log(Q/s)=-4,72+0,86logTOPO+1,01logLPRO+0,15logHLDIS+0,18logBED-0,55logSLOH–0,18log WAT–0,60logRELAL. It accounts for 25% of the variance in logarithmic specific capacity.

Table 72. Results from multivariable regression analysis with forward method for dependent variable $\log Q_w$ (n=200).

Step	Variable entered	F-value	Pr > F	Parameter estimate	t value	Pr > | t |	Partial R^2	Model R^2	C_p
	Intercept			-6,30822	-13,31	<0,0001			
1	LogTOPO	32,57	<0,0001	1,35661	3,56	0,0005	0,1412	0,1412	24,3713
2	LogLPRO	8,33	0,0043	1,22618	2,96	0,0034	0,0348	0,1761	17,4283
3	LogBED	3,84	0,0516	0,23938	3,07	0,0025	0,0158	0,1919	15,3698
4	LogSLOH	5,86	0,0164	-0,74153	-2,87	0,0046	0,0236	0,2155	11,3233
5	LogWAT	3,93	0,0489	-0,24283	-3,05	0,0026	0,0156	0,2310	9,3291
6	LogRELAL	9,30	0,0026	-0,89799	-3,05	0,0026	0,0354	**0,2664**	2,2539

The multivariable regression equation for well productivity (ms^{-1}) can be written as $\log Q_w$=-6,31+1,36logTOPO+1,23logLPRO+0,24logBED–0,74logSLOH-0,24logWAT–0,90logRELAL. It accounts for 27% of the variance in logarithmic well productivity.

Table 73. Results from multivariable regression analysis with forward method for dependent variable logT (n=200).

Step	Variable entered	F-value	Pr > F	Parameter estimate	t value	Pr > | t |	Partial R^2	Model R^2	C_p
	Intercept			-5,00139	-12,74	<0,0001			
1	LogTOPO	30,14	<0,0001	0,84729	2,20	0,0292	0,1321	0,1321	23,1358
2	LogLPRO	8,86	0,0033	0,99697	2,91	0,0040	0,0373	0,1695	15,7088
3	LogHLDIS	3,52	0,0623	0,15146	1,05	0,2928	0,0146	0,1841	14,0138
4	LogBED	2,81	0,0952	0,17849	2,76	0,0064	0,0116	0,1957	13,0857
5	LogSLOH	4,76	0,0303	-0,53825	-2,52	0,0126	0,0193	0,2150	10,2214
6	LogWAT	3,81	0,0525	-0,18072	-2,74	0,0067	0,0152	0,2301	8,3884
7	LogRELAL	5,51	0,0200	-0,59101	-2,35	0,0200	0,0215	**0,2516**	4,9688

The fitted model for transmissivity (m^2s^{-1}) can be written as logT=-5,00+0,85logTOPO +1,00logLPRO+0,15logHLDIS+0,18logBED–0,54logSLOH–0,18logWAT–0,59log RELAL. It accounts for 25% of the variance in logarithmic transmissivity.

Table 74. Results from multivariable regression analysis with forward method for dependent variable logK (n=200).

Step	Variable entered	F-value	Pr > F	Parameter estimate	t value	Pr > | t |	Partial R^2	Model R^2	C_p
	Intercept			-6,58705	-14,05	<0,0001			
1	LogTOPO	32,58	<0,0001	1,34162	3,56	0,0005	0,1413	0,1413	24,3729
2	LogLPRO	8,32	0,0044	1,21194	2,96	0,0034	0,0348	0,1761	17,4409
3	LogBED	3,84	0,0515	0,23681	3,07	0,0025	0,0158	0,1919	15,3802
4	LogSLOH	5,87	0,0164	-0,73390	-2,87	0,0046	0,0236	0,2155	11,3247
5	LogWAT	3,93	0,0489	-0,24024	-3,05	0,0026	0,0156	0,2311	9,3294
6	LogRELAL	9,32	0,0026	-0,88884	-3,05	0,0026	0,0354	**0,2665**	2,2382

From the parameter estimates, the fitted model for hydraulic conductivity (ms^{-1}) can be written as logK=-6,59+1,34logTOPO+1,21logLPRO+0,24logBED–0,73logSLOH-0,24log WAT-0,89logRELAL. It accounts for 27% of the variance in logarithmic hydraulic conductivity.

As was expected, the multivariable regression equations for Q/d_s, Q/s and T and those for Q_w and K are nearly identical, except for the intercepts. The well yield and hydraulic parameters modeled with multivariable regression have moderate regressions with the actual well yield and calculated hydraulic parameters of the same wells in the CF database (Fig. 107). Although the median and mean values of both calculated and modeled values are in a similar range, the models fail to predict low and high values of well yield and hydraulic parameters.

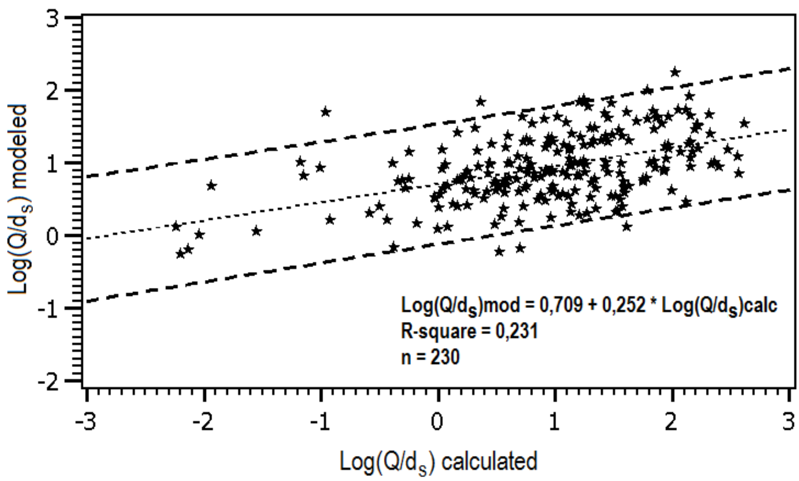


Fig. 107. Regression between modeled and calculated log(Q/d_s) in $Lhr^{-1}m^{-1}$ with 95% confidence limits for individual predicted values for all lineament wells in the CF database (n=230).

7.4.3 Factor procedure

Factor procedure was first performed for the lineament wells as an R-mode analysis with principal components method. The following 15 variables were included in the analyses: OVER, TOPO, ASLL, ASLH, SLOH, RELAL, RELAH, LDIS, HLDIS, LLEN, BED, FOR and WAT (Table 3). Well yield and hydraulic parameters were separately entered into successive analyses with these well variables. Test wells were included in the analyses.

The first two factors with well yield explain 54% of the variation in the data and four factors in all have eigenvalues > 1 with cumulative account of 69% of the total variance (Table 75).

Table 75. Eigenvalues of the correlation matrix for 14 variables including well yield for lineament wells in the CF database (n=203, test wells included).

Factor No	Eigenvalue	Difference	Proportion	Cumulative
1	5,01606268	2,50239521	0,3583	0,3583
2	2,51366747	1,35397470	0,1795	0,5378
3	1,15969277	0,13941151	0,0828	0,6207
4	1,02028125	0,12970224	0,0729	0,6936
5	0,89057901	0,09436164	0,0636	0,7572
6	0,79621738	0,15536137	0,0569	0,8140
7	0,64085601	0,04221331	0,0458	0,8598
8	0,59864270	0,20308237	0,0428	0,9026
9	0,39556033	0,07423914	0,0283	0,9308
10	0,32132119	0,02917022	0,0230	0,9538
11	0,29215096	0,07639936	0,0209	0,9746
12	0,21575160	0,10287012	0,0154	0,9901
13	0,11288148	0,08654629	0,0081	0,9981
14	0,02633518		0,0019	1,0000

In a two-factor pattern well YIELD enters into factor1 with seven other variables, for example HLDIS and TOPO with high positive factor loadings and LDIS and RELAL with high negative loadings (Table 76, Fig. 108). Factor1 might be named as a lineament factor and factor2 as a catchment factor.

Table 76. Varimax rotated orthogonal R-mode two-factor pattern with well yield for lineament wells in the CF database (n=203, test wells included).

Variable	Factor1	Factor2
HLDIS	**0,88709**	-0,06250
TOPO	**0,86751**	-0,09096
RELAH	**0,66861**	-0,42324
SLOH	**0,50367**	-0,34615
OVER	**0,49710**	0,20791
YIELD	**0,42369**	-0,01729
RELAL	**-0,60312**	0,47801
LDIS	**-0,89807**	0,11765
ASLL	-0,11638	**0,86304**
ASLH	-0,00708	**0,79547**
FOR	0,04979	**0,72377**
LLEN	0,30958	**-0,40826**
WAT	0,02511	**-0,66210**
BED	0,19412	**-0,66246**
Variance explained	3,9873666	3,5423636

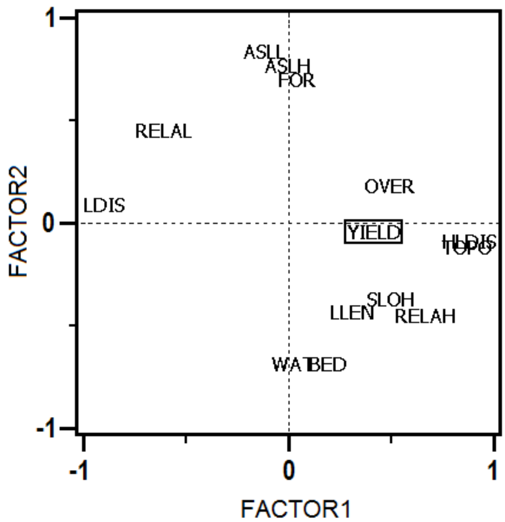


Fig. 108. Varimax rotated orthogonal R-mode two-factor pattern for variables with well yield for all lineament wells in the CF database (n=203, test wells included).

Final communality estimates for each variable in the two-factor pattern vary from 0,18 for YIELD to 0,82 for LDIS (Table 77).

Table 77. Final communality estimates for each variable in a two-factor pattern for lineament wells in the CF database (n=203, test wells included).

Variable	Communality estimate	Variable	Communality estimate
HLDIS	0,79083171	ASLL	0,75838204
TOPO	0,76085206	ASLH	0,63282774
RELAH	0,62617784	FOR	0,52632830
SLOH	0,37349979	LLEN	0,26251705
OVER	0,29034038	WAT	0,43901287
YIELD	0,17981357	BED	0,47652942
RELAL	0,59224696		
LDIS	0,82037042	Total	7,529730

In the R-mode four-factor pattern YIELD moves from factor1 to factor4 with LLEN. However, it still has a moderate factor loading in factor1, too (Table 78).

Factor results for hydraulic parameters are almost identical with well yield. Hence, they are not presented separately. On the other hand, when the well yield and hydraulic parameters were included in the analyses simultaneously, they only arranged into one strict factor leaving out other variables. They also remained there despite the order of the factor pattern examined.

The Q-mode factor analysis was executed for high-yield and low-yield wells only. In this analysis three of 14 factors had eigenvalues greater than one; the two first factors explained 58% of the variation in the data. The number of wells was one third of the previous analysis. In the two-factor pattern the factor variables remained the same as in Table 76 except the variable RELAL that moved to factor2.

Table 78. Varimax rotated orthogonal four-factor pattern with well yield for lineament wells in the CF database (n=203, test wells included).

Variable	Factor1	Factor2	Factor3	Factor4
HLDIS	**0,81129**	-0,01687	0,21458	0,32110
TOPO	**0,79947**	-0,03647	0,27116	0,23402
OVER	**0,68298**	0,05794	-0,07847	-0,29567
RELAL	**-0,65116**	0,54311	-0,11888	-0,12343
LDIS	**-0,76459**	0,02139	-0,29054	-0,41264
ASLH	-0,03575	**0,84116**	-0,09226	-0,08330
FOR	0,03860	**0,74732**	-0,10597	-0,06860
ASLL	0,00906	**0,71903**	-0,50834	-0,07710
WAT	0,10002	**-0,74677**	0,02130	0,00221
SLOH	0,30109	-0,06739	**0,81061**	-0,09527
RELAH	0,45134	-0,16988	**0,73180**	0,13029
BED	-0,12633	-0,35969	**0,62754**	0,43553
YIELD	0,26136	0,03059	-0,10485	**0,74549**
LLEN	0,08520	-0,23659	0,30579	**0,51386**
Variance explained	3,1711724	2,8607853	2,1938010	1,4839455

High-yield wells dominate the positive side of the lineament factor (1) whereas most low-yield wells stay at the negative side of the same axis (Fig. 109). As expected, the well groups for most part overlap each other.

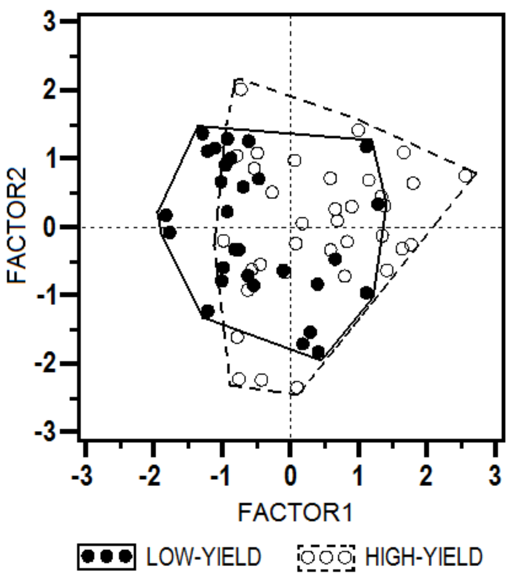


Fig. 109. Varimax rotated orthogonal Q-mode two-factor pattern with well yield for low-yield (Q≤100 Lhr^{-1}) and high-yield (Q≥4,000 Lhr^{-1}) lineament wells in the CF database (n=65, test wells included).

The factor procedure was also conducted for non-lineament wells in the CF database (n=214). The variables included in the analysis are shown in Table 79. In this analysis four of 12 factors in all had eigenvalues higher than one. The first two factors explained 51% of the variation in the data.

In the two-factor pattern YIELD appears with FOR, OVER and BED, the last one being negative (Table 79, Fig. 110). Final communality estimates for the two-factor pattern are presented in Table 80. In the four-factor pattern YIELD shows up with OVER as their own factor (Table 81). OVER has a relatively high factor loading with YIELD through two- to four-factor patterns. Interestingly, the variables OVER and BED change their sign compared to the lineament wells. This indicates that the thickness of overburden is of higher importance when explaining the well yields in non-lineament group.

