Kvithola at Fauske, northern Norway: an example of ice-contact speleogenesis

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Kvithola is a system of vertical, relict phreatic karst conduits. Passage cross-sections are elliptical or lenticular, averaging 1.5 m^2 in area. The cave extends parallel with, and close to the wall of a formerly glaciated valley. Water was transmitted downwards at a rate of $1m^3s^{-1}$, and the majority of passages were developed from sheet-fractures close to the surface. The corresponding paleohydraulic gradient was more than two orders of magnitude less than the surface slope. The only conceivable mode of phreatic flow is then by ice contact, of which the ice-marginal case seems the most probable with regard to CO_2 content and hydrology of glacial environments.

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The classical controversy around speleogenesis of relict, phreatic karst caves, hanging high in glacially sculptured topographies, has been to what extent the formation can be ascribed to preglacial (Tertiary), interglacial, subglacial or proglacial conditions. This particular configuration of cave morphology and surface topography is very common in Norway (Horn 1947, Lauritzen 1981). Down through the years, proglacial (Oxaal 1914, Corbel 1957) and subglacial (Horn 1935, 1947) speleogenesis have been suggested. As with most controversies, each view may have some merit. Many caves are also complex in morphology and probably polygenetic in origin (Lauritzen 1983).

Background problems: sub-glacial speleogenesis

Horn's general subglacial speleogenetic hypothesis consists of two main arguments (Horn 1947). First, the thermal and hydraulic conditions prevailing beneath large glaciers (Werenskiold 1922) allow water to circulate from the glacier base into the bedrock and therefore enhance the possibility of subglacial speleogenesis. Horn's second argument was that pre- or interglacial speleogenesis could be excluded from geomorphic considerations. Glacial erosion rates were considered to be so high that – even in interglacial times – the cave systems of today would have been deeply buried beneath bedrock ($\leq$500 m). At such depths, groundwater circulation was regarded as insignificant, and with no water circulation, speleogenesis could not occur. Horn emphasized that this argument was stronger for preglacial than for interglacial cases.

Recent research has revealed evidence that makes the second argument less valid than believed at the time. First, Uranium Series disequilibrium dating of calcite speleothems (cave dripstones) has proved that many major cave systems in glaciated areas around the world are several glacial cycles old (Schwarcz et al. 1982). This is also the case for Norway (Lauritzen & Gascoyne 1980, Lauritzen 1983, 1984a). Speleothem dates provide a minimum age for the cave. In the case of phreatic (sub-watertable) conduits, the cave must first be drained above the watertable before speleothem deposition may occur. In particular, two of the major cave systems discussed by Horn (1947), Hamarnesgrotta and Lapphullet, have been dated to > 350 ka and probably > 720 ka, respectively (Lauritzen 1984a, Lauritzen, Løvlie & Moe, in prep.)

Second, glacial erosion rates in valleys may be estimated by speleothem dates, giving averages through one or more glacial/interglacial cycles. These rates are quite low, often less than 0.13 m ka^{-1} (Ford et al. 1981). In Norway this rate has been determined in one case to 0.35 m ka^{-1} or less (Lauritzen & Gascoyne 1980). This value is in good accordance with the erosion rate of a Younger Dryas cirque glacier, derived from moraine volume estimates (0.50–0.60 m ka^{-1}; Larsen & Mangerud 1981 and references therein). Many of the caves discussed by Horn (1947) are situated several hundred metres above valley floors, and have therefore been above the bedrock-controlled base level of erosion through several glacial cycles.

The high minimum ages demonstrate that subglacial karst corrosion was unable to create these cave systems either during the last, or during any of the previous glacial cycles through the upper Pleistocene. Rather, the likelihood is increasing that these caves represent fragments of extensive groundwater drains which were mainly developed in early interglacials, or even in Tertiary, and later truncated by glacial erosion. Moreover, the dissolution of limestone lacks intrinsic thresholds (Ford 1980), and karstification must have taken place over large time spans since the Tertiary uplift of Norway. These time spans (40–60 ma) are much larger than those of the glaciations, and we should expect large karstforms (towers, poljes, large conduits) to characterise preglacial karst. Hence, karst megaforms at or within paleic surfaces may well be interpreted as surviving preglacial forms (Lauritzen 1986c). Such arguments are strictly valid only for the cave cases investigated, or for cases that are geomorphologically closely similar to them. Hence, examples of cave conduits which have undergone solutional modifications (paragenesis) under subglacial conditions do occur as well (Lauritzen 1983b).

