

A Fluidization Breccia in Granite at Skaget, Svellingen, Frøya

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The breccia consists of fragmented and comminuted host rock material and small amounts of hydrothermal quartz, epidote, laumontite, ?stilbite, calcite and montmorillonite. At least three brecciation and mineralization events have occurred. The breccia is considered to have formed by hydraulic fracturing by a mixed, high-pressure CO₂-H₂O fluid. Upon explosive venting to the surface, the fluid is thought to have fluidized and fragmented the fractured rock. The carbon dioxide of the fluid may have originated in the mantle, and could have ascended through the crust by hydraulic fracture propagation. It is suggested that the occurrence within the same region of the Skaget Breccia, Tertiary olivine-nephelinite plugs, and scattered zeolite mineralization may possibly define a palaeo-geothermal province.

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Introduction

A small breccia body cutting through granite at Skaget, Svellingen, Frøya (UTM co-ordinates NR 912752; Fig. 1) was described by Keilhau (1850) and mentioned by Reusch (1914), but has not been treated in newer literature. It occupies the tip of a small, pointed headland called Skaget (Figs 2 & 3). On land, the Skaget Breccia covers about 5000 m², but is surrounded by the sea on three sides; accordingly, its full extent and shape are unknown. In addition to the main breccia outcrop, there are a few, scattered, small occurrences of breccia over a distance of a couple of hundred metres southward along the coast.

Keilhau (1850) regarded the breccia as an epiclastic sedimentary rock, and interpreted its gradational boundary with the surrounding granite in terms of his neptunistic ideas: as evidence for the transformation of a sedimentary rock into granite. The present study indicates that the rock was formed by fragmentation of the host granite and fluidization of the fragments by streaming gas.

Regional and field geology

Host rock to the Skaget Breccia is the Frøya Granite, a large, homogeneous pluton of granitic to granodioritic composition. It covers the northern part of Frøya and the adjoining archipelago, including the Sula islands (Fig. 2). The granite is intrusive into migmatites and gneisses of the Northwestern Gneiss

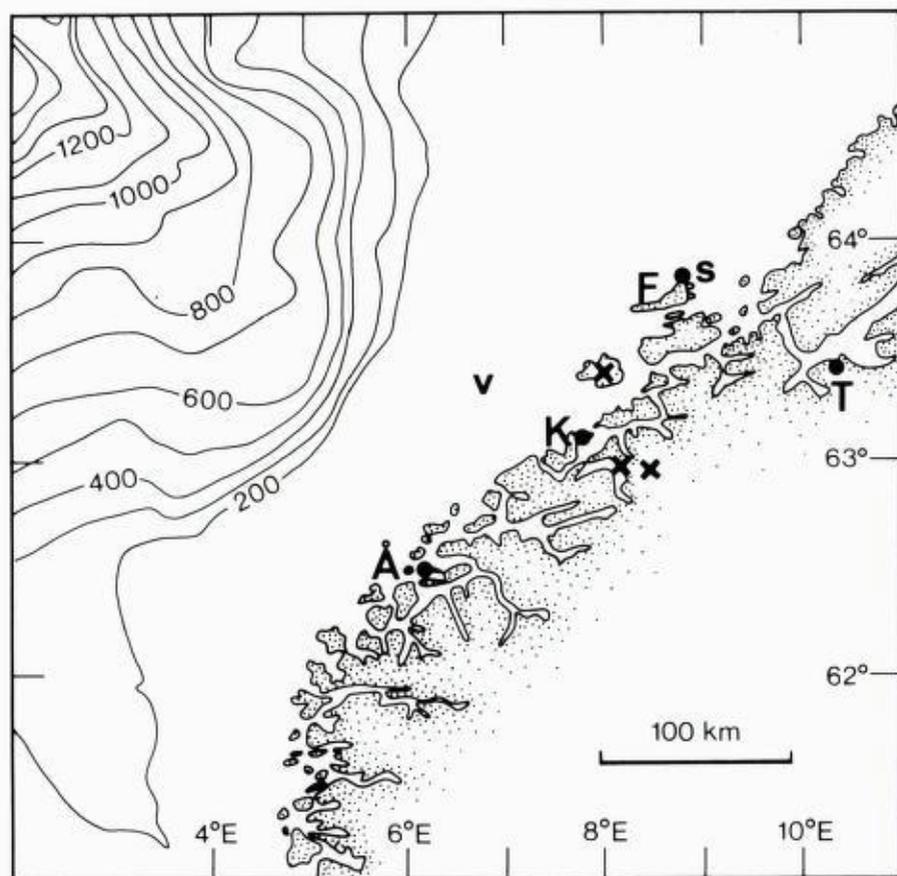


Fig. 1. Location map of west-central Norway. S: Skaget; F: Frøya; V: Olivine nephelinite plugs; crosses: known zeolite occurrences. K: Kristiansund; T: Trondheim; Å: Ålesund. Isobaths in fathoms.

complex of the Norwegian Caledonides (Strand 1960, Oftedahl 1981). In places, the gneisses contain small bodies of partly silicified marble (Reusch 1914). A provisional, model Rb.-Sr dating gave a Caledonian minimum age for the Frøya Granite (B. Sundvoll, pers. comm.).

The Skaget Breccia occurs well inside the pluton, about four kilometres from the nearest, steeply dipping contact between the Frøya Granite and the gneisses adjoining it to the south. It consists of fragments and matrix of comminuted host-rock material and secondary hydrothermal minerals; predominantly quartz, minor epidote, and small amounts of Ca-zeolites, montmorillonite and calcite.