Table 79. Varimax rotated R-mode orthogonal two-factor pattern with well yield for non-lineament wells in the CF database (n=214, test wells included).

Variable	Factor1	Factor2
RERE	**0,93379**	0,17177
RELAH	**0,83895**	0,12667
SLOH	**0,65198**	-0,04683
WAT	**0,61912**	-0,15318
TOPO	**0,57844**	0,11564
HDIS	**0,44381**	0,44079
ASLL	**-0,61516**	0,39748
RELAL	**-0,84275**	-0,15497
FOR	-0,40092	**0,58851**
OVER	0,22269	**0,51701**
YIELD	0,22020	**0,39583**
BED	0,27497	**-0,74641**
Variance explained	4,3388424	1,7883269

Table 80. Final communality estimates for each variable in a two-factor pattern for non-lineament wells in the CF database (n=214, test wells included).

Variable	Communality estimate	Variable	Communality estimate
RERE	0,90146425	FOR	0,50707860
RELAH	0,71987991	OVER	0,31689010
SLOH	0,42727131	**YIELD**	0,20516673
WAT	0,40678001	BED	0,63274042
TOPO	0,34797135		
HDIS	0,39126206		
ASLL	0,53641664		
RELAL	0,73424792	Total	6,127169

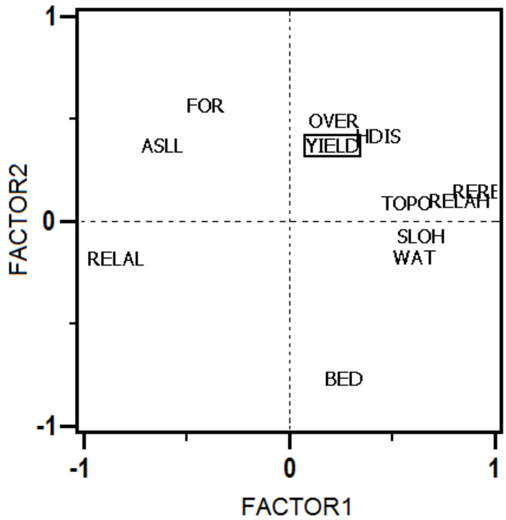


Fig. 110. Varimax rotated orthogonal R-mode two-factor pattern for variables with well yield for non-lineament wells in the CF database (n=214, test wells included).

Table 81. Varimax rotated orthogonal R-mode four-factor pattern with well yield for non-lineament wells in the CF database (n=214, test wells included).

Variable	Factor1	Factor2	Factor3	Factor4
SLOH	**0,86749**	-0,14103	-0,06110	-0,05181
RELAH	**0,82016**	-0,15332	0,35597	0,03026
RERE	**0,76761**	-0,27139	0,45421	0,21117
RELAL	**-0,54827**	0,35228	-0,47718	-0,31765
FOR	-0,00488	**0,84032**	0,01744	-0,05453
ASLL	-0,38322	**0,59144**	-0,16345	0,11602
WAT	0,31325	**-0,55733**	0,18068	0,26750
BED	0,12398	**-0,61868**	-0,05174	-0,50339
HDIS	0,10066	0,12553	**0,86402**	0,07743
TOPO	0,15620	-0,19643	**0,78585**	-0,04349
YIELD	-0,00226	-0,12788	0,06344	**0,80836**
OVER	0,40871	0,26991	-0,04759	**0,49234**
Variance explained	2,7769845	2,1340123	1,9971733	1,3942197

Factor results for hydraulic parameters in non-lineament wells are almost identical with well yield. Hence, they are not presented separately. As with lineament wells, the well yield and hydraulic parameters arranged into one strict factor and left out other variables when included simultaneously in the analyses.

Q-mode factor analysis was executed for high-yield and low-yield wells only. In this analysis three of 12 factors had eigenvalues greater than one; the two first factors explained 56% of the variation in the data. The number of wells was about half of the previous analysis. The factor groupings for two-factor pattern are shown in Table 82.

Table 82. Varimax rotated Q-mode orthogonal two-factor pattern with well yield for low-yield (Q≤100 Lhr^{-1}) and high-yield (Q≥4,000 Lhr^{-1}) non-lineament wells in the CF database (n=109, test wells included).

Variable	Factor1	Factor2
RERE	**0,74433**	-0,57749
RELAH	**0,70741**	-0,50714
TOPO	**0,68418**	-0,20405
HDIS	**0,68099**	0,00921
OVER	**0,62295**	0,14518
YIELD	**0,58337**	0,12441
RELAL	**-0,65330**	0,52969
FOR	0,09397	**0,78491**
ASLL	-0,11309	**0,77090**
SLOH	0,44369	**-0,50505**
WAT	0,06022	**-0,65957**
BED	-0,02115	**-0,68970**
Variance explained	3,3640576	3,3256796

High-yield wells dominate the positive side of the topography factor (1) whereas most low-yield wells stay at the negative side of the same axis (Fig. 111). As expected, the well groups for most part overlap each other.

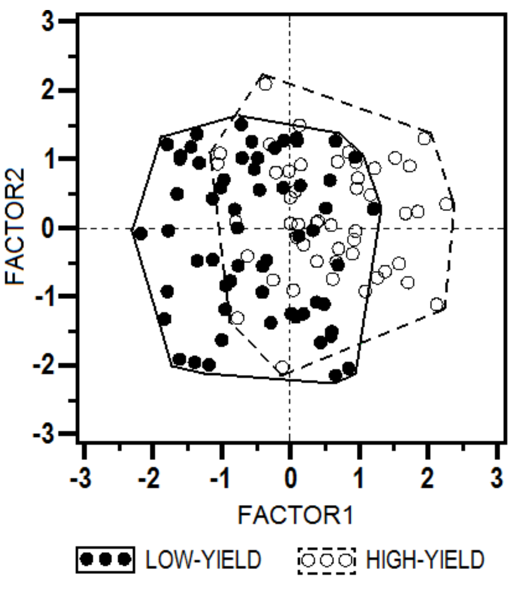


Fig. 111. Varimax rotated orthogonal Q-mode two-factor pattern with well yield for low-yield (Q≤100 Lhr^{-1}) and high-yield (Q≥4,000 Lhr^{-1}) non-lineament wells in the CF database (n=109, test wells included).

7.4.4 Discriminant procedure

The discriminant procedure was used to find the linear combination of well factors, which best separates low-yield and high-yield wells from each other. In this analysis the dependent variable YIELD was used as a categorical class variable. The independent variables included in the final analysis for lineament wells were OVER, TOPO, ASLH, ASLL, RELAL, RELAH, LDIS, LPRO, LLEN, BED, FOR and WAT (Table 3). Test wells were included in the analyses.

The method stepwise was used for selecting variables into the model. The significance level for each variable to enter into the model was ≤ 0,15. The use of a moderate significance level, in the range of 0,10 to 0,25, often performs better than the use of a much larger or a much smaller significance level (e.g. SAS 1990b). Each variable´s discriminatory power in the model was measured by Wilks´ lambda, the likelyhood ratio criterion. As a result of the stepwise discriminant analysis, four independent variables, LDIS, BED, ASLL and RELAL, were picked up (Table 83) and used in the actual discriminant analysis (Table 84).

Table 83. Result of the stepwise discriminant analysis for low-yield (Q≤100 Lhr^{-1}, n=51) and high-yield (Q≥4,000 Lhr^{-1}, n=67) lineament wells in the CF database (test wells included).

Step	Variable entered	Partial R-square	F value	Pr > F	Wilks' Lambda	Pr < Lambda
1	LDIS	0,0849	10,76	0,0014	0,91508962	0,0014
2	BED	0,0311	3,69	0,0573	0,88666081	0,0010
3	ASLL	0,0239	2,79	0,0975	0,86546979	0,0009
4	RELAL	0,0267	3,10	0,0810	0,84235949	0,0006

Around 25% of the high-yield wells and 35% of the low-yield wells were misclassified when resubstituting them using the linear discriminant function of the four variables LDIS, BED, ASLL and RELAL. The total error count was 29%, which is rather high (Table 84).

Table 84. Results of the discriminant analysis for low-yield (Q≤100 Lhr^{-1}) and high-yield (Q≥4,000 Lhr^{-1}) lineament wells in the CF database (n=126, test wells included).

From class YIELD	Into class YIELD		
	LOW	HIGH	Total
LOW	37 65%	20 35%	57 100%
HIGH	17 25%	52 75%	69 100%
Total	54 43%	72 57%	126 100%
Priors Rate	0,45 35%	0,55 25%	**29%**

The corresponding results for low-yield and high-yield non-lineament wells are presented in Tables 85 and 86. The independent variables used in the analysis were ASL, OVER, UPLIFT, HSL, TOPO, ASLH, ASLL, RELA, RERE, RELAL, RELAH, BED, FOR and WAT (Table 3).

Table 85. Results of the stepwise discriminant analysis for low-yield ($Q \leq 100$ Lhr^{-1}, n=76) and high-yield ($Q \geq 4{,}000$ Lhr^{-1}, n=72) non-lineament wells in the CF database (test wells included).

Step	Variable entered	Partial R-square	F value	Pr > F	Wilks' Lambda	Pr < Lambda
1	OVER	0,0825	13,12	0,0004	0,91753500	0,0004
2	RELAL	0,0334	5,02	0,0266	0,88685177	0,0002
3	BED	0,0190	2,79	0,0970	0,86998844	0,0002

High-yield wells have low RELAL- and BED-values and high OVER-values in contrary to low-yield wells.

One third of the wells in both groups were misclassified when resubstituting them using the linear discriminant function of the three variables OVER, RELAL and BED. The total error count was 33%, which is considered rather high (Table 86).

Table 86. Results of the discriminant analysis for low-yield ($Q \leq 100$ Lhr^{-1}) and high-yield ($Q \geq 4{,}000$ Lhr^{-1}) non-lineament wells in the CF database (n=149, test wells included).

From class YIELD	Into class YIELD		
	LOW	HIGH	Total
LOW	51 67%	25 33%	76 100%
HIGH	24 33%	49 67%	73 100%
Total	75 50%	74 50%	149 100%
Priors Rate	0,51 33%	0,49 33%	 **33%**

The categorical variable FRAC was shown to be an important factor affecting well yield and hydraulic parameters (Chapter 7.3.1). Because of this, it was included in the discriminant procedure as a dependent variable to find the linear combination of independent variables, which best separates lineament wells from each other in low, medium and highly fractured bedrock well sites. The stepwise method picked up the variables LDIS, ASLL, OVER, BED and LLEN (Table 87).

Table 87. Results of the stepwise discriminant analysis for the dependent variable FRAC of the lineament wells in the CF database (n=165, test wells included).

Step	Variable entered	Partial R-square	F value	Pr > F	Wilks' Lambda	Pr < Lambda
1	LDIS	0,1675	16,29	<,0001	0,83252941	<0,0001
2	ASLL	0,0645	5,55	0,0047	0,77883314	<0,0001
3	OVER	0,0469	3,93	0,0215	0,74234296	<0,0001
4	BED	0,0274	2,24	0,1096	0,72197968	<0,0001
5	LLEN	0,0246	1,99	0,1395	0,70420129	<0,0001

Three quarters of the wells in the group ‘medium’ and nearly 60% of the wells in the group ‘high’ were misclassified when resubstituting them using the linear discriminant function of the independent variables. Because of this, the total error count went high, 42%, though the misclassifying percent in the biggest group ‘low’ was only 15% (Table 88).

Table 88. Results of the discriminant analysis for the dependent variable FRAC of the lineament wells in the CF database (n=165, test wells included).

From class FRAC	**Into class FRAC**			
	LOW	**MEDIUM**	**HIGH**	**Total**
LOW	70 85%	8 10%	4 5%	82 100%
MEDIUM	33 60%	14 25%	8 15%	55 100%
HIGH	7 25%	9 32%	12 43%	28 100%
Total	110 67%	31 19%	24 14%	165 100%
Priors Rate	0,50 15%	0,33 75%	0,17 57%	 **42%**

8 DISCUSSION AND CONCLUSIONS

8.1 Drilled well yield and hydraulic properties in Central Finland

In Central Finland, some 9,000 bedrock wells have been drilled since the late 1940's. Just over a quarter of them have been compiled in the CF database. The database wells show a median depth and yield of 73 m and 700 Lhr^{-1}, respectively. However, recalling the depth and yield progress during the last 30 years, a typical drilled well in Central Finland today might be a 110 m deep well with a 10 m long casing and a (short-term) yield of 600 Lhr^{-1} of good-quality groundwater with a drawdown of 15 m. An average well needs to be equipped with a submersible pump and it can easily satisfy the water demand in normal household activities.

At present, the overall groundwater extraction from bedrock wells in Central Finland is around 10,000 m^3d^{-1}. This is estimated as the total number of wells used in the water supply multiplied with the average water consumption per well. If the recharge area for bedrock groundwater is estimated to be two thirds of the land area of Central Finland (Chapter 7.3.2) and if the amount of groundwater recharge is estimated to lie around 5% of the average annual rainfall of 650 mm a^{-1} (Chapter 5.1), then the potential groundwater production from bedrock aquifers is around 1,000,000 m^3d^{-1}. This means that the amount of bedrock groundwater used in daily water supply is only 1% of the total bedrock groundwater potential in Central Finland. Moreover, in the CF database, the proportion of high-yield wells ($\geq$ 4,000 Lhr^{-1}) is approximately 12% of all drilled wells. This means that at present there are nearly one thousand high-yield wells in Central Finland with total (short-term) capacity of around 130,000 m^3d^{-1}. This is twice that of the total water demand in Central Finland.

These assessments show that bedrock groundwater is a capacious water resource in Central Finland. It could more often be used in the common water supply of small villages and towns not to mention single households and farms in rural areas. Moreover, bedrock groundwater can provide a valuable source for water supply of large communities in various times of crisis. But as Martin and van de Giesen (2005) state, groundwater availability is not only determined by the amount of groundwater potential, but also by the accessibility, exploitability and supply reliability of the resource and whether its development is worthwhile (e.g. water demand and quality issues).

Hydraulic fracturing has proven to be a successful method in increasing the yield of dry and low-yield wells in the crystalline bedrock of Central Finland. Hydraulic fracturing should be used whenever enough water has not been encountered but at a depth of 120…150 m at the latest, which may be considered as the maximum recommendable depth for drilled wells in Central Finland. Test well results from Central Finland suggest that hydraulic fracturing can successfully be used in the development of medium-yield wells, too.