In summary, many caves in Norway are very old compared to what was previously believed. Subglacial modification of already established cave conduits has been demonstrated. Subglacial speleogenesis sensu stricto (i.e. the complete formation of a cave conduit from preexisting secondary porosity within the rock) is still neither proven nor rejected. To penetrate this problem we have to consider the limiting conditions for speleogenesis.

Speleogenetic time constraints from guiding fractures

Cave passages develop from pre-existing voids. Within the cave-bearing marbles in Norway, these void types are mainly restricted to joints and faults. The cave conduit cannot be older than the joint it developed from (guiding fracture). Some fracture zones in Norway are recently reactivated (Bungum & Husebye 1979), but the fracture systems in Norway as such are very old (Gabrielsen & Ramberg 1979, Gabrielsen et al. 1981). Therefore, the possible age of the fractures, which confine the age of the cave, may often be several orders of magnitude higher than that of the glaciations. One way to approach this problem is to search for a cave developed from geologically young fractures. This would then put a confining age of relevant magnitude on the speleogenesis.

This is far from an easy task. Many caves are parts of extensive systems of great complexity, showing evidence of multi-phase development, sometimes with glacial truncation and being partially or completely filled with exotic sediments and local breakdown. The passages are locally guided by tectonic features which may exist within a very complicated structural setting, as is common within the Nordland nappes. Such complex situations involve many additional parameters influencing cave development which are poorly known, and these in turn make it difficult to draw definite conclusions about speleogenesis. Generally, caves which approach the necessary requirements of simplicity are much rarer than complex systems with multiple phases of development under different climatic and geomorphic conditions.

During the last 10 years, the author has visited about one third of the ~800 karst caves known in Norway. So far, only a few seem to display sufficient simplicity to allow unique distinctions with respect to speleogenesis. Kvithola at Fauske, northern Norway, is one example of a simple cave system which involves relatively few uncertain variables and which also appears to have formed during a former glaciation. This paper is an attempt to present a concise discussion of Kvithola as an example of ice-contact speleogenesis.

Geomorphological and geological setting

The cave is situated within the southern wall of the valley forming the Øvrevatn fjord-lake at Fauske (Fig. 1). The fjord posesses the same morphological characteristics as other glacial fjords in Norway; a U-shaped cross-sectional profile, incised into a 'paleic' plateau (Gjessing 1967), 3–500 m a.s.l. The cavernous carbonates (Fauske marble group, Nicholson 1973) consist of a marble belt, measuring ~15 by ~55 kilometres