There are two, texturally different types of breccia (Fig. 3): a dark, unsorted, granite breccia (Fig. 4) and a light silicic microbreccia (Fig. 5a, b). The latter forms a dyke-like sheet along part of the boundary between the main, granite breccia and the host granite (Fig. 3). Its emplacement post-dated that of the local granite breccia, because it cross-cuts a small portion of the main breccia body. On the other hand, quite similar microbreccia occurs

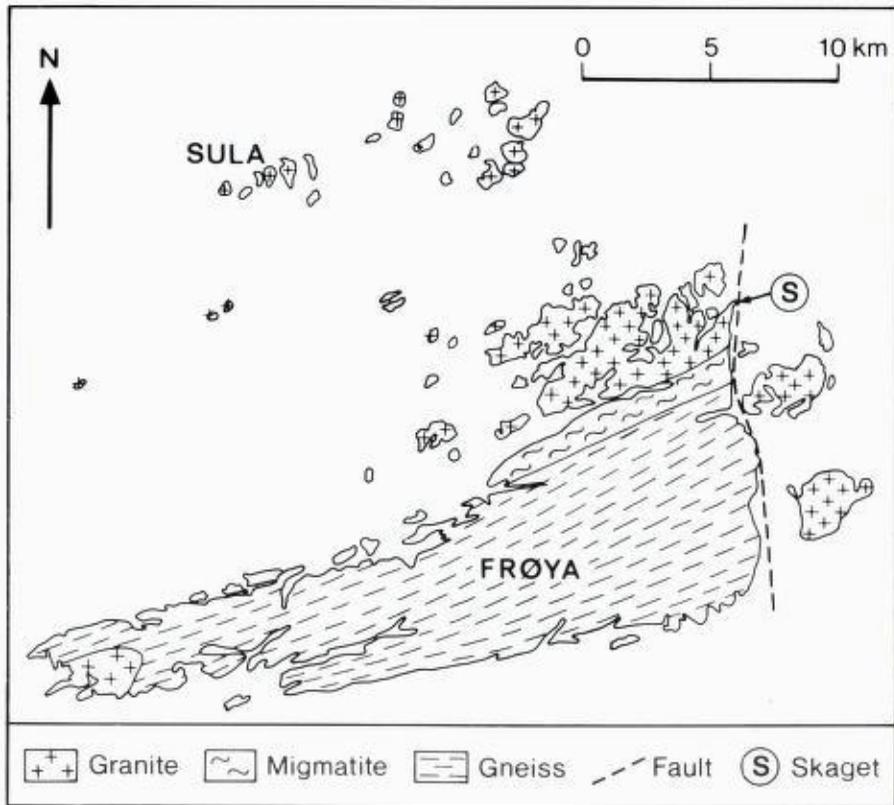


Fig. 2: Bedrock geology of Froya. Redrawn after Askvik (1979). S: Skaget.

as fragments in the granite breccia near the boundary between the two breccia types. The microbreccia sheet is sharply delimited, up to 12 m thick, and dips about 35° ESE. This orientation parallels that of a joint system in the bordering granite. The limited exposure of the granite breccia precludes an assessment of the overall shape and extent of this variety.

Petrography and mineralogy

The Skaget Breccia and its surroundings may be described in terms of three main rock types: 1) the host granite; 2) the granite breccia; and 3) the microbreccia.

1) *The Froya Granite* is a massive, medium to coarse-grained, red, biotite granite to granodiorite. In places it is semi-porphyric, with up to several centimetres long, stubby megacrysts of brick red microcline. The megacrysts may locally have a crudely defined parallel orientation. Sparse and scattered, dark, thin, long, biotite schlieren have the same orientation as the feldspar megacrysts, and are probably strongly flattened and partly digested xenoliths.

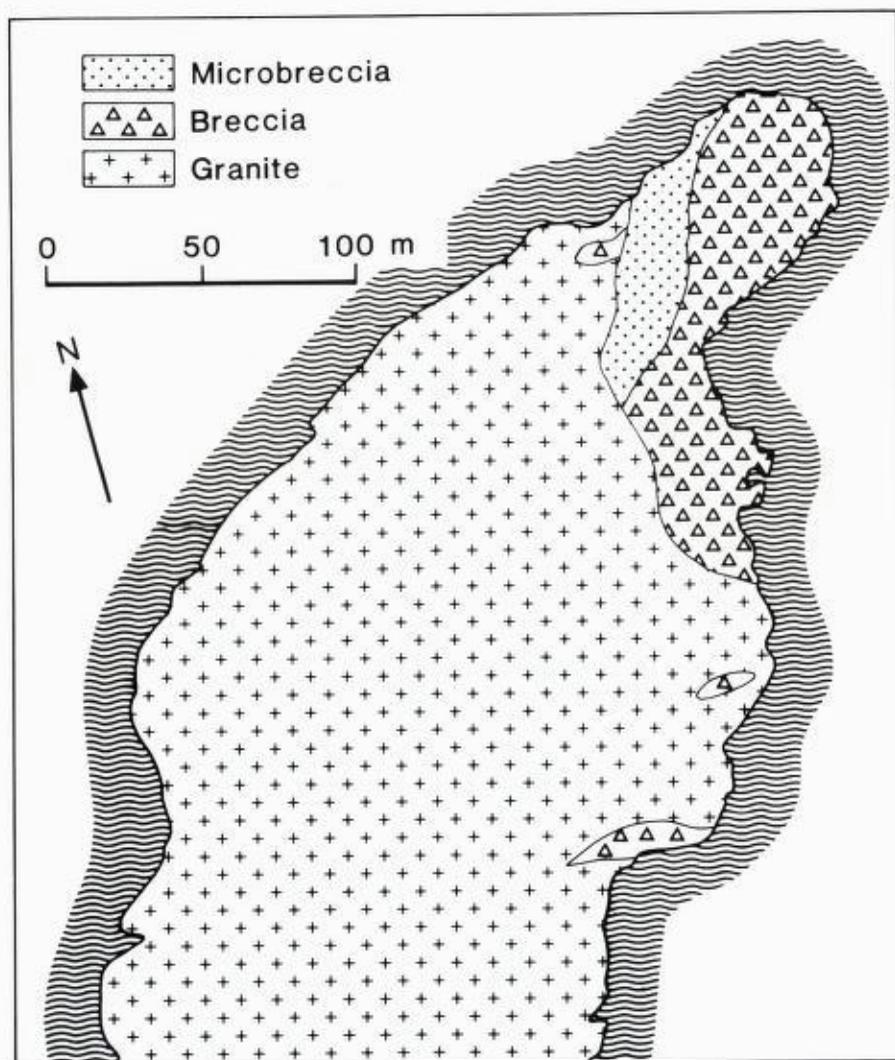


Fig. 3. Geological sketch map of Skaget.