In Central Finland, typical normalized well yields (Q/d_s) are about one quarter of the corresponding specific capacities (Q/s) and less than half of the corresponding transmissivities (T). Short-term specific capacities and well productivities (Q_w) are in turn nearly twice that of transmissivity and hydraulic conductivity (K) values, respectively (Table 89). Hence, the Q/d_s, Q/s and Q_w are not, as such, well enough suited for T and K approximation. Yet, adjusting the hydraulic parameters with their mutual empirical relations, as accomplished in this study, the Q/d_s, Q/s and Q_w can be used as representative estimates for bulk T and K in regional studies of fractured rock aquifers. Concurrently it must be kept in mind that the accuracy and precision in the determination of the well yields (if any) are fundamental prerequisites for hydraulic parameter estimations, the importance of which cannot be overemphasized. Moreover, it would be of great value if relatively fast and inexpensive specific capacity determinations were more often carried out in connection with water well drillings. Empirical equations, which are derived in one geologic setting for estimating T or K, are seldom directly transferable to another.

Table 89. Summary statistics for well depth and production properties of all drilled wells with depth and yield information in the Central Finland drilled well database (n=1300). Median values are in bold. The yield of a dry well has been entered into calculations as a nominal figure of 1 Lhr^{-1}.

Parameter	Dimension	Central Finland drilled well database (n=1300)				
		Mean	Std dev	Median	Minimum	Maximum
Depth	m	88	50	**80**	9	355
Q	Lhr^{-1}	1594	2306	**700**	1	24000
Q/d_s	$Lhr^{-1}m^{-1}$ m^2s^{-1}	46 $1,28x10^{-5}$	161 $4,47x10^{-5}$	**12** **$3,31x10^{-6}$**	0,004 $1,18x10^{-9}$	3500 $9,72x10^{-4}$
Q/s	$Lhr^{-1}m^{-1}$ m^2s^{-1}	186 $5,16x10^{-5}$	628 $1,74x10^{-4}$	**50** **$1,39x10^{-5}$**	0,02 $5,54x10^{-9}$	13530 $3,76x10^{-3}$
Q_w	md^{-1} ms^{-1}	0,43 $4,95x10^{-6}$	6,02 $6,97x10^{-5}$	**0,02** **$2,08x10^{-7}$**	0,000002 $2,36x10^{-11}$	162,36 $1,88x10^{-3}$
T	m^2d^{-1} m^2s^{-1}	2,28 $2,63x10^{-5}$	7,44 $8,62x10^{-5}$	**0,63** **$7,34x10^{-6}$**	0,0003 $3,26x10^{-9}$	159,14 $1,84x10^{-3}$
K	md^{-1} ms^{-1}	0,21 $2,47x10^{-6}$	2,96 $3,42x10^{-5}$	**0,01** **$1,10x10^{-7}$**	0,000001 $1,39x10^{-11}$	79,57 $9,21x10^{-4}$

Drilled well depths and yields in Central Finland are in a similar range than in most fractured non-carbonate and non-volcanic crystalline rocks around the world. Median well yields around 600-700 Lhr^{-1} have been reported, for example, from Sweden, Norway, Canada, Nigeria and Sri Lanka and in parts of USA and Brazil despite the different lithologies, climates and tectonic histories. Sander et al. (1996) state that in Ghana the overall drilling success rates for 600 Lhr^{-1} have averaged approximately 55 percent. This is exactly the same as in this study. The similarity in well yields of crystalline rock areas has also been noticed by Krásny and Sharp (2007b) and Banks et al. (2010). Also Black (1987) suggests that single-borehole data from crystalline rocks around the world are similar and independent of detailed geology and test techniques.

The similarity holds true with hydraulic parameters, too. The median normalized yield (Q/d) is nearly the same in Swedish (10 $Lhr^{-1}m^{-1}$) and Norwegian drilled wells (12 $Lhr^{-1}m^{-1}$) as well as in Central Finland (11 $Lhr^{-1}m^{-1}$). In addition to Central Finland, hydraulic conductivities ranging commonly between 10^{-8}-10^{-6} ms^{-1} have been met in many crystalline rock areas, for instance in Sweden and Norway, in northeastern USA and in the Canadian Shield. Also the compilation of Stober and Bucher (2007) suggests that K values for fractured basement to about 1 km depth are typically restricted to the range from 10^{-8} to 10^{-6} ms^{-1}. Based on well yield distributions and typical drawdowns, Banks et al (2010) have set forth median T and K values for Fennoscandian bedrock wells, which are similar with this study.

In some basement areas, the median well yields have been reported to be clearly higher than in Central Finland (e.g. New Hampshire, North Carolina and Virginia in the United States) but also lower yields have been presented (e.g. Paraiba State in Brazil, Colorado in USA). Drawing comparisons between different crystalline rock areas is complicated by the fact that dry and low-yield wells are not necessarily included in the available datasets.

The observed similarities in well yields and hydraulic properties indicate that these parameters are largely governed by factors associated with crystalline rocks themselves (mainly fracturing characteristics) and that the recharge and (overburden) storativity, for example, are not primary factors controlling short-term yields in crystalline rock environments (Krásny & Sharp 2007b, Banks et al. 2010). Also Chambel et al. (2003) suggest that the dominating effect of rock fracturing results in similar transmissivities in crystalline basement areas. Locally, however, crystalline rock aquifers almost always are highly varying complex and heterogeneous groundwater environments (e.g. Mabee 1999).

In crystalline carbonates and sandstones the well yields and hydraulic values may be 2-3 orders of magnitude higher than in igneous and metamorphic rocks. For example, in wells of the Dammam hard rock formation (limestone and dolomite) in Kuwait the median specific capacity is $2{,}8 \times 10^{-3}$ m^2s^{-1} and the median transmissivity $2{,}2 \times 10^{-3}$ m^2s^{-1} (Sayed & Al-Ruwaih 1995). An average transmissivity of $1{,}1 \times 10^{-2}$ m^2s^{-1} was obtained by Rayne et al. (2001) for a Silurian dolomite aquifer in Wisconsin, USA. In the Kristianstadsslätten plain in Scania, southern Sweden, drilled well yields up to 450,000 Lhr^{-1} have been met in the porous Ordovician aquifer called 'glaukonitsand' (Gustafsson 1995). In Australia, sandstones constitute the most productive rocks with yields ranging from 700 to 216,000 Lhr^{-1} (Bestow 1990).

8.2 Factors affecting well yield and hydraulic properties

In the Central Finland drilled well database most well construction factors correlate statistically significantly with well yield and hydraulic parameters. Their significance, however, largely fades away when partialled with well depth and construction year. Moreover, their predictive value in well siting is fairly limited. Yet some well construction results merit a brief discussion.

The introduction of the fast and efficient DTH drilling technique has indisputably increased the well depths in Central Finland during the last three decades. At the same time the median well yields have remained unchanged or even decreased. This may sound contradictory but it can be explained, for the most part, by the deeper casing through surficial water bearing fractures and by the fact that some water-bearing fractures in crystalline bedrock may become clogged with bedrock cuttings during DTH-drilling. It has also been addressed that due to price fixing with drilling depth DTH-wells have in places been drilled deeper than necessary (Rönkä 1983, 1993, Gierup 1994).

On the other hand, the DTH method may enable better surface water protection because of the deeper casing and safer water supply from (deep) low-yield wells due to greater borehole water storage. There are some findings, too, that clogged water-bearing fractures reopen gradually with time, which may increase the initial well yields. Geologic or topographic features or water demand do not explain the increased well depths in Central Finland.

The depth to the first (main) water strike in a well correlates negatively with the well yield and hydraulic parameters. The negative correlations remain statistically significant even when controlling for well depth and construction year and they stand for both in private and test wells. These findings support the common view that well productivity and hydraulic conductivity gradually decrease downwards in crystalline bedrock, at least to the drilled well depths. This productivity decrease with depth, which obviously is related to fracture properties, is the ultimate reason for the low yield and hydraulic values of deep wells.

It may thus be concluded that the increased well depths and reduced (drill-time) well yields during the last three decades are, for the most part, related to drilling-technical factors, whereas the negative correlation between drilled well depth and well production properties is mainly associated to georelated factors.

In Central Finland, the static bedrock groundwater level usually is near the ground surface and closely follows the bedrock morphology reflecting the overall low transmissivity of crystalline rocks. The static groundwater level does not extend below the water level of Lake Päijänne, which is the lowest surface water body in the study area. Most of the artesian water wells are located in valleys while none of them are situated on hilltops. These findings support the view of Tóth (1963,1995) and others (e.g. Uhl & Sharma 1978, Freeze & Cherry 1979, Fetter 1994, Heath 2004), who state that in a bedrock groundwater basin with adequate local surface topography and slope, three distinctly different groundwater flow regimes, namely the recharge, midline and discharge regions, occur with recharge on hilltops and discharge in valley floors. Moreover, three types of flow system are recognized: local, intermediate and regional.

When regarding the geological factors the soil type at well site did not prove to be of any great importance to the drilled well yield and hydraulic properties in this study. In fine-grained well sites the well production properties appeared to be at their best, but in Central Finland these well sites are more often situated in flatlands and valleys near lineaments and away from hilltops. The well production properties were positively correlated with the thickness of overburden. Moreover, in multivariate analyses the overburden thickness was positively related to well yield and hydraulic values also in non-lineament wells. The connection between overburden thickness and well yield was not, however, as clear as reported elsewhere, e.g. in Sweden. The thickness of overburden neither explained the high yield and hydraulic values in valley wells.

Taylor and Howard (2000) have concluded that, because fractured crystalline rock has a low storativity, the presence of a hydraulic connection to the porous overburden is critical to the long-term productivity of boreholes drawing water from bedrock fractures. This is supported by test well experiences from Central Finland. In most long-term pumping tests, where monitoring wells have been installed in overburden soil layers, the groundwater table in the overburden wells has more or less followed that of the abstraction well (drawdown, recovery). This indicates that part of the sustained yield of the abstraction wells comes from the groundwater storage in the overburden. Also Davis and DeWiest (1966), Daniel (1990), Sander (2007) and Holland and Witthüser (2011) state that best targets for sustainable well sites are located where fractured zones can interact with storage in overlying unconsolidated material.

According to Gustafson (1988), the bedrock wells that have the highest long-term yields very often are situated in a fracture zone that goes into a lake basin and thus the ground water is supplied by induced recharge. This could not be confirmed in this study. Rather the findings from Central Finland suggest that there are no significant hydraulic connections between surface water courses and drilled wells in crystalline rock areas. This is in agreement with the results of Rosenberry and Winter (1993), Mabee (1999) and Oxtobee and Novakowski (2002).

The different rock types in Central Finland possess astoundingly similar median well yields and hydraulic values, while at the same time the variation of the well production properties in each group is several orders of magnitude. This variation has been confirmed in other basement areas, too (e.g. Havlík & Krásný 1998, Krásný 2002, Caine & Tomusiak 2003, Singhal 2008, Banks et al. 2010, Boutt et al. 2010). It thus appears to be possible to drill both very good and very poor boreholes in any crystalline rock type. Lithological differences between crystalline rock types in Central Finland may be considered insignificant from the well production point of view.

In Norway, Rohr-Torp (1994a, 1995, 2000) and Morland (1997) have suggested a positive correlation between drilled well yield and postglacial land uplift rate. In this study, however, the uplift rate did not indicate any clear trend in well yields and hydraulic properties. Also in Henriksen´s (2003a) study, where the yield of crystalline-rock wells (n~13600) was arranged in three profiles parallel to gradients of postglacial rebound in Sweden and Norway, the uplift rate had a low explanation power and a considerable amount of variation in well yields was left unexplained. According to Henriksen, there were two likely explanations for this: the uplift process is more complex than envisaged or other factors contribute to borehole flow rates.

Later Henriksen (2006b) has proposed that it is not the uplift rate but linear belts separating regions undergoing differential uplift (possibly with abrupt changes in uplift rate) that might explain the variation in bedrock hydraulic properties. This could not be confirmed in this study. Recently Henriksen (2008) has argued that more thorough understanding of the connection between uplift and hydraulic conductivity needs to be associated with the geodynamics of the late glacial and Holocene deglaciation. This subject will be discussed in the following Chapter 8.3.

The topographic setting of a drilled well is important for well production properties in Central Finland. Valley wells are most productive whereas hilltop wells, though deepest, yield the least amounts of water. This is concordant with most studies in crystalline rock areas. Also the distance to the nearest (bedrock) hilltop and the relative height difference between the well site and the lowest location in a well´s catchment area are important factors influencing the well yield and hydraulic values in Central Finland. The latter had a significant affect to well production properties in the non-lineament well group, too. This apparently shows that low altitude areas between ridges and hilltops often represent areas of fractured bedrock although any distinct fault zones are not to be recognized.

The proximity of lineaments may be considered the most important factor entity controlling the productivity of drilled wells in Central Finland. This is largely the situation in other crystalline rock areas, too. The most significant single lineament factors appeared to be the prominence of the nearest lineament, the distance to the nearest lineament, and the orientation of the nearest lineament.

The total lineament length and the lineament length density had only a minor influence on the well yield and hydraulic values. But what came as a greater surprise was that the distance to the nearest lineament intersection and the number of lineament intersections had no influence at all on the well production properties. These findings are contradictory to most other studies in crystalline rock areas where lineament intersections have been considered to be of great importance to well yield and hydraulic properties (e.g. Lattman & Parizek 1964, Woodruff et al. 1974, Astier & Paterson 1989, Sander 1996, Dash 2003, Kenny et al. 2006).

In Central Finland, short to medium length lineaments (≤ 10 km) seem to be hydraulically widest. Moreover, the highest yields and hydraulic values are met in wells drilled on short to medium length lineaments, whereas several wells located in major fault zones (> 10 km) produce only small quantities of water. Hence, this study casts doubt upon the general applicability of a rule that the most pronounced fracture zones are those that will yield the most groundwater. The results of this study are supported by the findings from Norway, where many drilled boreholes in topographically prominent fracture zones yield spectacularly little water (Banks et al. 1994, Banks & Robins 2002). Moreover, the largest fracture-zones crossed by tunnels often give rise to very few water leakage problems. Instead, the majority of large water leakages tend to arise from smaller fracture zones or individual fractures/groups of fractures in relatively massive bedrock, often near a major fault zone (Banks & Rohr-Torp 1990, Olofsson 1991, Banks et al. 1992a, 1992b, 1994, Nilsen & Palmström 2001, Banks & Robins 2002, Mabee et al. 2002).