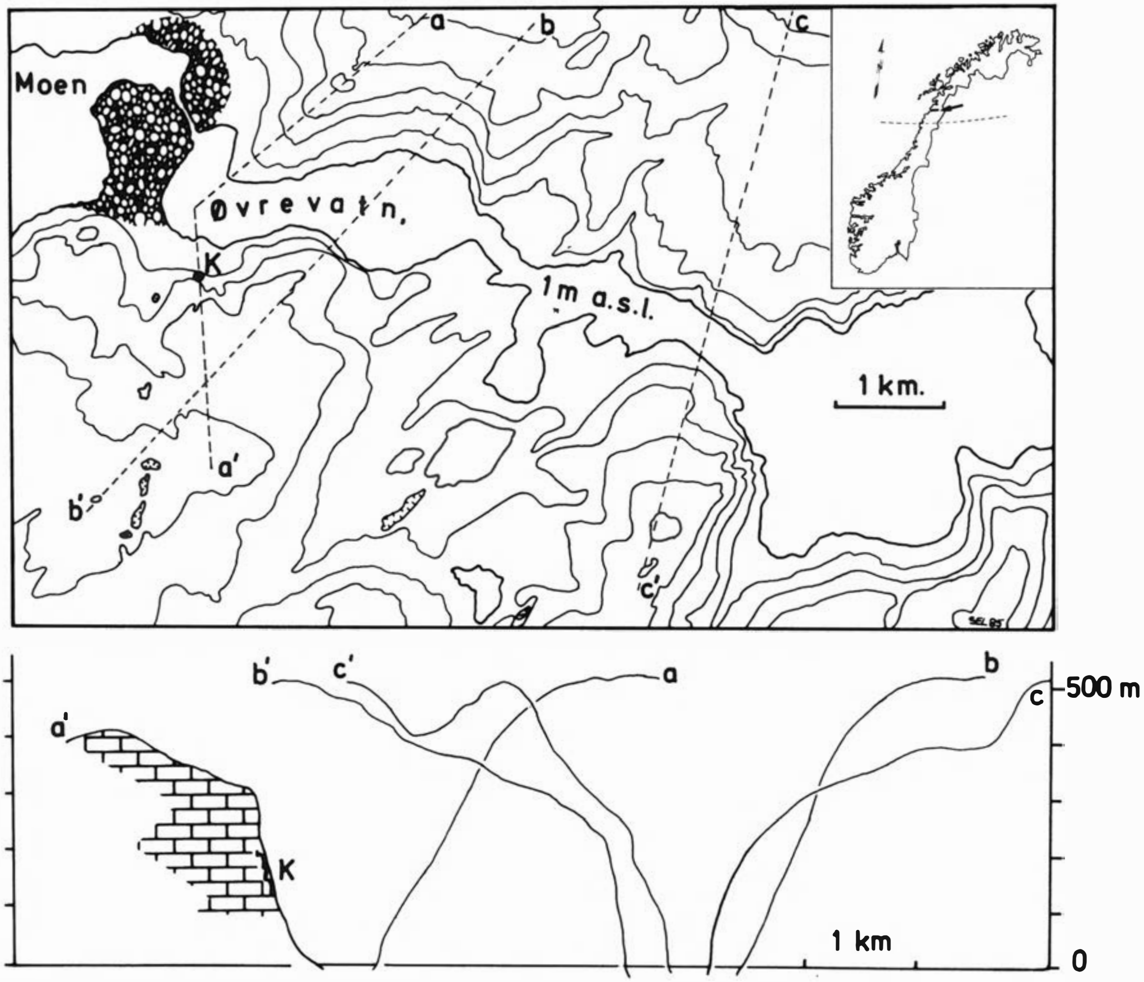

Fig. 1. Above: The fjord-lake Øvrevatn, between Fauske and Sulitjelma, Northern Norway. The Øvrevatn moraine (correlated to a Preboreal event, Andersen 1975) is shown on the left. The location of Kvithola (K) is indicated. Karstic macroforms are shown as closed depressions on the paleic plateau. Below: Cross-sections of the glacial incision, looking westwards.

in areal extent. Several of the largest known caves in Norway are situated within this formation. The local strike of foliation planes is N-S with a steep westward dip (60–70°).

Cave description and paleohydraulic analysis

The present (lower) entrance to the cave is situated at 124 m a.s.l., and it has been explored upwards to at least 200 m a.s.l. (H. M. Herstad, pers. comm. 1984). The entrance shows evidence of truncation, either from glaciers, or from rockfalls guided by exfoliation. The present survey (Fig. 2) covers most of the known passages, with the exception of tubes extending about 30 m beyond the top syphon, and minor narrow fissures. Exploration confirms that these passages all follow the same trends as those in the survey. The conduits have a very steep gradient, closely following the valley wall, less than 10 m from the surface (Fig. 2).