Some granite pegmatite dykes have up to 10 cm long, red, euhedral K-feldspar crystals, much quartz, and subordinate plagioclase. Pegmatites with smaller grain sizes form narrow, straight and parallel-walled dykes, a few centimetres wide and up to some tens of metres in length. The granite has a variably developed parallel jointing, with joints spaced from a few decimetres to a couple of metres apart. They appear to be oriented more or less perpendicular to the indistinct foliation, defined by parallel oriented feldspar megacrysts and biotite schlieren.

Along its external border, against the gneiss complex to the south, the Frøya pluton is associated with a zone of migmatite. There, more or less irregular dykes and veins of massive red granite, muscovite-bearing aplite, and pegmatite invade the older gneisses.



Fig. 4. Granite breccia, Skaget. Subangular to rounded granite fragments in unsorted matrix. Coin is 2.5 cm across.

The texture of the typical Frøya Granite is hypautomorphic-granular (Johannsen 1939); with subhedral to anhedral, predominantly unstrained microcline and albite-oligoclase crystals, unoriented flakes of biotite, and intersertal, granular aggregates of fine- to medium-grained, anhedral, strained quartz. Sphene is the predominant accessory mineral, but apatite is also very common. In addition, there are scattered grains of zircon and allanite. The plagioclase may be slightly sericitized, but has apparently not been epidotized in the typical Frøya Granite. Biotite may show incipient chloritization.

In the nearest several metres to the breccia boundary, the Frøya Granite is strained, fractured and hydrothermally altered. The characteristic deformation features are: strain effects in plagioclase, like undulatory extinction and bent or kinked twin lamellae; and small-scale fracturing and veining by microbreccia and quartz, as well as by laumontite. The effects of hydrothermal alteration are: complete chloritization of biotite, minor adularization as well as partial epidotization of plagioclase, and local dissolution of primary, strained quartz and subsequent deposition within the same space of unstrained, subhedral quartz grains.

In many cases chlorite flakes bordering on plagioclase are accompanied by a K-feldspar rim in the latter, in such a way that the two secondary minerals together form a composite pseudomorph after biotite (Fig. 5 c). From the restricted scale of this type of plagioclase alteration, and from the apparent non-participation of the primary K-feldspar of the granite in these secondary reactions, it may be inferred that potassium for the secondary K-feldspar in