The reasons for low well yields in lineaments have been suggested to be rock gouge, breccia, mineral precipitates or clay weathering products, which may have decreased fracture porosity and hydraulic conductivity specifically in fault core zones. With regard to the most prominent fracture zones, the phenomenon of decreased productivity is ascribed, in many cases, to tightening due to secondary clays, which result from weathering or hydrothermal activity (Cesano et al. 2000, Nilsen & Palmström 2000, 2001). For instance, smectite and kaolinite fracture fillings are known from many Norwegian hard-rock lithologies and areas and they are expected to be rather efficient at tightening fractures (Olesen et al. 2006). Because of the fracture clogging by clay products, water well drillers in Norway have avoided regional fracture zones (Olesen et al. 2006). This has also been the case in the Northern Territory of Australia, where drilled wells have been sited with caution to assure that they are not in the main faults because of clayey materials blocking the boreholes (Verma 2003).

To find out the possible reason behind the low productivity of some test wells drilled in major lineaments in Central Finland, the author placed an order for a diamond core drilling project in December of 2005. The drilling project was conducted by the Geological Survey of Finland next to the Jurvansalo test well no. 931098 in the municipality of Viitasaari. The well is situated near the center of a NW-SE orientated major lineament. The yield of the well is 170 Lhr^{-1}; hydraulic fracturing did not managed to improve the yield. The core sampling and analysis indicated that it is not the clayey material but mainly iron oxides, which have sealed the few fractures open in the near past (Fig. 112, Kesola 2005).

Fig. 112. Core samples from a diamond drilling site near the Jurvansalo test well no. 931098 in the municipality of Viitasaari, Central Finland. The rock type is granodiorite. Note the reddish colour due to apparently hydrothermal Fe-oxyhydroxides (limonite, hematite). The diameter of the sample core is 38 mm. Sample depth 68,59 m below ground surface is marked in the photo. Photo adopted from Kesola (2005); original photo Reino Kesola (GTK).

Iron oxides are the most abundant group of low-temperature fracture fillings also in the Eye-Dashwa granite pluton in Canada and include secular hematite, limonite and goethite (Kamineni & Stone 1983). Low-temperature fracture fillings such as Fe-oxyhydroxides and zeolites occur also in the most recent cross-cutting microfractures in the Molberget postglacial fault in the Lansjärv area, northern Sweden (Eliasson et al. 1991).

The azimuth of the nearest lineament turned out to be of high importance to well production properties in this study. In Central Finland, the most productive lineaments are orientated in the NE-SW and NW-SE directions, whereas the least productive lineaments are found between these directions. The medium-length and long lineaments are most productive when orientated in the NE-SW direction, while the short lineaments are most productive when trending NW-SE. This subject is discussed more thoroughly in Chapter 8.3.

The common existence of high-yield wells (58%) in the non-lineament well group suggests that many water-bearing fracture zones may go undetected in the lineament mapping. While a major part of these fracture zones probably are buried and obscured vertical or subvertical (narrow) faults, some of the fracture zones might well be gently dipping structures or more or less horizontal zones of weakness or fractures (e.g. sheeting; Chilton & Smith-Carington 1984, Houston & Lewis 1988, Greenbaum 1990c, Ahlbom & Smellie 1991, Howard et al. 1992, Greenbaum et al. 1993, Gustafsson 1994, Mabee at al. 1994, Wladis 1995, Kellgren & Sander 1997, Mabee & Hardcastle 1997, Sander et al. 1997, Taylor & Howard 2000, Tam et al. 2004, Adepelumi et al. 2006, Juhlin & Stephens 2006).

Multivariate analyses reaffirmed the concurrent relationships between certain well factors and well production properties. Especially the variables OVER, TOPO, RELAL, LPRO, LDIS and HLDIS stood out as important well factors related to the high well yields and hydraulic values. In addition, the catchment factors BED and WAT, which did not show any great importance in bivariate analyses, were selected into different multivariate models as significant variables. In other respects, however, the results of the multivariate analyses were rather modest and somewhat disappointing. Moreover, because of the missing well data, the number of observations went low, which per se is a quite usual drawback in multivariate data analysis.

Thus it may be summarized that there are no simple relationships between various well factors and the success of boreholes in crystalline basement areas. Despite the similarities in some factors that influence borehole productivity on a regional scale (e.g. lineaments, topography), the heterogeneity of the bedrock aquifers suggests an overriding influence of more local features such as small-scale structures (e.g. Holland & Witthüser 2009). Local factors presumably are difficult or even impossible to be specified properly on topographic maps or by means of remote sensing, especially when defined on two-dimensional analyses of an initially three-dimensional geologic architecture (e.g. Lewis 1990, Brown 1994, Diop & Tijani 2008).

8.3 Role of seismotectonics

Tectonic activity in the form of earthquakes is responsible both for creating fractures and for reactivating them. Crustal faults may be activated when they are favorably oriented and critically stressed with respect to the contemporary tectonic stress field. During the tectonic evolution, reactivation in core and damage zones keeps faults permeable. On the other hand, crystallization and sealing of fractures make faults impermeable with the sealed discontinuities constituting zones of weakness (Chapter 4.2.3; Brown & Bruhn 1996, Cox et al. 2001, Ito & Hayashi 2003, Constantin et al. 2004, Sibson 2004). Therefore, in any hard rock environment, geologically recently activated fractures are the most important for groundwater flow (e.g. Banks et al. 1993, Fernandes & Rudolph 2001).

In Central Finland, the N40°E and N50°W lineaments possess highest median well yields and hydraulic values (Fig. 93a). Consequently, they may be assumed to represent the most recently activated faults. The rare spatiotemporal occurrence and especially the low magnitudes of present-day and historical earthquakes suggest that the modern-day earthquakes are not responsible for the reactivation of the N40°E and N50°W faults in Central Finland. It should be recalled that these faults obviously are moderately to steeply dipping kilometer scale features, whose reactivation has and will require a great amount of energy (e.g. Juhlin et al. 2010). Yet in Finland only three earthquakes per annum at most have magnitudes greater than 2,0 and the magnitude of all earthquakes in human records is less than five (Chapter 4.2.2; Ahjos & Uski 1992, Ojala et al. 2006, http://www.helsinki.fi/geo/seismo/maanjaristykset/suomi.html). Ahjos et al. (1984) and La Pointe and Hermanson (2002) have estimated that the present-day maximum magnitude earthquake for Finland is M_L 5,0…5,5 (see also Mäntyniemi et al. 1993). Moreover, based on the figures given by Bödvarsson et al. (2006) from Sweden, one should expect this magnitude earthquake in Finland perhaps every one thousand years. Thus, it seems that in Central Finland there must have been paleoearthquakes of much larger spatiotemporal amplitudes and frequencies than is the case today.

The N40°E faults in Central Finland are orientated parallel to the general direction of the large postglacial thrust (reverse) faults in the Lapland Fault Province, northern Fennoscandia (Fig. 113, Chapter 4.3). The orientation of the paleostress field implied by these postglacial faults is subparallel to the contemporary tectonic stress caused by spreading of the North Atlantic Ridge ($S_H \approx$ NW-SE; Richardson et al. 1979, Clauss et al. 1989, Stephansson 1989, Müller et al. 1992, Lindholm et al. 1995, Reinecker et al. 2005). There are also a couple of NW-SE trending postglacial faults in the LFP. The postglacial faults have been activated in connection with severe earthquakes (even M 8) during the last phases of the Weichselian deglaciation some 10 ka ago (Chapter 4.3).

Postglacial faults have been related to favorable groundwater flow conditions. For example, large quantities of spring water leaking from scarps of postglacial faults have been documented from northern Fennoscandia (Lagerbäck & Henkel 1977, Lagerbäck 1979, 1988, 1989, Lagerbäck & Witschard 1983, Muir Wood 1989, 1993a, Dehls et al. 2000b, Lagerbäck & Sundh 2008). According to Lagerbäck and Witschard (1983), this is a typical feature of most of the PG faults and indicates an extensive, open fracturing of the bedrock (see also Lundqvist & Lagerbäck 1976, Adams 1993, Anon 1999). Exceptionally large amounts of groundwater poured out of the Stuoragurra escarpment some time after the January 1996 M_L 3,9 earthquake, which was most likely located along the fault itself at a depth of ca. 10 km (Fig. 114; Olesen et al. 1999, 2000a, 2010, 2011b).

Moreover, geophysical investigations have revealed low-resistivity and low-seismic zones within the postglacial faults, especially in the hanging-wall blocks. This has been interpreted to be caused by a generally high incidence of fracturing and faulting of the bedrock (Lagerbäck & Henkel 1977, Paananen 1987, 1989, Olesen et al. 1991, 1992a, Kuivamäki et al. 1998, Anda et al. 2002). Indeed, exposed bedrock outcrops and drilling results show that the bedrock is strongly fractured and shattered (Lagerbäck & Witschard 1983, Kuivamäki & Vuorela 1985, Talbot 1986, Kuivamäki et al. 1988, 1998, Lagerbäck 1989, 1992) and has a relatively high hydraulic conductivity ($10^{-5}\ldots10^{-7}$ ms^{-1}; Vuorela et al. 1987, Kuivamäki et al. 1988, 1998, Larsson 1989, Paananen 1989). It may be mentioned that drilling trough the Stuoragurra fault revealed a groundwater yield among the highest ever recorded in Norway. The yield was estimated to greatly exceed 17,000 Lhr^{-1}, which was the capacity of the most powerful pump available for the 5,5" diameter borehole (Klemetsrud & Hilmo 1999).

Because of the geologically short time period since the last glaciation, Ahlbom and Smellie (1991) have proposed that no significant fracture infilling process, which could otherwise have sealed the fracture system, has occurred in postglacial faults.

There are also other indications of groundwater associated with postglacial faulting. According to Muir Wood (1993a), the liquefaction of the till deposits in connection with postglacial landslides may not have solely been due to ground shaking but could reflect the expulsion of groundwater as a result of strain changes accompanying the fault movements. In the Lansjärv area, northern Sweden, Lagerbäck and Sundh (2008) have found sand dykes interpreted to have been formed by expulsion of groundwater from bedrock in connection with postglacial faulting.

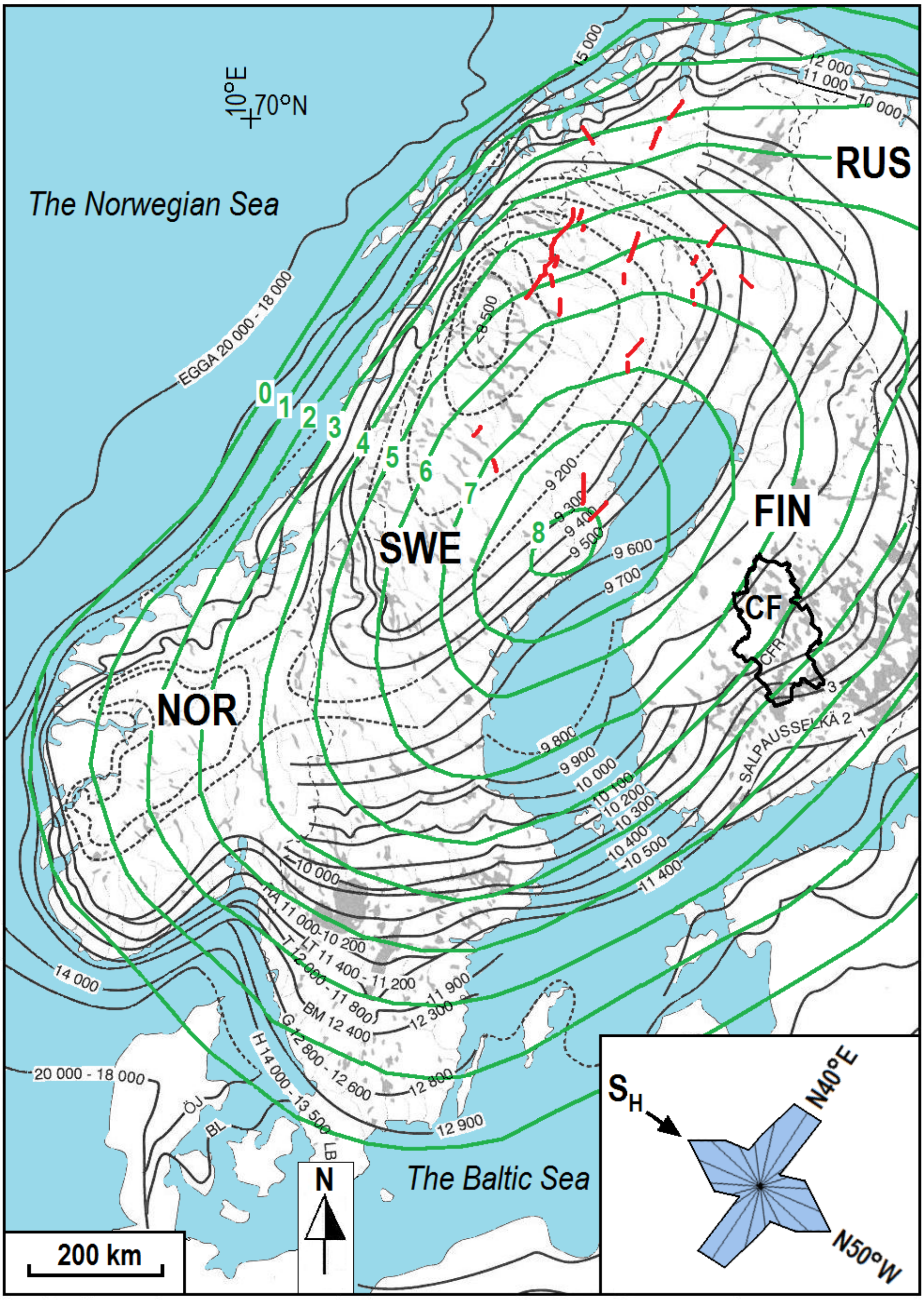


Fig. 113. Location of postglacial faults in northern Fennoscandia (red lines; Kuivamäki et al. 1998, Lagerbäck & Sundh 2008), successive ice-marginal lines for the Weichselian deglaciation between ca. 20,000 and 8,500 years BP (black lines; Sollid et al. 1973, Kujansuu 1992, Lindström et al. 2000) and isolines of the apparent land uplift rate in mm a^{-1} (green lines; Ågren & Svensson 2007). The median yield of drilled wells vs. lineament azimuth and the orientation of the maximum horizontal stress S_H in Central Finland (CF) are presented in the lower right inset. NOR=Norway, SWE=Sweden, FIN=Finland, RUS=Russia. Modified from Lindström et al. (2000).

Fig. 114. Oblique aerial photograph of the Stuoragurra fault as it crosses Finnmarksvidda, northernmost mainland Norway. Groundwater is pouring out of the escarpment (scarp height ca. 7 m) to the right in August 1996. Photo taken from the west. Photo adopted from Olesen et al. (2000a).