The passages are tubular, with an elliptical or lenticular cross-section, providing evidence of a phreatic (sub-watertable) formation. Moreover, scallops (dissolution-bedforms on the walls) invariably demonstrate that the conduits conveyed water downwards from the plateau edge, and out through the lower entrance. The presence of scal-

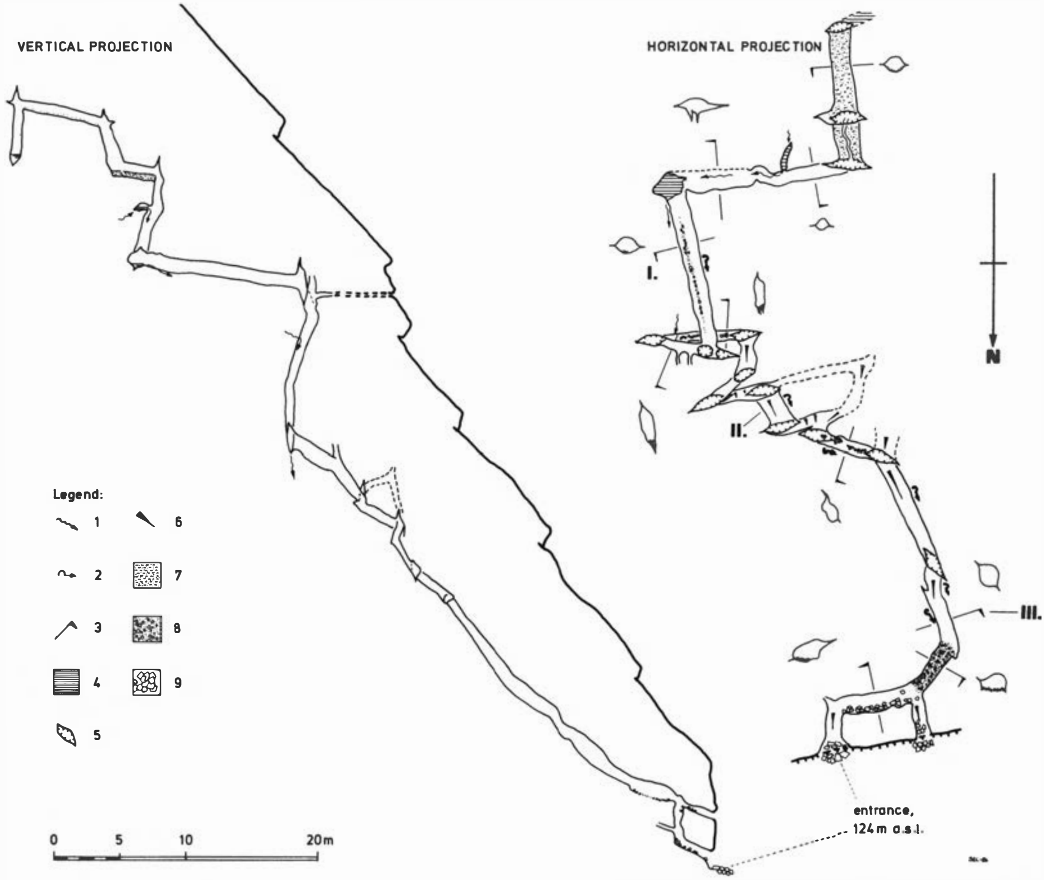

Fig. 2. Plan and vertical section of Kvithola. Vertical projection plane approximately N-S. On the horizontal projection, passage cross-sections, scallop directions and stations for paleocurrent calculations (I–III) are shown. The total surveyed length is 130 m, depth is 57 m. The cave survey accuracy standard is grade 5B of the British Cave Research Association (Ellis 1976). Legend: 1: small stream, 2: observed scallop direction, 3: site and direction of cross section, 4: water surface, 5: vertical shaft, 6: slope of passage floor, 7: sandy silt, 8: fluvial gravel and sand, 9: angular breakdown.

lops on the roof demonstrates that the cave was full of water. The dimensions of the scallops and the passage cross-section were measured at three points, in the upper, middle and lower parts of the cave (sites I–III, Fig. 2). The paleodischarge in equilibrium with the scallops was calculated using the procedures outlined by Lauritzen (1982), using equations of Blumberg & Curl (1974). The results are shown in Table 1.

The different sections of passage transmitted between 0.38 and 1.00 m^3s^{-1}, respectively (Table 1).The deviations from continuity of flow along the conduits (i.e. section II) may be explained by the numerous oxbows (branched loops), which

Table 1. Flow rate calculations from scallops in Kvithola (+1°C). x and y are passage diameters, assuming an elliptical cross-sectional profile. $L_{32} = \Sigma L_i^3/\Sigma L_i^2$ (Sauter-mean of scallop lengths), n = number of scallops measured.