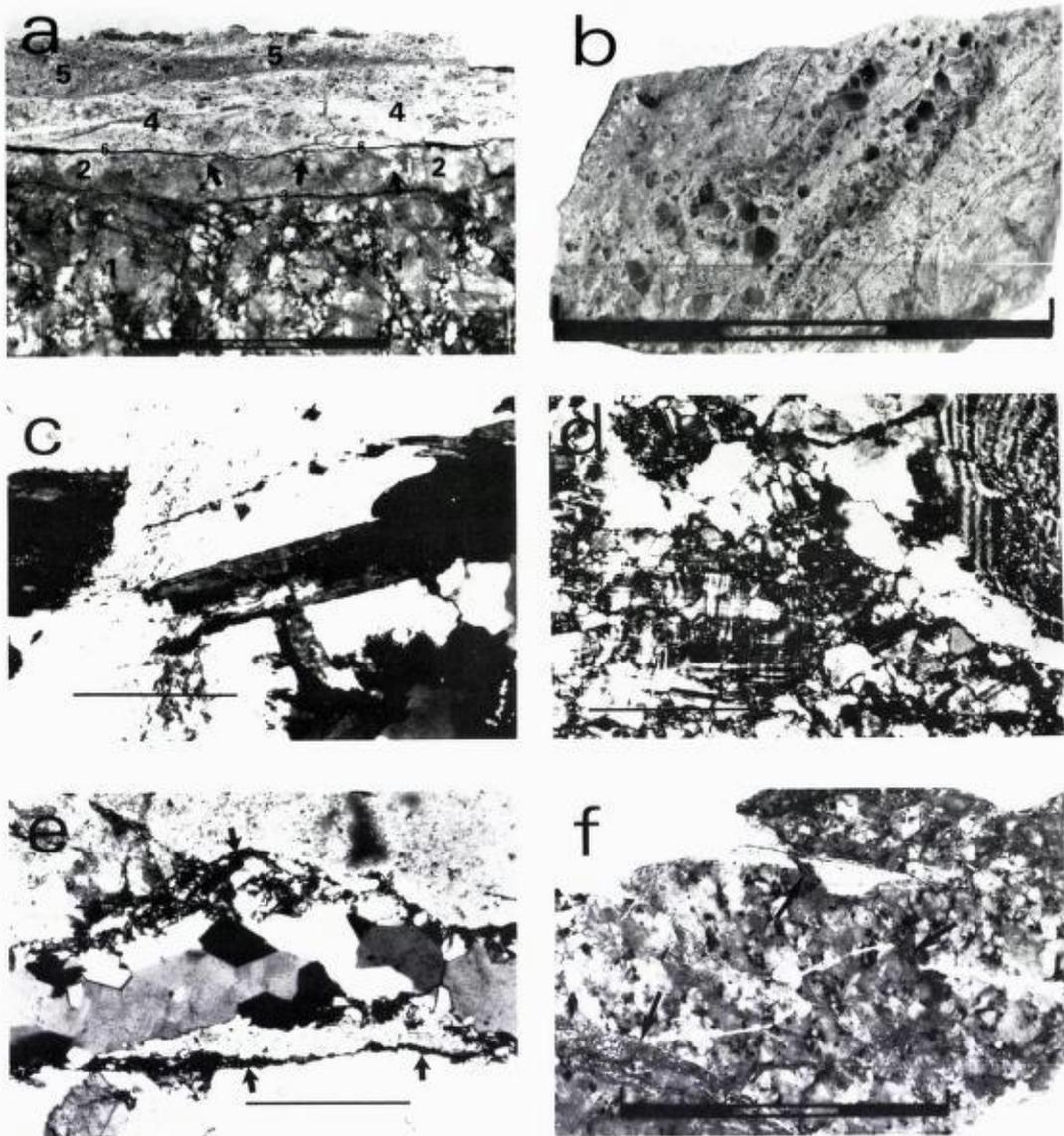


Fig. 5(a): Border between host granite (below) and micro breccia: 1, microfractured granite; 2, quartz-free rim, consisting of highly strained feldspar (cf., Fig. 5 d), and with microbreccia-filled embayments (arrows); 3, younger quartz vein (cf., Fig. 5 e); 4 and 5 are two generations of microbreccia; crack (i) is sampling artifact. Polished slab. Bar scale in cm.

(b) Microbreccia with train of rock crystal fragments. Polished slab. Bar scale in cm.

(c) Adularia (medium grey) and chlorite (dark) pseudomorph after biotite in plagioclase (light). Frøya granite. Planepolarised light. Bar scale = 0.5 mm.

(d) Severely strained plagioclase and microcline in feldspathic rim in granite bordering against microbreccia (cf., Fig. 5a). Crossed nicols. Bar scale = 0.5 mm.

(e) Quarts-filled fracture in granite (cf., Fig. 5a). Note euhedral outlines of some quartz grains, and older microbreccia-filled fractures (arrows). Crossed nicols. Bar scale = 0.5 mm.

(f) Laumontite veins (light) in granite offset by younger microbreccia-filled fractures (arrows). Polished slab. Scale in cm.

plagioclase was derived mainly from the breakdown of local biotite, in general accordance with the process proposed by Chayes (1955). Most K-feldspar patches in plagioclase are turbid; they have very low birefringence and small but indeterminate, apparently variable $2V_x$, indicating adularia; and thus pointing to hydrothermal alteration of the affected plagioclase. Epidote in plagioclase consists mostly of fairly large grains, frequently only one large epidote grain per plagioclase host grain, rather unlike the fine-grained products of typical saussuritization.

The strain effects in plagioclase and quartz give evidence of ductile deformation of these minerals. Microcline appears to be unaffected by this, except at the immediate contact with the microbreccia (Figs. 5a, d). There, a feldspathic rim, about 5 mm thick, shows evidence of intensive ductile as well as brittle strain. This, together with intensive micro-fracturing, indicates that the deformation behaviour of the granite in the contact zone was near the brittle-ductile transition field. The thin rim with highly strained feldspar appears to have lost its primary quartz content by dissolution. It is embayed by microbreccia-filled solution pits and cracks (Fig. 5 a), and invaded by younger, unstrained veins of hydrothermal quartz (Figs 5 a, e).

Four types of microfracture occur in the altered granite and in larger fragments in the breccia: a) healed fractures (Batzle & Simmons 1976), where fractured mineral grains have recrystallized across the fracture; b) sealed fractures (*ibid.*), where new mineral grains have crystallized within the fracture (Figs 5 e, f); c) filled fractures, where minute fragments, forming a microbreccia, occupy the fracture space (Fig. 6 a); and d) closed fractures, where little or no material intervenes between the fracture walls; these are in immediate contact, and only small offsets along the crack or, for instance, a thin coating of iron oxide delineate the fracture. Any particular fracture may change type one or several times along its extent. The fracture-sealing mineral is predominantly quartz (Fig. 5 e), often with small amounts of epidote, but laumontite veins have also been identified (Fig. 5 f). Some laumontite veins are cut by thin, microbreccia-filled fractures (Fig. 5 f). Calcite is a very subordinate vein mineral.

2) *The granite breccia* forms the major part of the Skaget occurrence. It consists of more or less rounded, unsorted, granite clasts and mineral fragments, interspersed with scattered fragments of light-coloured microbreccia and vein quartz; all set in a fine-grained to dense, greyish brown to green groundmass. The granite clasts are mostly less than 10 cm, but may reach 50 cm in size. Clasts of microbreccia are mostly less than 2 cm across.

In a few places, the granite breccia has irregular, locally anastomosing, steep, tubular or podiform cavities, up to 2 cm in diameter and possibly up to several tens of centimetres long. They are either empty, or filled with pasty, white, pure montmorillonite. Some are lined with a thin layer of zeolite or quartz and calcite.

The border between granite breccia and host granite is diffuse and irregular: there is a gradational transition over one to several metres from large

blocks of granite in situ, partially surrounded by thin veins of granite breccia, to matrix-supported breccia with rounded, smaller fragments. The boundary between the granite breccia and the microbreccia, on the other hand, is a well-defined, sharp, straight to smoothly undulating contact.