To date, convincing examples of postglacial faulting have not been described in southern and central Fennoscandia (Chapter 4.3; Lagerbäck & Sundh 2008). However, as Kukkonen et al. (2010a) state, bedrock deformation and seismic activity may have appeared on a much larger scale than that signaled by the sharp and easily identified PG faults in northern Fennoscandia. According to Lagerbäck and Witschard (1983), it is likely that a large number of lesser features of postglacial faulting may have escaped detection (see also Kuivamäki & Vuorela 1985). Cloos (2009) argues that very few earthquake-generating ruptures propagate to the surface. According to him, it is uncommon even for events between M 6...7. Moreover, low, short or discontinuous scarps, resulting from small/moderate (M~6) earthquakes, can easily be overlooked, particularly if they follow old Precambrian structures (Björck & Svensson 1992). According to Adams (1981, 2005), there are from 10 to 100 times more postglacial faults than have been found by date and only 1/7th of the expected number of earthquake events between magnitudes 6 and 7 might have been found from neotectonic mapping. It must also be admitted that even relatively small events may be capable to modify hydrogeological conditions and increase bedrock permeability manifold (Chapter 4.2.4; Muir-Wood & King 1993). Then, from the groundwater production point of view, there is no need for fault reactivation to be morphologically and/or structurally discernible in the field.

Additional evidence of palaeoseismic activity at the end of the Weichselian deglaciation comes from soft-sediment studies carried out in different parts of Fennoscandia and adjacent areas (Chapter 4.3; Feyling-Hanssen 1966, Lagerlund 1977, Björkman & Trägårdh 1982, Lagerbäck 1988, 1990, 1991, 1992, 1994, Anundsen 1989, Bäckblom & Stanfors 1989, Mörner 1989, 1991, 1996, 1997, 2001, 2004, 2005, 2009, Hansbo 1993, Mörner & Tröften 1993, Nisca 1995, Tröften 1997, 2000, Tröften & Mörner 1997, Mörner et al. 2000, Forwick & Vorren 2002, Kotilainen & Hutri 2004, Olesen et al. 2004, Hutri 2007, Hutri & Kotilainen 2007, Hutri et al. 2007, Virtasalo et al. 2007, 2010, Lagerbäck & Sundh 2008, Bungum et al. 2010, Hoffmann & Reicherter 2011).

In recent years several authors have examined with 2D and 3D finite element ice-earth models, how the waxing and waning of the Weichselian ice sheet during the last glacial cycle have affected the state of stress in the Earth, and how those changes in stress have influenced the stability of faults (Chapter 4.5; Lund & Zoback 2003, 2007, Hetzel & Hampel 2005, 2006, Lund 2005a, 2006a, Hampel & Hetzel 2006, Hampel et al. 2007, 2009, 2010a, 2010b, Turpeinen et al. 2008, Karow 2009, Lund et al. 2009, 2011, Lund & Schmidt 2011). The results of these modeling studies generally show that the rate of faulting decreased during the presence of the ice sheet and strongly increased during deglaciation. Importantly, causal relationships between the increased seismicity and deglaciation also apply to the southern and south-central parts of Fennoscandia. Moreover, Turpeinen et al. (2008) state that the area affected by deglaciation-induced seismicity appears to have migrated northward with the margin of the melting Fennoscandian ice sheet.

Summarizing the aforementioned, it seems quite reasonable to suggest that, during the Weichselian deglaciation, favorably oriented fault zones were regionally activated not only in northern Fennoscandia but in central Finland, too. This analogy with the postglacial faults in the Lapland Fault Province may well explain the permeability of the N40°E and N50°W faults in Central Finland.

The following proposal for fault reactivation in Central Finland is based on the relations between drilled well yields and tectonic lineaments presented in this study and on the results of the modeling studies listed above. The reader is referred to the original model studies for detailed information on the modeling, which is beyond the scope of the present study. The general spatiotemporal evolution of principal stresses and their implications for fault stability in different phases of the glacial loading-unloading cycle in Central Finland are schematically illustrated in Fig. 115. The corresponding changes of the stress state on a NE-SW trending thrust fault are exemplified with Mohr diagrams (Chapter 4.1.3).

Prior to the glacial loading, the brittle crust is in a frictional-failure equilibrium on optimally oriented pre-existing NE-SW faults in the plate tectonic thrust regime of Central Finland (Fig. 115:A). The direction of the maximum horizontal stress is NW-SE ($S_H = \sigma_1$). The Mohr circle lies close to the line of the Mohr-Coulomb criterion sometimes touching it due to the accumulation of the horizontal tectonic stress along locking faults. As a consequence of this, the critically stressed NE-SW faults may periodically slip in a reverse movement causing earthquakes. However, earthquakes in this phase are a rare phenomenon and mostly of low magnitudes (<< 2 M), which essentially resembles the contemporary situation in Central Finland.

Then, glacial loading gradually increases the mass on top of the crust, which increases the vertical stress σ_3, but also leads to flexure, which increases the horizontal stress σ_1 in the upper crust (Fig. 115:B; Watts 2001, Hetzel & Hampel 2005, Pascal et al. 2010b, 2010c). According to Lund et al. (2009) and Lund and Schmidt (2011), glacially induced stresses have a significant effect on the ambient tectonic stress field especially in the surficial part of the crust. The changes in the vertical stress with time reflect simply the changes in ice sheet thickness. Also glacially induced horizontal stresses generally follow the build-up of the ice.

When the glaciation proceeds, the stress increase during loading is more pronounced for the vertical stress σ_3 and the differential stress ($\sigma_1 - \sigma_3$) decreases. Thus, in compressive thrust fault regimes, for any angle between σ_1 and the normal to the fault plane, this will cause shear stress (τ_s) to decrease and normal stress (σ_n) to increase promoting fault stability (Johnston 1989b). This means that the Mohr circle shifts to the right away from the criterion line and its diameter becomes smaller with the result that faulting is suppressed for most of the lifetime of the ice sheet (Fig. 115:B). During this stabilization, horizontal tectonic stresses can gradually increase and accumulate beneath the ice sheet (e.g. Johnston 1987) until the glacial unloading begins. According to Lund et al. (2009) and Lund and

Schmidt (2011), the maximum shear stresses, i.e. the highest failure potential, at the end of the glaciation are concentrated in northern Fennoscandia, in general agreement with the region of large postglacial faults. Also Central Finland belongs to the area with elevated shear stresses.

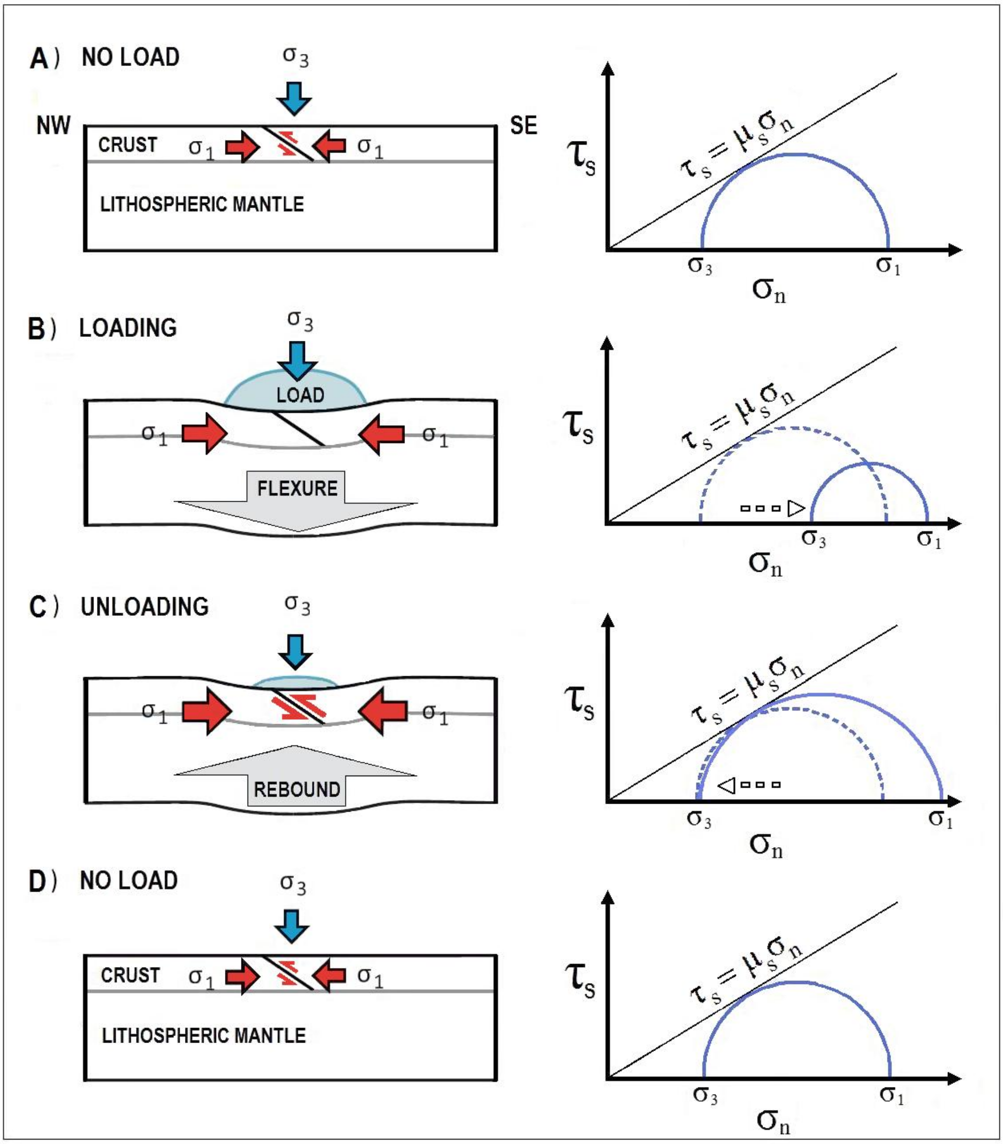


Fig. 115. Schematic sketch of the principal stresses (max σ_1, min σ_3) in the lithosphere A) prior to glacial loading, B) during loading, C) during unloading, and D) after completion of the loading-unloading cycle for a pre-existing cohesionless and critically stressed thrust fault trending NE-SW in the thrust stress regime of Central Finland. The maximum principal stress σ_1 is horizontal (S_H) and orientated in a NW-SE direction; the minimum principal stress σ_3 is vertical (S_V). For each phase A-D, the Mohr circle is plotted, in reference to the stress state prior to loading (dashed circle) and under load (σ_n is the normal stress, τ_s is the shear stress and μ_s is the coefficient of friction). Note that the size of the arrows marking the principal stresses represents the relative stress change. Modified from Karow (2009).

Postglacial unloading reduces σ_3 but σ_1 also begins to decrease owing to the unloading-induced rebound. Because the isostatic rebound of the depressed elastic lithosphere is a much slower process than the retreat of the ice sheet and corresponding removal of the ice load, the vertical stress can decrease earlier and faster than the accumulated horizontal stress with the result that the differential stress increases and may become even greater than in the initial state. In other words, the Mohr circle strongly shifts to the left while its diameter gets larger (Fig. 115:C). Hence, it touches the criterion line with the consequence of abrupt release of accumulated strain energy and reverse slips producing earthquakes on the favorably orientated NE-SW fault planes. According to Lund et al. (2009), the faults generally become unstable before deglaciation is complete, although the maximum measures of instability are not reached until the ice disappears.

As a consequence of this thrusting on the NE-SW fault planes, the orthogonal NW-SE orientated (short) faults, which run more or less parallel to the S_H, slip in a strike-slip mode. That is, about the way tear faults accommodate the abrupt changes in the amount of displacement along thrust faults (Hatcher 1990) or transform faults slip laterally near transpressive plate boundaries (Stein & Klosko 2002). Also Fenton (1993) suggests that in addition to postglacial thrusting there should be strike-slip fault movement in Fennoscandia orientated approximately parallel to the direction of S_H. In Scotland, about 50% of the PG faults are orientated NW and display cumulative strike-slip movement of up to ca. 160 m (Fenton 1993; see also Davenport et al. 1989, Ringrose 1989b, Stewart et al. 2000).

After the completion of deglaciation, viscoelastic relaxation slowly brings the stresses back toward the initial state (Fig. 115:D). According to Lambeck and Purcell (2003) and Lambeck (2005), the present-day observed stress state is largely free from residual glacial-cycle deviatoric stresses.

In summary, the response of favorably orientated faults is characterized by cessation of slip during glacial loading followed by pronounced slip acceleration during postglacial unloading before a return to a lower level of seismicity and slip rate. Importantly, it is the temporal increase in the differential stress, which causes high slip rate and, eventually, abrupt fault reactivation (Karow 2009, Hampel et al. 2010a, 2010b).

According to the modeling results of Lund et al. (2009) and Lund and Schmidt (2011), before and during the initial phase of glaciation the glacially induced horizontal minimum stress S_h (oriented in the NE-SW direction) may decrease and even become negative, i.e. tensional. This is due to the positive flexure of the lithosphere in the peripheral bulge area, which is located at some distance to the southeast of the advancing ice margin. Although the simulations made by Lund et al. (2009) indicate that glacially induced stresses have not been able to overcome the effect of the tectonic background stress at depth ($\geq$ 500 m), they still may have dominated the superficial part of the bedrock (Lund & Schmidt 2011, Valli et al. 2011). Had this been the case in Central Finland, the NW-SE trending vertical short (and obviously shallow) faults may have reacted in a strike-slip mode with the result of increased permeability (Fig. 100, Table 64).

Most of the large postglacial reverse faults in northern Fennoscandia run across the withdrawing ice margins and cross the contours of postglacial uplift-rate (Fig. 113; Kujansuu 1992 and his Fig. 10, Kleman et al. 1997 and their Fig. 8, Kuivamäki et al. 1998 and their Fig. 42, Johansson 2007 and his Fig. 3, Bungum et al. 2010 and their Fig. 2a). Hence, they do not seem to coincide with the retreat pattern of the Weichselian ice sheet. Instead, it is evident that the thrust regime due to plate tectonics has exclusively determined their optimal orientation in accordance with the maximum horizontal stress S_H. The modeling results of Lund et al. (2009) and Lund and Schmidt (2011) reassert that at larger depths in the earth, where the background stress field is significantly larger than the glacially induced field, the orientation of optimally oriented faults is determined solely by the tectonic background stress.

Similarly, in the deglacial stress field of Central Finland, the medium length to long (and obviously deep) NE-SW lineaments were favorably oriented, and still are, with respect to S_H suggesting reactivation via tectonic thrusting. This is concordant with their generally high permeability. In contrary, the long (and obviously deep) NW-SE lineaments were not, and still are not, optimally oriented in the tectonic thrust regime of Central Finland. Hence, they do not easily experience any reactivation. This is in agreement with their low to moderate permeability (Fig. 98, Table 64).

A recent assessment of the deglacial earthquakes in northern Fennoscandia suggests that they represent the stress-strain release of some 50,000 years accumulation at the level of seismic activity typical of stable cratons (Adams 2006). According to Olesen (2010), the lithospheric strength of the crust in the northern Fennoscandia was sufficiently high to accommodate this stress accumulation during the last glaciation. According to Muir Wood (1993a), the time of strain accumulation under ice sheet was about four times shorter in southern Sweden compared to that of northern Sweden. Hence, in southern Sweden, the accumulated strain could be released by a series of earthquakes no larger than magnitude 5 (Muir Wood 1993a). Moreover, recent studies indicate that the permanent ice cover in southern and central Finland prior to the LGM seems to have lasted a significantly shorter time span than earlier has been thought (Ukkonen et al. 1999, Svendsen et al. 2004, Lokrantz & Sohlenius 2006, Lunkka 2007, Nenonen 2007). Hence, despite their optimal orientation in the thrust regime, the longest NE-SW lineaments (>> 10 km) in Central Finland might not have experienced reactivation during the deglacial time due to an inadequate stress-strain accumulation, though their counterparts in northern Fennoscandia may well be several tens of kilometers in length and have over 10 meters high fault scarps.

Lund (2006b) suggests that the shape of the depression caused by the ice sheet, and therefore, the magnitude and distribution of induced stresses, depend critically on the shape (e.g. spatial extent and slope) and thickness of the ice. According to Turpeinen et al. (2008), the amount of postglacial slip strongly depends on the ice sheet thickness, the shortening rate of the crust due to tectonic stress and the viscosity of the lower crust and the lithospheric mantle, whereas the width of the ice sheet, the rate of deglaciation and the fault dip have a minor effect (see also Karow 2009).

Abrupt changes in rates of faulting may occur with a time lag of several thousand years after a glacial load has been applied or removed. Subsequently, the accelerated or decelerated motion of faults may be sustained over periods of several thousand years (Hetzel & Hampel 2005). Moreover, in contrast to the assumptions made by Johnston (1987) that large continental ice sheets in general suppress seismic activities, the model results of Karow (2009) indicate that the seismicity is mainly affected by a change in the load, not the load itself. Hence, a static load should lead to the readjustment of the steady-state stress field over very long time spans and, consequently, to a seismicity similar to that of the preloading state (Karow 2009). It should also be noted that fault reactivation may well occur at any rapid increase or decrease in ice volume and/or areal extent during the glacial cycle, not just at the actual end of a glacial cycle (Lund 2006b, Lund & Näslund 2009).

There are also other findings of faulting and well yields in Fennoscandia, which might be implications of postglacial fault reactivation suggested above. Niini (1967, 1968, 1973) found that in the bedrock valleys of southern Finland the NE-SW direction, in addition to valley intersections, represents the most fractured bedrock (see also Anttila 1988). According to Niini (1968), there have been strong changes, geologically perhaps quite recently, in the fracture zones of southern Finland. In the municipality of Leppävirta, east-central Finland, Leveinen (2001) and Leveinen and Vallius (2005) found that the main bedrock aquifer at the local water work comprises two N35-40ºE striking fracture zones with drilled well yields between 14,580-30,000 Lhr^{-1} (350-720 m^3d^{-1}). A major NW-SE striking fault zone crossing these main aquifers is poorly conductive showing no response to pumping.

Larsson (1959, 1963) has stated that tensile fractures parallel to the NNE-SSW dikes in the Precambrian basement areas of south-central Sweden would be more transmissive and open than shear fractures, because the latter are held closed by a component of normal stress. However, because of some weak arguments in Larsson's presentation (Chapter 4.6), some authors (e.g. Rohr-Torp 1987) have argued that later reactivations of fracture zones are needed. Indeed, an alternative explanation to Larsson's findings might be the reactivation of the NNE-SSW fracture zones via postglacial thrust (reverse) faulting. Similarly, instead of considering the relationships between the stress orientations originating from the late Precambrian folding phase and the permeability of the fracture zones, Rohr-Torp's (1987, 2003) findings in Norway (Chapter 4.6) might be explained from the point of view of postglacial stress field and fault reactivation.

According to Wu (1996, 1998, 2000), the orientations of postglacial thrust faults in southeastern Canada (Adams 1981) indicate that the maximum horizontal principal paleostress (S_H) orientation was mainly in the NW-SE direction, i.e. perpendicular to the ice margin at glacial maximum and consistent with the direction of ice retreat (e.g. Adams 1989b). The present S_H orientation (NE/ENE-SW/WSW) is almost perpendicular to this paleostress orientation (Sbar & Sykes 1973, Herrmann 1979, Horner et al. 1979, Richardson et al. 1979, Hasegawa & Adams 1981, Yang & Aggarwal 1981, Adams 1987, 1991, 1995, Wahlström 1987, Adams et al. 1988, 1989, Adams & Bell 1991, Richardson 1992, Zoback 1992a, 1992b).

Wu (1996, 1998, 2000) suggests that the rebound stresses gradually diminished until, at present, the plate tectonic stresses caused by spreading of the Mid-Atlantic Ridge appear to be the dominant forces influencing the orientation of S_H. So, according to Wu, a large stress rotation must have occurred in southeastern Canada during postglacial time. However, as stated earlier (Chapter 4.3), a number of postglacial features compiled by Adams (1981) were later understood to have resulted from other, non-tectonic, mechanisms (e.g. Adams 1989a, Fenton 1994a). Therefore Fenton (1994a) suggests that the contemporary stress field has been acting in eastern Canada and northeastern United States since the Cretaceous, that is, about the time that the North Atlantic plate margins attained their present configuration. This is considered to be the case in Fennoscandia, too (Muir-Wood 2000, Torsvik & Cocks 2005).

Several studies from northeastern USA (Beebe & Rauch 1979, Helvey & Rauch 1993, Sidle & Lee 1995, Mabee et al. 2002, Moore et al. 2002a, Robinson 2002, Folan et al. 2004; Chapter 7.3.4) show that there are permeable fracture zones lying perpendicular to the current (and obviously deglacial) NE/ENE-SW/WSW oriented maximum horizontal stress direction in a thrust/strike-slip fault regime of the region (e.g. Zoback 1992a, 1992b). Then it might not be implausible to suggest that these favorably orientated fault zones had experienced a general reactivation during the Laurentide deglaciation in a similar way as in Fennoscandia.

Postglacial faulting may generate unfavorable conditions for the safe behavior of the spent nuclear fuel repositories in bedrock. Because of this, understanding postglacial faulting has a high societal relevance in countries preparing for final disposal of nuclear waste and attempting to predict the future behavior of bedrock during forthcoming glaciations (Kukkonen et al. 2010a). Munier (1993) and Bäckblom and Munier (2002) have proposed that the Fennoscandian bedrock has accumulated fracture zones with a sufficiently wide range of orientations that strains anticipated in the next few million years are likely to reactivate pre-existing zones of weakness rather than generate new fracture zones (see also Kumarapeli 1987, Milnes et al. 1998). Autio (1973) has set forth the same idea for central Finland. Mörner (1989), however, states that the latest glacial and postglacial faults and fractures do not necessarily represent reactivations of old weak zones but also – and, maybe even predominantly – may represent new ruptures due to a new stress and strain field.

The larger regional brittle deformation zones are believed to be where any substantial future movement and seismic activity could take place during the early stages of glacial retreat (Lambeck & Purcell 2003, McEven & Anderson 2009). However, the results of the present study indicate that short to medium length fracture zones may more easily become reactivated in the deglacial stress field of southern and central Fennoscandia. According to Fenton et al. (2006), faults as short as 5 km are capable of producing damaging earthquakes.

According to Lowry et al. (2006), there appears to be no evidence to support or reject a conceptual model that includes the long-term influence of active tectonics on the regional hydrogeology. Hence, their conclusion is that in nuclear disposal studies there needs to be a better quantitative understanding of paleohydrogeology.

Indeed, the findings of the present study should be addressed in the assessment of long-term safety for long-lived radioactive waste in deep bedrock vaults in regions that might be subject to future glaciations. Further work is especially required to identify what the study results are revealing about failure or reactivation of discontinuities in the crust during glacial cycles in southern and central Fennoscandia, where the rock vaults are currently under construction.

8.4 Criteria for siting high-yield drilled wells in crystalline bedrock

This paper shows that several factors can control drilled well productivity in crystalline rock areas. Although there do not appear to be any unique factor that would guarantee finding large amounts of water, such factors can allow, when coupled, obtainment of high-yield well sites with a higher rate of success. Additionally, these factors may also serve to identify geological circumstances in which some bedrock localities are likely to be insignificant in hydrogeological terms (e.g. Black et al. 1986).

At first it must be admitted that although various factors affecting the well production properties evaluates the relative favorability of potential high-yield drilling sites, detailed hydrogeologic and geophysical field investigations generally are needed for the selection of actual drilling sites (e.g. Mäkelä 1990b, 1993, Clarke & McFadden 1991, Teeuw 1995). It would also be unwise to use any single factor for different investigation areas without first questioning whether that factor is able to help in locating high-yield wells in that particular geologic setting (Mabee 1999).

The use of lineament interpretation to aid in locating high-yield bedrock wells is of vital importance. Indeed, well siting in crystalline rock areas should always start with lineament mapping. Most groundwater exploration projects have had higher success rates when sites for detailed geophysical surveys and well drilling have been guided by lineament mapping (e.g. Lattmann & Parizek 1964, Mäkelä 1990b, Teme & Oni 1991, Gustafsson 1993, 1994, Sander et al. 1996, Magowe & Carr 1999, Savané & Biémi 1999, Kellgren & Sander 2000, Kellgren 2002, Brito Neves & Albuquerque 2007). In Central Finland, topographic maps have proved to be very valuable means of investigation in lineament interpretation. In test well projects aerial photographs and various aerogeophysical data sets have additionally been used.

Deducing the hydrogeological importance of different lineament sets needs a dynamic understanding of the regional lineament pattern in terms of the state of stress and associated deformation styles of the rock body (e.g. Cohen 1993, Domoney et al. 2003, Henriksen 2006b). Such an understanding has the potential to significantly reduce the time and costs involved in borehole site location and improve the borehole success rate and median borehole yield. The regional stress state can be ascertained from global data sets published in the WSM Project (Chapter 4.1.4; Heidbach et al. 2008b, Heidbach 2009).

The key point in obtaining the highest probability of success in high-yield well siting is to be able to identify those lineament sets, which are optimally orientated and critically stressed in a present-day stress regime and/or have been geologically recently reactivated. These lineaments are the most important for groundwater flow in crystalline bedrock. It presumably would be an advantage for the lineament-stress analysis, if the tectonic stress had a preferred orientation over a large region. But what is still more important, cycles of glacial loading-unloading in formerly glaciated areas seem to have been of great relevance for lineament reactivation and thus for well yields and hydraulic properties even today.

With the aid of the lineament-stress analysis it is possible to predict, which structural features in different stress regimes would likely yield the highest capacity water wells. In this aim, the following approach is recommended to any area where a reasonable understanding of the local tectonics and stress distribution exists: 1) in a thrust fault regime one should go for detecting lineaments, which strike perpendicular to the maximum horizontal stress direction S_H; 2) in normal fault regimes the most promising lineaments should lie parallel to the S_H, while 3) in strike-slip regimes they either coincide with the S_H or diverge at various angles (<< 45°) to it. Concurrently it is important to remember that steep fault dips in normal and especially in reverse faults under (near) hydrostatic pore pressures would result in very few of them being active today, regardless of the orientation of S_H (Zoback 2011).

In Central Finland, the median well yield and the proportion of high-yield wells are at their highest in the NE-SW (perpendicular to S_H) and NW-SE (parallel to S_H) orientated lineaments. Hence, in well siting these lineaments should be preferred if possible to other lineament sets. However, along with the lineament orientation, the prominence of lineaments must also be considered. Most high-yield wells (86%) are located along short to medium length lineaments. The long (> 10 km) NW-SE lineaments, although being the most prominent lineaments in Central Finland, are not recommended as drilling sites for major water supply. This is because of their low to moderate permeability and a marked number of low-yield wells indicating that these lineaments have not recently been activated. Instead, the short (< 2 km) NW-SE lineaments frequently are highly productive and may well be suggested as potential drilling sites for high-capacity wells. The situation is largely contrary with the NE-SW lineaments, where the medium length to long lineaments seem to be most advantageous for high-yield well drilling. However, the most prominent NE-SW lineaments (>> 10 km) might not suit sites for high-yield wells.

In order to ensure high well yields, drilled wells in Central Finland should be located near the lineament centers. Both the median yield and the proportion of high-yield wells peak at the central parts (≤ 100 m) of the lineaments. Moving away from lineament centers decreases the probability to catch a high-yield well site. This does not mean, however, that high-yield wells are not to be found outside lineaments. To the contrary, there are many high-yield wells even in areas without any distinct lineament pattern. They only are intentionally hard to site in these areas without any such features possibly implicating fractured bedrock.

It has been proposed that siting of wells on the "flanks" of larger fracture zones instead of their central parts would lead to higher success rates (e.g. Sander 1996, 1997, Dewandel et al. 2005). Although there may be some single wells supporting this proposition, the results of this study generally show that in Central Finland there is no need to avoid the lineament centers. Rather the question is, which lineament sets (azimuth, prominence) are most suitable for high-yield well drilling.

From the findings of this study, the benefit of the topographic analysis in high-yield well siting is lesser than that of lineament interpretation. This is concordant, for example, with the results of Yin and Brook (1992a, 1992b), who conclude that lineament mapping is far more successful in locating high-yield wells than the topographic method and should be employed whenever possible. On the other hand, it should be recalled that only 5% of high-yield wells in Central Finland are situated on hilltops (total percentage of wells 10%) whereas 30% of them are located in valley positions (total percentage of wells 19%). Hence, staying in low-elevation areas away from hilltops and ridges may well increase the chances to catch a high-yield well site. Many topographically related well factors support the validity of this view (Table 68).