The fragments in the granite breccia vary in size from around 10 cm down to about 10 μm . The matrix is composed essentially of fine-grained to dust-sized, fragmented rock material (Fig. 6 b). Fragments of quartz and feldspar may show internal, pre-brecciation strain phenomena; whereas primary quartz from the granite is strained and polygonized, hydrothermal vein quartz fragments are nearly always unstrained. Flakes of chlorite may show externally induced deformation from being pinched between other mineral fragments or between the walls of closed fractures. The breccia matrix is cemented by secondary quartz and smaller amounts of epidote-clinozoisite; the grain size of the latter is only about 5 μm . In addition, very small but irregularly distributed amounts of zeolite minerals form thin veins and replacement textures. Zeolites partially replacing quartz and other minerals (Fig. 6 c) have been tentatively identified under the microscope as ?stilbite; and vein-forming laumontite was identified by X-ray diffraction methods, as was the montmorillonite mentioned above. Calcite occurs in trace amounts.

The bulk chemical composition of the granite breccia is very similar to that of the host granite (Table 1), the most notable difference being a higher content of Na_2O and slightly lower CaO and SiO_2 .

3) *The microbreccia* has well-defined and sharp boundaries against both the granite breccia and the granite. It sends a few small, thin veinlets into the latter. The rock is fine-grained to dense, and breaks with a rough, irregular to flinty fracture. It is light grey with varying shades of pale green, caused by finely dispersed epidote. Some microbreccia varieties have small, reddish stains of hematite. Thin, irregular quartz veins occur locally.

The microbreccia commonly contains small (less than 2 cm), more or less rounded fragments of similar but older microbreccia. Nearly all these fragments are transected by older quartz veins, clearly predating the latest fragmentation. Scattered, discrete quartz vein fragments are also macroscopically visible. In a few places, small concentrations of fragmented, euhedral low-quartz crystals are embedded as clasts in the microbreccia (Fig. 5 b).

Under the microscope, the microbreccia is seen to consist of a fine-grained, varied mixture of mineral fragments, mostly between 0.01 and 0.50 mm in size (Fig. 6 d). The fragments are mainly of feldspar and quartz, with trace amounts of chlorite. The feldspar and quartz fragments derived from granite may show severe, internal, pre-brecciation strain. The near-ubiquitous fragments of older, quartz-veined microbreccia are commonly richer in secondary epidote than the younger, surrounding microbreccia matrix. In other cases, however, the matrix of these older fragments of microbreccia is so similar to the younger host matrix that such fragments can be identified only by their old quartz veins, abutting against the fragment borders (Fig. 6 e). Even though the breccia matrices in these instances are almost identical, this

Table 1: Chemical analyses and CIPW norms of rock types at Skaget.

	Froya Granite wt %	Granite breccia wt %	Microbreccia wt %
SiO ₂	69.77	68.47	91.37
Al ₂ O ₃	15.58	15.92	3.68
CaO	2.19	2.05	0.97
MgO	0.91	0.79	0.10
Fe ₂ O ₃	0.72	1.40	0.96
FeO	1.58	1.28	0.21
Na ₂ O	4.00	5.29	1.23
K ₂ O	4.05	3.24	0.81
TiO ₂	0.40	0.46	0.09
MnO	0.05	0.03	0.00
P ₂ O ₅	0.08	0.11	0.02
CO ₂	0.09	0.11	0.04
H ₂ O ⁺	0.68	1.04	0.23
Total	100.10	100.19	99.71
CIPW norms			
Q	24.6	20.1	79.5
C	1.0	0.5	-
Or	23.9	19.1	4.8
Ab	33.8	44.8	10.4
An	9.8	8.8	2.1
Wo	-	-	0.7
Di	-	-	0.5
Hy	4.0	2.5	-
Mt	1.0	2.0	0.4
Hm	-	-	0.7
Il	0.8	0.9	0.2
Ap	0.2	0.3	-
Cc	0.2	0.3	0.1
H ₂ O	0.7	1.0	0.2
Total	100.1	100.2	99.7

Analyst: Per-Reidar Graff.

mode of occurrence of vein quartz proves that the microbreccia fragments did not form by agglomeration in the present host matrix, but represent one or more, separate, older microbreccia generations. Some microbreccia fragments carry even older, rounded, fragments of microbreccia (Fig. 6 f); thus, there are at least three generations of microbreccia.

The secondary minerals in the microbreccia are quartz and minute (about 5 µm) grains of epidote; in addition there are trace amounts of calcite. Unlike the granite breccia, the microbreccia does not appear to have been subject to zeolite and montmorillonite deposition.

Chemically, the microbreccia is highly siliceous, and is enriched in calcium relative to alkalis and alumina, as compared with both the Froya Granite and the granite breccia (Table 1). The microbreccia is diopside and wollastonite normative, whereas the two other lithologies are hypersthene and slightly corundum normative.