High-yield wells often draw water from overlying soil and weathered layers where the storage capacity is usually considerably higher compared with unweathered crystalline rock (e.g. Davis & DeWiest 1966, Chilton & Smith-Carington 1984, Daniel 1990, Greenbaum et al. 1993, Kellgren & Sander 1997, Bocanegra & Silva 2007). It has frequently been suggested that boreholes should be sited on areas with high lineament frequency and a thick overburden (e.g. Beeson & Jones 1988, Mambali & Kongola 1990, Carruthers & Smith 1992, Chilton & Foster 1995, Okereke et al. 1998). In this study, the thickness of overburden did not have any major effect on drilled well yields. However, low permeability groundwater reservoirs in thick till deposits and under the flanks of glaciofluvial formations (deltas, sandurs, eskers) might effectively be harnessed by wells drilled into the underlying bedrock. Of course, the most favorable conditions should exist when coarse-grained sediments overlay fracture zones in topographic depressions, i.e. a good hydraulic contact exists between a large unconsolidated groundwater reservoir and a high-yielding fracture zone below it (Olofsson 1994).

The siting of high-yield wells may also more effectively be facilitated, if previous borehole information exists or water well inventories will be executed in the area of interest.

Detailed surface-geophysical surveys can provide valuable data for high-yield well siting. In Central Finland, various geophysical methods have been applied in test well projects from the late 1980's: refraction seismics, ground-penetrating radar, vertical electrical resistivity sounding (VES), multi-electrode resistivity survey (CVES), azimuthal resistivity survey, and very low frequency measurements (VLF, VLF-R) (Mäkelä 1989a, 1990a, 1990b, 1993, Penttinen et al. 2000).

Of these geophysical techniques, the refraction seismics has proved to be the most reliable and successful method in locating high-capacity drilled wells in Central Finland (Fig. 116). The study of Lindsberg (2011), which was conceived under the umbrella of the present study, confirms the preliminary results (Mäkelä 1990b, 1990d) that the bedrock seismic velocity at high-yield well sites in Central Finland lies near 4000 ms^{-1} (Fig. 117). This is in agreement with the results of Breilin et al. (2003) from crystalline rocks in east-central Finland. Clearly lower (< 3000 ms^{-1}) and higher (> 5000 ms^{-1}) seismic velocities indicate low-permeability bedrock conditions due to weathering and sealing of fractures and due to compact rock, respectively (Mäkelä 1990b, 1990d, Lindsberg 2011). These findings together with the lineament-stress analysis have greatly improved the success of test well drillings in Central Finland during the last decade (Table 90).

Fig. 116. Refraction seismic sounding next to the Höytiä test well no. 892015 in the municipality of Uurainen, Central Finland, in the summer of 2004. Photo adopted from Lindsberg (2011). Original photo Ossi Alho/CETECF.

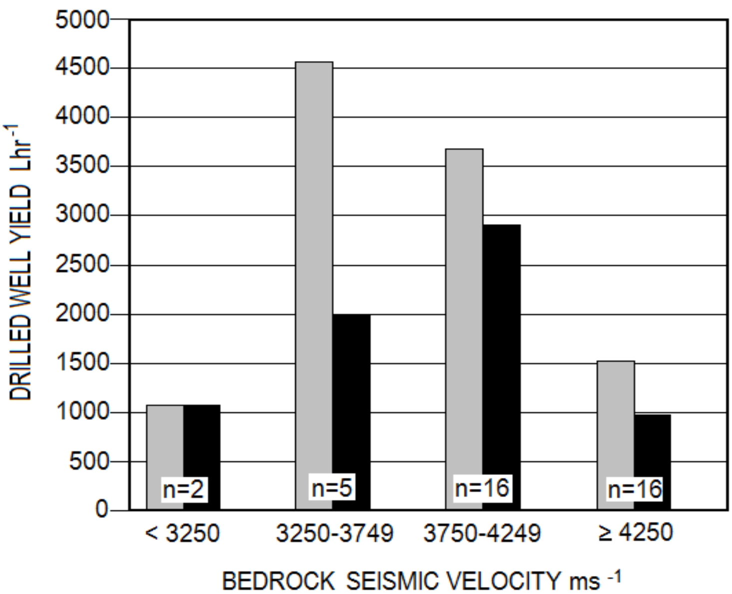


Fig. 117. Bar chart showing the mean (■) and median (■) drilled well yield (Lhr^{-1}) vs. bedrock seismic velocity (ms^{-1}) at valley well sites in intrusive crystalline rocks of Central Finland (n=39). Modified from Lindsberg (2011).

Table 90. Statistics for test well depth, yield and hydraulic parameters grouped according to the construction decade of the test wells in the CF drilled well database (n=73). Highest values are in bold. Comparisons significant at the α≤0.05 level between the groups tested with the nonparametric Kruskal-Wallis test for group means and with the median one-way analysis for group medians are indicated by different letters [A and B].

Variable	Parameter	TEST WELL CONSTRUCTION DECADE		
		1980´s n=18 25%	**1990´s n=32 44%**	**2000´s n=23 31%**
DEPTH m	mean median	59^{B} 56^{B}	**105^{A}** **103^{A}**	104^{A} 91^{A}
Q Lhr^{-1}	mean median	2463^{AB} 1475^{B}	2517^{B} 1375^{B}	**5197^{A}** **4600^{A}**
Q/d_s $Lhr^{-1}m^{-1}$	mean median	62 30^{AB}	36 13^{B}	**64** **46^{A}**
Q/s m^2s^{-1}	mean median	$6{,}97x10^{-5}$ $3{,}48x10^{-5\ AB}$	$4{,}13x10^{-5}$ $1{,}55x10^{-5\ B}$	**$7{,}24x10^{-5}$** **$5{,}28x10^{-5\ A}$**
Q_w ms^{-1}	mean median	**$2{,}02x10^{-6}$** **$1{,}15x10^{-6\ AB}$**	$8{,}05x10^{-7}$ $1{,}38x10^{-7\ B}$	$9{,}21x10^{-7}$ $6{,}75x10^{-7\ A}$
T m^2s^{-1}	mean median	$3{,}59x10^{-5}$ $1{,}82x10^{-5\ AB}$	$2{,}13x10^{-5}$ $8{,}20x10^{-6\ B}$	**$3{,}72x10^{-5}$** **$2{,}75x10^{-5\ A}$**
K ms^{-1}	mean median	**$1{,}04x10^{-6}$** **$5{,}99x10^{-7\ AB}$**	$4{,}15x10^{-7}$ $7{,}30x10^{-8\ B}$	$4{,}74x10^{-7}$ $3{,}51x10^{-7\ A}$

Based on the study results and test well experiences in Central Finland, a conceptual hydrotectonic model of the local hydrogeologic system was developed to guide the selection of the most favorable well sites (Fig. 118). The lineament-stress analysis forms the fundamental basis of the model. In addition, the optimal field for bedrock seismic velocity has been included in the model to aid in selecting the actual sites for drilling.

The information presented in the hydrotectonic model are currently factored into the groundwater exploration strategies in Central Finland. The model can also be applied to other crystalline rock settings especially in formerly glaciated areas in Fennoscandia and northern North America. However, expectations should not be too optimistic. Despite the general productivity, even critically stressed lineaments with optimal bedrock seismic velocities are no guarantee of siting successful wells, as they sometimes also possess low-productivity wells. The proposed criteria do not either mean that the probability of attaining high yields in wells drilled in supposedly unfavorable sectors is nil. Crystalline rock aquifers being very heterogeneous, these possibilities cannot be excluded (e.g. Lachassagne et al. 2001).

Finally, it must be stressed that - despite any knowledge or technique that aids in the search of higher yielding well sites - only the drilling and performance of aquifer pumping tests provide the decisive information required to determine if sufficient groundwater exists to support the safe and long-term withdrawal of large quantities of groundwater from a crystalline bedrock aquifer (e.g. Talkington 2004).

Water quality issues are of utmost importance when prospecting high-yield well sites. In Central Finland, bedrock groundwater quality and factors affecting it have been under constant monitoring and research from the mid 1980's. The noble expectation is that the results of this work will be published in the forthcoming years.

Economic and logistical criteria should also be considered when selecting potential drilling sites for major water supply. These include, for instance, land ownership and the proximity to roads, water lines, and electrical power lines. Environmental criteria to be considered include proximity to landfills, underground storage tanks, septic tanks, industrial or agricultural activities, or other potential sources of contamination (e.g. Mäkelä 1990b, Clarke & McFadden 1991, Dinger et al. 2002, Batte et al. 2008b).

Further systematic work in the field of high-yield well siting in crystalline basement areas is to be desired, both on a regional and local scale, with a combination of seismotectonic, hydrogeologic and hydrogeophysic studies.

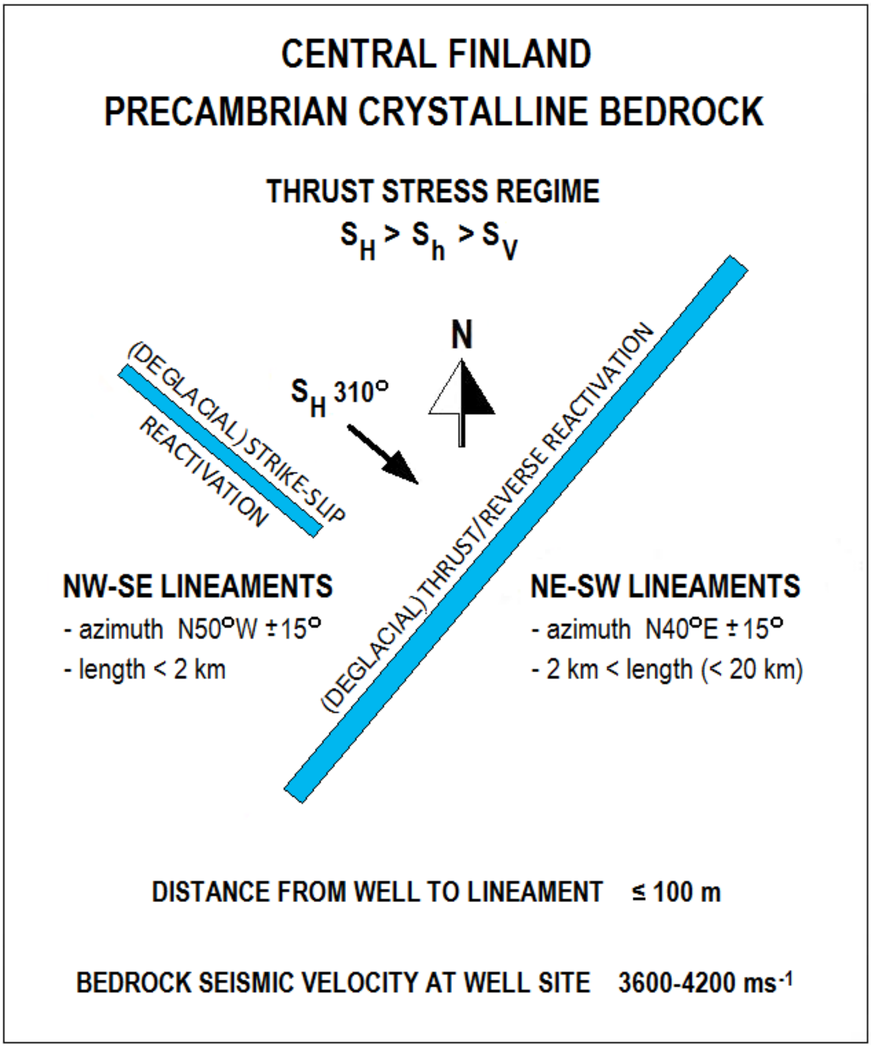


Fig. 118. Conceptual hydrotectonic model for selecting high-yield drilled well sites in the crystalline bedrock of Central Finland.

9 SUMMARY

The drilled well yield and hydraulic properties and their relationships to different well factors related to the location of the wells were investigated in the Precambrian crystalline bedrock of Central Finland. Data from 2,352 private wells constituted the primary study material. In addition, test well data from 73 drilled wells sited using various hydrogeological and hydrogeophysical techniques and expertise were utilized as a support material.

Based on technical well data and single well pumping tests, estimates for hydraulic parameters, which included normalized yield, specific capacity, well productivity, transmissivity, and bulk hydraulic conductivity, were statistically determined. Nearly 60 well factors were extracted. They were divided into five groups: construction, geologic, topographic, lineament and catchment factors. In addition, the role of seismotectonics was considered.

The main results and conclusions of the study are briefly outlined as follows:

Drilled well yield and hydraulic properties in Central Finland

1. Some 9,000 bedrock wells have been drilled in the study area since the late 1940’s. Their present groundwater abstraction is around 10,000 m^3d^{-1}, which is 1% of the total bedrock groundwater potential in the area. Drilled wells are most often used in domestic water supply by single households and farms. In addition, some 30 villages and small towns use bedrock groundwater for their common water supply. Bedrock groundwater can provide a valuable source for water supply of large communities in various times of crisis. Hydraulic fracturing has proven to be a successful method in increasing the yield of low to medium yield wells.

2. The median drilled well depth and yield are 73 m and 700 Lhr^{-1}, respectively. The median hydraulic parameters are as follows: normalized yield (Q/d_s) 12 $Lhr^{-1}m^{-1}$, specific capacity (Q/s) 50 $Lhr^{-1}m^{-1}$, well productivity (Q_w) $2{,}1x10^{-7}$ ms^{-1}, transmissivity (T) $7{,}3x10^{-6}$ m^2s^{-1}, and bulk hydraulic conductivity (K) $1{,}1x10^{-7}$ ms^{-1}. Much the same values have been reported in most unweathered non-carbonate and non-volcanic crystalline rock settings around the world despite different lithologies, climate and tectonic histories. This is suspected to be largely due to similar rock fracturing characteristics.

3. Q/d_s, Q/s and Q_w can be used as representative estimations for bulk T and K in regional studies of fractured rock aquifers. A prerequisite for this is, however, that their mutual relations have first been statistically adjusted. Empirical equations, which are derived in one geologic setting for estimating T or K, are seldom directly transferable to another.

Factors affecting well yield and hydraulic properties

4. Mainly due to the introduction of the DTH drilling technique in the mid 1970's the median drilled well depth increased from ca. 45 m to more than 100 m during the next two decades. Despite this, the median well yield remained unchanged or even decreased. In comparison to cable-tool wells, the DTH wells may offer better surface water protection due to longer casing and safer water supply from deep low-yield wells due to greater borehole water storage. The well yield and

hydraulic conductivity decrease downwards in bedrock, at least to the drilled well depths.

5. The soil type at well site and the thickness of overburden are not of any great importance to the well yield and hydraulic properties. However, their role in storaging groundwater may be significant. Lithological differences between well sites may be considered insignificant from the well production point of view. In Central Finland, it is possible to drill both very good and very poor boreholes in any rock type. The drilled well yield and hydraulic properties are not statistically related to surface water bodies. Neither does the land uplift rate nor does the highest shore level indicate any clear trend in well production properties.

6. In the study area, the productivity of a drilled well is clearly related to its topographic setting. Valley wells are most productive whereas hilltop wells, though deepest, yield the least amounts of water. Also the distance to the nearest (bedrock) hilltop and the relative height differences in a well's catchment area are statistically related to the well yield and hydraulic properties.

7. The proximity of lineaments is considered the most important factor entity controlling the productivity of drilled wells in the study area. The most significant single lineament factors are the azimuth and prominence of lineaments and their perpendicular distance to the drilled wells. The median well yield and hydraulic properties are at their highest along medium length to long NE-SW lineaments and on short NW-SE lineaments. Lineament intersections have no statistical relations to the well production properties.

8. Lineament centre (≤ 100 m) wells have nearly one order of magnitude higher median yields and hydraulic values compared to the wells situated well beyond (> 300 m) lineaments. High-capacity drilled wells are still found in areas without any distinct lineament pattern. This suggests that many water-bearing fracture zones go undetected in the lineament mapping. Some high-yield wells obviously get their water from fractures, which are not related to lineaments in any way.

Role of seismotectonics

9. Tectonic reactivation of faults is considered of utmost importance for groundwater flow and drilled well hydraulics. Analogously to the large postglacial faults in northern Fennoscandia, favorably orientated and critically stressed NE-SW and NW-SE lineaments have most probably been activated in the study area during the last phases of the Weichselian deglaciation some 10,000 years BP. This well explains their higher permeability compared to other lineament sets and may be considered as an implication of postglacial faulting in Central Finland.

10. The medium length to long NE-SW lineaments, which are orientated perpendicular to the maximum horizontal stress S_H, were activated by tectonic thrusting while the short NW-SE lineaments experienced strike-slip movement. The crust stabilization and stress accumulation beneath the ice sheet during the glaciation and the abrupt stress release during the deglaciation were most probably of a minor scale in the study area in comparison to the northern Fennoscandia, where they lead to severe earthquakes and huge faults scarps on the ground surface. Elevated well yields elsewhere in glaciated terrains of Fennoscandia and

northern North America are possibly related to similar fault reactivation during the deglacial time.

11. The possibility of postglacial fault reactivation during glacial cycles and its consequences for groundwater movement are taken into account in various arrangements of the future disposal of nuclear fuel waste in Fennoscandia. The results of this study show, however, that a better understanding of bedrock hydrogeology and paleohydrogeology is still needed.

Criteria for siting high-yield drilled wells in crystalline bedrock

12. High-yield well siting in crystalline rock areas should always start with lineament mapping. Along with remote sensing data, topographic maps suit well for lineament interpretation. The key point in obtaining the highest probability of success is to be able to identify those lineament sets, which are optimally orientated and critically stressed in a present-day stress regime and/or have been geologically recently reactivated. Hence, in a thrust fault regime, one should go for detecting lineaments that strike perpendicular to the maximum horizontal stress direction S_H. In normal fault regimes the most promising lineaments should lie parallel to the S_H, while in strike-slip regimes they either coincide with the S_H or diverge at various angles ($<< 45°$) to it.

13. In the thrust regime of Central Finland, the NE-SW and NW-SE lineaments should be preferred if possible to other lineament sets. Their central parts (≤ 100 m) are optimal for high-yield well drilling. Along with the lineament orientation, the prominence of lineaments must also be considered. The long (> 10 km) NW-SE lineaments, although being the most prominent lineaments, are not recommended as drilling sites for major water supply because of their low to moderate permeability and a marked number of low-yield wells. Instead, the short (< 2 km) NW-SE lineaments frequently are highly productive and may well be suggested as potential drilling sites for high-capacity wells. The situation is largely contrary with the NE-SW lineaments, where the medium length to long lineaments seem to be most advantageous for high-yield well drilling. However, the most prominent NE-SW lineaments (>> 10 km) might not suit for high-capacity well sites.

14. Although not as successful as lineament-stress analysis, the topographic approach can also serve useful information for high-yield well siting in crystalline rock areas. In the study area, staying in low-elevation areas away from hilltops and ridges statistically increases the chances to catch a high-yield well site. The siting of high-yield wells may also benefit from previous borehole information and water well inventories to be executed in the area of interest.

15. A conceptual hydrotectonic model of the local hydrogeologic system was developed to guide the selection of the most favorable well sites in the study area. The lineament-stress analysis forms the fundamental basis of the model. In addition, the optimal field for bedrock seismic velocity derived from test well investigations has been included in the model to aid in selecting the actual sites for drilling. It is stressed that only the drilling and proper pumping test with water quality determinations provide the ultimate information required to decide whether a high-yield well site will suit for long-term withdrawal of groundwater.

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APPENDIX I Unpublished reports from bedrock groundwater investigations 1983-2008 (in Finnish). Centre for Economic Development, Transport and the Environment for Central Finland (CETECF).

NO	MUNICIPALITY	SITE/VILLAGE	YEAR
1	SAARIJÄRVI	Mahlu	1983
2	KIVIJÄRVI	Purala	1984
3	VIITASAARI	Kotvala	1986
4	VIITASAARI	Keihärinkoski–Valkeisjärvi	1986
5	SAARIJÄRVI	Mahlu	1987
6	PIHTIPUDAS	Seläntaus	1987
7	SAARIJÄRVI	Kotimäki-Kuoppala-Lehtomäki	1987
8	VIITASAARI	Huopana-Jurvansalo-Keitelepohja-Niinilahti	1988
9	VIITASAARI	Permoskylä-Soliskylä-Tereniemi-Viitakangas	1988
10	SAARIJÄRVI	Paajala-Pajumäki-Puolimatka	1988
11	UURAINEN	Höytiä	1989
12	JOUTSA	Tammihaara	1989
13	SAARIJÄRVI	Pajupuro	1989
14	SAARIJÄRVI	Koskenkylä-Lehtola	1989
15	JOUTSA	Rutalahti	1991
16	PETÄJÄVESI	Kintaus	1991
17	KEURUU	Jukojärvi	1992
18	KANNONKOSKI	Kannonjärvi	1992
19	UURAINEN	Kirkonkylä	1993
20	VIITASAARI	Vuorilahti	1994
21	VIITASAARI	Ilmolahti	1996
22	KIVIJÄRVI	Lokakylä	1997
23	KANNONKOSKI	Käräjämäki	1997
24	LUHANKA	Kirkonkylä	1998
25	JOUTSA	Kälä-Mieskonmäki	1998
26	JYVÄSKYLÄ	Moksi	1998
27	JOUTSA	Uimaniemi	1998
28	JYVÄSKYLÄ	Oittila	1998
29	JYVÄSKYLÄ	Putkilahti	1998
30	JÄMSÄ	Mon Repos	2000
31	HANKASALMI	Kärkkäälä	2002
32	VIITASAARI	Keitelepohja	2003
33	KONNEVESI	Kirkonkylä	2003
34	KUHMOINEN	Harmoinen	2003
35	HANKASALMI	Säkinmäki	2004
36	PETÄJÄVESI	Kumpujärvi	2004
37	VIITASAARI	Luotolansaari	2004
38	VIITASAARI	Kiminki	2004
39	SAARIJÄRVI	Koskenkylä	2004
40	HANKASALMI	Ristimäki	2005
41	VIITASAARI	Luotolansaari	2006
42	TOIVAKKA	Heiska	2008

APPENDIX II Drilled well information used in testing the reliability of the questionnaire data (n=42). The explanations of different abbreviations are given in Table 3.

WELL ID	WELL DRILLERS´ DATA						WELL OWNERS´ DATA					
	CY	DR	OVER m	DEPTH m	DIA mm	YIELD Lhr^{-1}	CY	DR	OVER m	DEPTH m	DIA mm	YIELD Lhr^{-1}
077015	1990	3	1	96	156	4000	1991	3	2	95	160	4000
172027	1962	20	4	50	150	864	1962	20	3	50	150	750
172040	1992	3	7	109	160	650	1992	3	7	109	160	650
172092	1990	1	20	47	115	5000	1990	1	20	47	115	5000
172094	1990	2	12	88	115	2000	1990	2	18	88	.	2000
172097	1992	2	4	178	116	400	1992	2	4	178	125	400
172098	1992	3	0	127	160	750	1992	3	3	127	160	750
179235	1976	3	0	75	110	200	1976	3	0	85	.	.
179240	1975	3	1	54	110	2400	1975	3	0	54	120	1500
182018	1976	3	2	41	110	400	.	3	.	.	.	.
182019	1976	3	3	41	115	3000	1976	3	0	.	.	.
182066	1987	3	1	96	155	300	1976	3	1	96	60	300
182080	1989	3	0	79	158	2500	1989	3	0	79	158	2500
182190	1971	3	3	74	110	400	1972	3	3	76	100	450
182232	1989	3	0	103	155	720	1989	3	2	103	155	720
226001	1978	3	5	72	112	300	1976	3	3	80	110	400
249015	1978	3	4	65	110	350	1978	3	4	67	100	350
249031	1991	1	9	127	115	1080	1991	1	6	125	115	1080
249046	1992	3	0	175	160	90	1992	3	3	175	160	90
275029	1992	3	36	139	160	50	1992	3	36	139	160	80
275034	1953	19	.	25	110	1000	.	19	3	23	.	.
291022	1989	3	1	337	155	220	1989	3	1	337	155	220
410044	1989	21	40	106	110	500	1989	21	.	118	100	.
410065	1954	19	.	51	110	190	.	19	1	56	.	250
410066	1956	19	.	30	110	170	.	19	1	30	.	200
435039	1961	20	2	16	110	350	1961	20	2	16	120	400
435098	1961	20	6	26	110	1800	1962	20	2	36	110	1500
495001	1977	3	1	80	110	400	1976	3	1	79	110	300
495002	1977	3	3	50	110	700	1978	3	3	56	110	800
495004	1978	3	3	82	110	700	1978	3	4	87	100	300
601019	1982	3	2	59	115	250	1981	3	2	59	80	250
601029	1957	19	9	50	110	1800	1957	19	10	50	110	1800
601032	1957	19	.	83	110	1200	1957	19	4	82	127	2000
729044	1958	19	3	28	110	150	1958	19	3	31	100	300
729105	1991	1	1	94	115	180	1991	1	1	94	115	100
729234	1991	1	25	25	125	1600	1991	1	25	25	140	1600
931020	1987	1	1	29	116	210	1987	1	1	29	100	230
931067	1988	3	4	109	155	150	1988	3	4	108	155	150
931087	1989	2	15	61	151	20000	1989	2	15	60	150	.
931103	1991	1	15	48	115	6000	1991	1	12	48	.	6000
931105	1991	1	7	64	115	700	1991	1	9	60	80	700
931114	1990	1	2	85	115	2400	1990	1	.	85	115	2400
N = 42	42	42	38	42	42	42	38	42	39	40	34	36

APPENDIX III Results from testing the adequacy of 25 sampling points for the land use of wells´ catchment areas (n=50). The explanations of different abbreviations are given in Table 3.

WELL	RESULTS FROM 25 SAMPLING POINTS %						RESULTS FROM 50 SAMPLING POINTS %					
ID	BED	FOR	PEA	CUL	SET	WAT	BED	FOR	PEA	CUL	SET	WAT
077015	0	52	0	0	8	40	0	52	0	0	8	40
172027	0	56	0	24	4	16	2	38	6	36	6	12
172040	0	64	0	28	8	0	0	64	2	30	4	0
172092	8	48	0	20	24	0	4	52	0	18	26	0
172094	0	68	0	16	16	0	0	70	2	14	14	0
172097	0	68	4	24	4	0	0	66	4	24	6	0
172098	4	24	4	16	12	40	2	34	6	12	14	32
179113	0	76	0	20	4	0	2	68	0	24	6	0
179235	0	48	0	44	8	0	0	40	0	48	10	2
179240	4	48	0	28	16	4	8	46	4	30	10	2
182005	24	48	0	20	4	4	32	42	0	16	5	4
182010	4	40	4	36	12	4	6	38	2	42	10	2
182012	0	36	40	12	4	8	0	40	38	8	4	10
182018	8	56	4	16	16	0	6	58	0	18	16	2
182019	4	72	0	8	8	8	4	64	0	10	8	14
182066	0	68	12	16	4	0	0	74	10	12	4	0
182080	24	44	4	20	8	0	24	46	2	22	6	0
182190	12	12	0	48	8	20	12	14	0	44	10	20
182218	4	68	0	16	8	4	2	68	2	14	8	6
182232	8	76	8	0	8	0	4	84	4	2	6	0
226001	0	60	0	20	20	0	2	52	2	28	16	0
249015	8	60	0	20	4	8	6	68	4	12	4	6
249031	4	36	4	40	4	12	2	40	6	26	12	14
249046	8	56	4	8	12	12	4	58	6	4	12	16
275029	0	64	0	20	16	0	0	58	0	22	20	0
275034	0	48	0	28	12	12	0	50	2	26	14	8
291022	20	40	20	12	8	0	20	50	10	6	12	2
410044	0	40	4	16	12	28	0	42	4	22	8	24
410065	12	40	0	40	8	0	10	40	0	40	8	2
410066	4	32	0	0	64	0	2	38	0	0	60	0
435039	12	44	0	36	8	0	14	44	0	32	8	2
435098	12	64	0	16	4	4	12	58	0	18	6	6
495001	0	68	12	16	4	0	0	56	18	22	4	0
495002	0	76	4	16	4	0	4	72	6	12	6	0
495004	0	40	8	44	8	0	0	46	4	42	8	0
601019	8	48	0	24	12	8	6	46	0	30	8	10
601029	0	20	4	40	36	0	0	18	0	36	46	0
601032	0	12	0	48	4	36	0	10	0	38	4	48
729044	4	32	0	44	20	0	2	38	0	42	18	0
729105	0	92	0	4	4	0	2	90	0	4	4	0
729234	0	44	4	20	12	20	0	52	2	22	6	18
931015	0	60	8	28	4	0	0	66	4	24	6	0
931020	0	72	8	16	4	0	0	74	8	14	4	0
931065	0	56	0	20	12	12	2	36	0	36	14	12
931067	0	88	0	8	4	0	0	88	0	8	4	0
931087	0	56	12	12	16	4	0	52	10	14	16	8
931092	0	16	4	44	8	28	0	16	2	48	10	24
931103	4	40	0	24	4	28	4	30	0	30	8	28
931105	8	72	0	16	4	0	12	68	0	16	4	0
931114	16	56	0	0	4	24	12	62	0	0	6	20