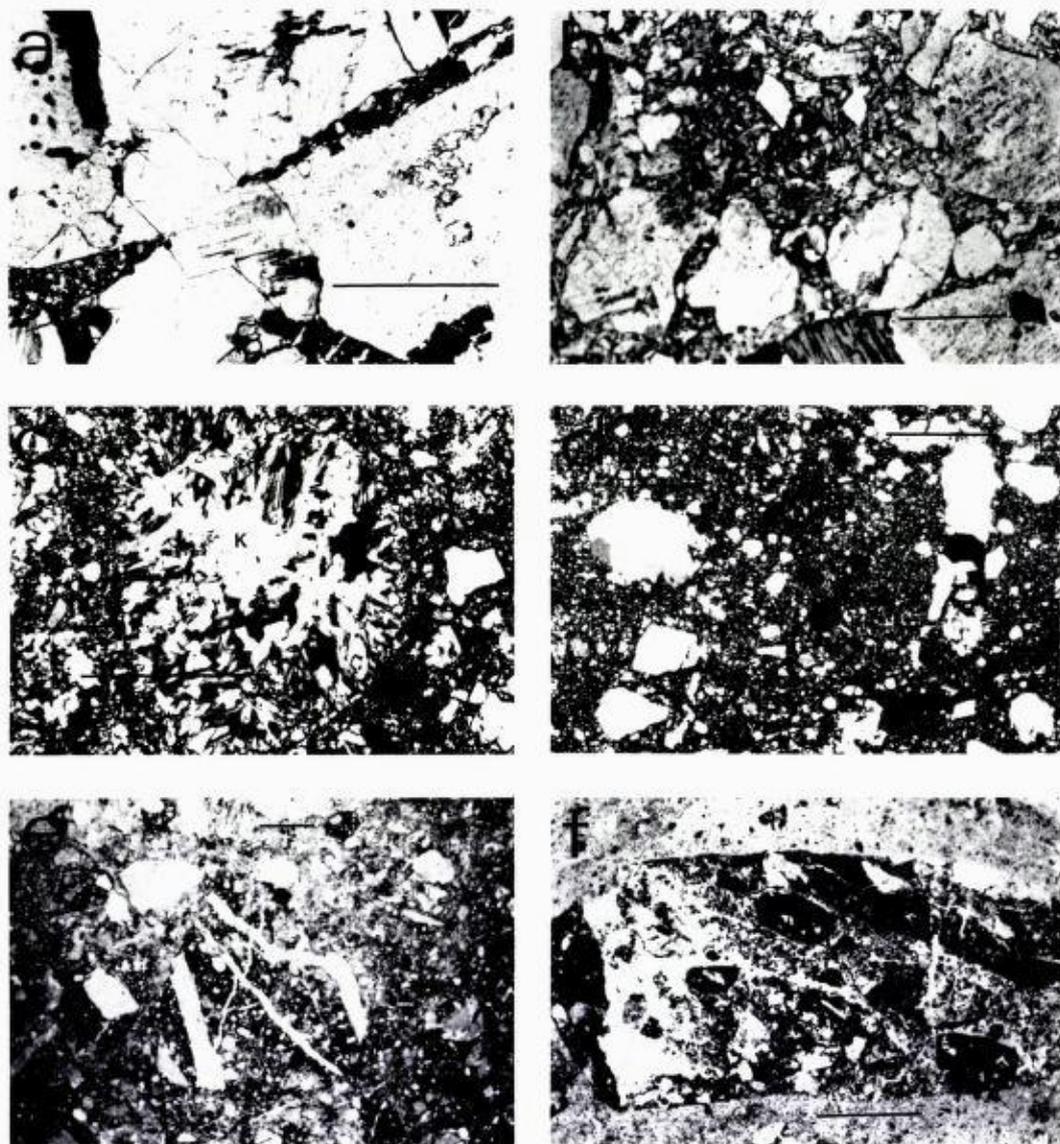


Fig. 6(a): Epidotized microbreccia-filled fractures in granite. Plane polarised light.

Bar scale = 0.5 mm.

(b) Granite breccia matrix; unsorted rock and mineral fragments. Plane polarised light.

Bar scale = 0.5 mm.

(c) Zeolitized fragment in granite breccia. ?Stilbite, replacing calcite (K), has grown more or less perpendicularly inward from the rim of the fragment. Crossed nicols.

Bar scale = 0.5 mm.

(d) Microbreccia with fragments of vein quartz. Crossed nicols. Bar scale = 0.5 mm.

(e) Quartz-veined microbreccia fragment in younger, similar microbreccia. Plane polarised light. Bar scale = 1 mm.

(f) Microbreccia fragment with older microbreccia fragments in microbreccia. The older fragments (dark) are more epidote-rich than younger matrix. Plane polarised light.

Bar scale = 0.5 mm.

Discussion

States of stress in the Earth's crust producing localized brittle fragmentation of rocks can arise from a variety of geological causes, such as: igneous intrusion and extrusion, tectonic activity, meteorite impact, and the release of thermal and potential energy from subsurface fluids.

The following points are considered important for an understanding of the formation of the Skaget Breccia. Brecciation by magmatic processes can be dismissed because no igneous material is present in or associated with the breccia; except the host granite, which clearly had crystallized completely prior to the breccia formation. Tectonic fragmentation appears unlikely because of the irregular and gradational boundary between the host granite and the granite breccia, and also because of the virtual lack of strain effects in fragments of vein quartz and other secondary minerals even in the most intensively comminuted fractions. Impact brecciation can be excluded as a mechanism because of the recurrence of brecciation events, separated by episodes of mineral deposition and veining; besides, no evidence of shock phenomena indicative of impact have been found.

On the other hand, brecciation of the granite by volatile fluid overpressure with subsequent gas streaming and fragment comminution appears to be consistent with all the known facts about the Skaget Breccia. The following discussion presents a short examination of this interpretation.

By combining the observations presented above, it is possible to identify some of the effects and properties of the fluid medium:

1) Initially, the fluid exerted a pressure sufficient to break up the Frøya Granite, probably by hydraulic fracturing (Fyfe et al. 1978), and to cause incipient ductile deformation of feldspar as well as dissolution of quartz in the wall-rock margin.

2) Through sudden pressure release, probably by explosive venting to the surface, the fluid thereupon expanded greatly and formed a fluidizing gas, which enabled inter-particle collisions among the granite fragments to abrade and comminute the material during vigorous, bubbling fluidization.

3) By entrainment and elutriation of the smallest particles, the streaming gas separated a fraction of fines from the main and unsorted part of the fluidized system to form the microbreccia in a marginal portion of the system (Woolsey et al. 1975).

4) When the fluidized material had settled after the eventual escape of the gas a hydrothermal, liquid solution following in its wake rapidly deposited minute grains of epidote/clinozoisite and, much more slowly, large amounts of quartz – some of it as euhedral, inclusion-free crystals. These have obviously grown in open vugs, cavities and fractures in the breccia. Subsequently, Ca-zeolites were formed along veins and fractures; some of them probably by reaction between the solution and unstable minerals (quartz, calcite), and montmorillonite was deposited in small cavities, which were formed as fluidization bubbles (Reynolds 1954).

The material deposited from solution was probably derived, at least in part, from quartz and plagioclase (anorthite) in the fragmented granite.

5) This sequence of events: fracturing, break-through to the surface, gas fluidization and hydrothermal mineral deposition was repeated at least twice (Fig. 6 f), in the same order and with apparently identical effects.

6) The temperature was probably never very high, since microbreccia-filled fractures cross-cutting and offsetting veins with unaltered laumontite in the granite and in breccia fragments indicate that the upper stability temperature of this zeolite (ab. 220 °C; Liou 1971, Bird & Helgeson 1981) was not overstepped during the fracturing and fluidization events. This is well below the critical temperature for pure water (374 °C). At a temperature of, say, 200 °C the vapour pressure of pure water is only 15.5 bars (Fyfe et al. 1978). This would probably be insufficient to break through the granitic overburden at a depth where the rock did not already contain shallow joints more or less open to the surface.

The most probable fracturing and gas fluidization medium at such moderate temperatures is, therefore, considered to have been a fluid composed of carbon dioxide and water, and separated into a H₂O-rich, liquid phase and a CO₂-rich, incondensable fluid phase (Takenouchi & Kennedy 1964). Such mixtures have thermal expansivities, compressibilities and specific heats much higher than those of pure water (Schubert & Straus 1981), and would expand much more than water under the same conditions. The fluidizing agent would then consist of the CO₂-rich gas phase.

Chemically, the CO₂ content of the initial fluid mixture would induce a relatively high solubility for anorthite as compared to albite and K-feldspar (Thompson 1974.) This is consistent with the deposition of Ca minerals upon loss of CO₂, whereas there is no mineralogical evidence that the alkali feldspars have been affected by the fluid. While calcium ion activities could be lowered due to complexing with carbonate species in solution (Wigley & Plummer 1976, Blatt et al. 1980, Anderson 1981); alkali feldspar solubilities, on the other hand, could be reduced by the common-species effect if the fluid were at or close to silica saturation (Fyfe et al. 1978). The sudden escape of CO₂ during fluidization would lead to rapid deposition of epidote/clinozoisite and probably calcite, which would redissolve during falling temperatures as a consequence of its retrograde solubility (Fyfe et al. 1978). Calcium zeolites may in part have formed at the expense of calcite.

The many complexities in this system, such as disequilibrium conditions, alternation between open and closed-system behaviour, and spatial variations in the relative importance of fluid flow versus diffusion transport, discourage further detailed discussions based on the available mineralogical and chemical data for the Skaget Breccia.

With only moderate temperatures prevailing in and near the system during brecciation activity, there is no evidence that the fluid was significantly heated in or by the host rock. On the contrary, the onset of incipient ductility of the granite material at the very contact may indicate that the rock margin was heated by the fluid (Paterson 1978, Handin & Carter 1979).

Therefore, *in situ* thermal expansion does not seem to afford a viable explanation for the dynamics of hydraulic fracturing, explosion and fluidization in

the present case. This would also seem to disqualify circulating groundwater as an immediate source for the fluids; such waters provide only hydrostatic pressures (Henley & McNabb 1978). The fluids must have come from deeper levels.

A deep-lying fluid and energy source within the granite pluton itself would entail the involvement in the brecciation events of a magmatic fluid (Henley & McNabb 1978), equilibrated at high temperatures with crystallizing or hot crystalline granite in the interior of the pluton, below a thick, cool granite shell, and perhaps driven out and upward by retrograde boiling (*ibid.*). However, the chemical effects of the fluid on the brecciated granite do not appear to be compatible with the expected composition of a 'granitic' magmatic fluid (Burnham 1967, Ellis & Mahon 1967, Hibbard 1980). On the other hand, the magmatic fluids could have had an unusual composition; perhaps in part resulting from decomposition of marble inclusions derived from the older gneisses. This is considered less likely, however, because the Frøya Granite appears to be notably free from inclusions of the adjacent gneisses. Moreover, during the late crystallization history of the accessible parts of the Frøya Granite pegmatitic melts, in all probability saturated with volatiles, were retained within the pluton or its migmatitic aureole even at the stage where the solidified granite could sustain large-scale fracturing. This indicates that the Frøya Granite was not emplaced at shallow levels in the crust, where late to post-magmatic fluids would tend to actively force their way upwards. Although there are a number of similarities between the Skaget Breccia and syn-plutonic breccias in granitoid and other igneous rocks, for instance in the British Caledonides (Brindley et al. 1976) and in and near Variscan granites (Goode & Taylor 1980), these plutons seem to have intruded at higher crustal levels than did the Frøya Granite; and thus to have been able to interact, as hot rocks, with water of meteoric origin. The same appears to be the case with breccias associated with intrusive stocks of porphyry copper systems (Phillips 1973, Henley & McNabb 1978). Diatremes and breccia pipes in appinitic plutonic rocks in the Caledonides in Scotland and Ireland differ from the Skaget Breccia in being intimately associated with magmatic products and clearly connected with intrusive processes (e.g. Pitcher & Berger 1972). They are considered to have been produced by volatile escape from water-rich, mantle-derived magma (Wright & Bowes 1979).

The brecciating fluid at Skaget and its energy content may possibly have been independent of and, by implication, younger than the hosting, Frøya Granite.

One alternative derivation for the fluid could have been prograde, devolatilizing metamorphism at deep levels in the crust of metasedimentary rocks associated with deep thrusting. Such a proposition is without supporting evidence at the present time.

A possible mantle origin for the Skaget fluids is an interesting alternative in view of the present, widely held opinion that carbon dioxide is widespread and common in the mantle (e.g., Eggler 1975, Mysen & Boettcher 1975, a, b,

Wyllie 1978, Barnes & McCoy 1979, Wendlandt & Harrison 1979, Newton et al. 1980). Newton et al. (1980) consider that such fluids can enter the overlying crust, and scavenge water from crustal rocks on their way upwards. For carbon dioxide to penetrate into the crust from below would probably require anomalously high temperatures. Appropriate conditions would be expected to occur in the thermal regime of a hot spot or mantle plume.

The heat energy and intrinsic buoyancy of such fluids could probably not alone provide a mechanism for their upwards penetration from the mantle to shallow crustal levels. Under certain conditions, however, release of potential energy could possibly bring this about, as shown theoretically by Secor & Pollard (1975), who applied their model specifically to water at high crustal levels. In this model, the fluid acts as a pressure medium, conveying high static pressures upwards from deeper levels (Fig. 7). In the situation depicted in Fig. 7, a horizontal least-principal rock stress has a vertical gradient (dS_x/dZ) which is moderately higher than the vertical pressure gradient in the fluid (dP_f/dZ). The fluid holds open a long vertical hydraulic fracture by means of the enhanced pressure induced on it by the rock stress at the lower closure of the fracture. If the fluid pressure at the top of the fracture exceeds the sum of the local least principal stress and the tensile strength of the rock, the crack may propagate upwards by hydraulic fracturing; the lower end of the fracture would close, in step with the upward advance. Thus, the length of the fracture will remain constant as long as the fluid ascends under unchanged conditions, assuming that the fluid is incompressible and that the process occurs under

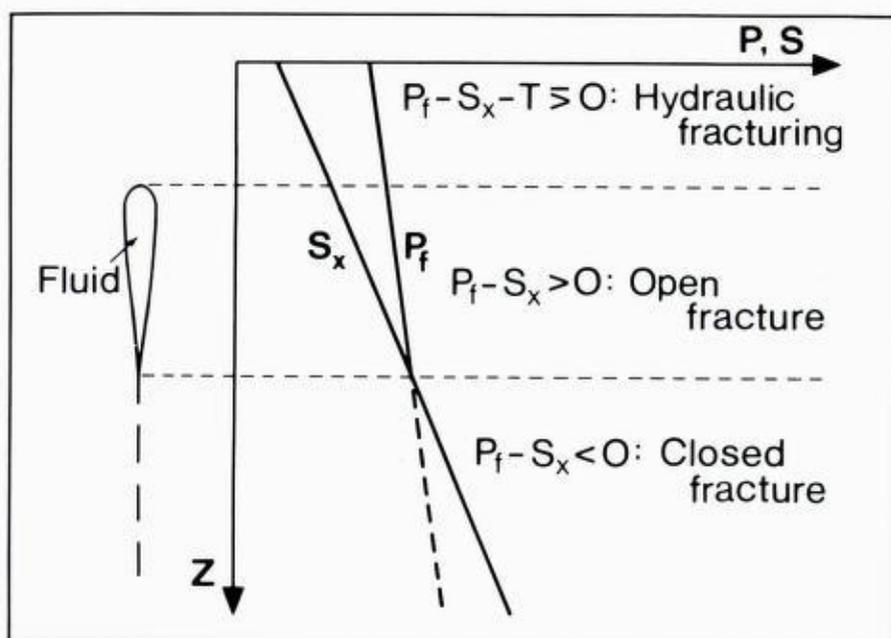


Fig. 7. Depth (Z) vs stress (P, S) diagram showing idealized, plane-strain conditions for upward propagation of vertical, fluid-filled, hydraulic fracture. Adapted from Secor & Pollard (1975). P_f : fluid pressure; S_x : horizontal, least principal stress; T : Tensile strength of rock.

plane-strain conditions (Secor & Pollard 1975). For the present application, these last assumptions make a rather crude approximation. Energy dissipation associated with the work of fracture, etc., would probably halt further ascent of an isolated batch of fluid such as this after a finite distance. However, since the length of the stable, open fracture is determined by the *local* stress and pressure variables and gradients, it should be entirely possible for new batches of fluid to follow in the hydro-fractured wake of the first, and to overtake it. The merging of sequential, open fractures and their contained fluids would probably add renewed impetus to hydraulic fracturing, and so continue a stepwise or pulsed ascent of the fluid. A possible consequence of this latter mechanism could be cyclic brecciation with intervening quiet periods, similar to the processes inferred for the Skaget Breccia, where self-sealing (Keith et al. 1978) by quartz deposition would probably result from hydrothermal activity following each explosive outburst of brecciation and fluidization triggered by the arrival of new pulses of mantle fluid.

High-pressure melting experiments with synthetic and natural mantle material have indicated that melts yielding, for instance, larnite-normative olivine nephelinite can form under realistic conditions from mantle material only in systems with carbon dioxide and water (Mysen & Boettcher 1975 a, b, Egger & Holloway 1977, Burnham 1979). It is therefore particularly pertinent to the present discussion about possible deep sources for carbon dioxide that plugs of larnite-normative olivine nephelinite, dated at about 56 Ma, occur on the continental shelf only about 120 km from Skaget (Fig. 1; Bugge et al. 1980). In this context it is also of interest to note that this part of the coast has been recorded by previous authors as a region of widespread zeolite mineralization on joints and fractures in basement gneisses and other rocks (Fig. 1; Sæther 1950, Strand 1952, Fediuk & Siedlecki 1977). Based on other considerations, a mantle-plume generated uplift has been tentatively proposed for this region (Torske 1975).

Conclusions

Based on the foregoing discussion on the Skaget Breccia and on the occurrence of young(?) zeolite mineralization as well as Tertiary volcanic activity, it is suggested that the coastal and offshore Nordmøre region may constitute a palaeo-geothermal province. This could conceivably have a bearing on the hydrocarbon potential in this part of the continental shelf in terms of palaeo-geotherms and permeability effects of mineralizations.

A critical topic in the assessment of the possible genetic affiliations of the Skaget Breccia would be its mineralization age. This is unknown at the present time.

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