Station	x (m)	y (m)	L_{32} +/− σ_{32}	n	ū (cm s^{-1})	Q (m^3 s^{-1})
I	1.50	1.15	8.14 $^{+0.98}_{-0.87}$	47	49.5	0.67 $^{+0.10}_{-0.08}$
II	0.70	1.30	7.10 $^{+0.81}_{-0.73}$	44	53.8	0.38 $^{+0.06}_{-0.05}$
III	1.50	1.30	6.59 $^{+0.95}_{-0.83}$	50	65.1	1.00 $^{+0.18}_{-0.15}$

could not be reached for surveying or for scallop morphometry (Fig. 2). However, the section close to the lower entrance (section III) would collect the discharges from most of the known passages upstream. Hence, 1.0 m^3s^{-1} is taken as representative for the scallop discharge through the cave. In the timescale of the total history of a cave conduit, scallops represent the last discharges involving corrosive water. In situ process studies indicate that scallops belong to the upper part of the flow regimes, i.e. they represent the highest stages of flow (Lauritzen et al. 1983, 1986).

The slope of the piezometric surface may be calculated from the D'Arcy-Weissbach equation for turbulent flow (Smith et al. 1976):

$$[\Delta h/\Delta 1] = Q^2 f P_w / 8a^3 g \qquad (1)$$

where $[\Delta h/\Delta 1]$ = hydraulic gradient, Q = discharge (m^3s^{-1}), f = a dimensionless friction factor, a = the cross-sectional area of the conduit (m^2), P_w = wetted perimeter (m) and g = acceleration of gravity (ms^{-2}). 'f' may be calculated from scallop lengths and passage radius (Gale 1984, Lauritzen et al. 1986), and from observation in active conduits, either by direct hydrodynamic (Lauritzen et al. 1986) or aerodynamic analysis (Atkinson et al. 1983). There appear to be large variations between apparent 'f' s based on active systems, and the 'f''s derived directly from wall roughness. $[\Delta h/\Delta 1]$ was calculated for the various places where scallops were measured, using the different friction factors available (Table 2).

Fig. 3. Photograph of the lower parts of Kvithola, looking upstream. At this place, the passage follows the line of intersection between foliation and sheet-fracturing. Note the scalloped surfaces and the symmetrical development around the guiding fracture, indicating a sub-watertable (phreatic) development. Protruding micaeous bands indicate that the effect of abration was minor.

Table 2. Hydraulic gradients for scallop discharges in Kvithola. 1: 'f' measured from scallop roughness, this work. 2: Apparent 'f' determined in situ from actual hydraulic gradient (Lauritzen et al. 1986) 3: Apparent 'f' determined from wall roughness and from airflow (Atkinson et al. 1983).

D'Arcy-Weisbach 'f'	Hydraulic gradient at stations			Ref.
	I	II	III	
0.06–0.07	6–7 E–4	9–11 E–4	9–11 E–4	1
0.016	2 E–4	3 E–4	2 E–4	2
0.3–0.9	3–8 E–3	4–14 E–3	4–14 E–3	3
1.5–2.3	1–2 E–2	2–4 E–2	2–4 E–2	3

Calculations based on wall roughness alone do not take bends or constrictions into consideration. Bends and constrictions will increase the resistance to flow, and hence the hydraulic gradient that is necessary to maintain the flow through the cave. The values of Atkinson et al. (1983) were based on a cave passage which was much longer and had a much higher frequency of constrictions than Kvithola. This would make the predicted $\Delta h/\Delta 1$ of ~0.04 to a maximum value, which is almost two orders of magnitude less than the present-day surface slope, or gradient of the cave conduits themselves (~70° i.e. $[\Delta h/\Delta 1]$ = 2.7).

If bedload or suspended sediments could abrade the walls and floor of the cave, the rate of passage growth would have been considerably higher than if corrosion acted alone. Such an effect was, for instance, suggested for the Grønli cave system by Oxaal (1914). The paleocurrent regime inferred from scallops ($\bar{u}$ = 50 cm s^{-1}) could have kept sand-sized sediments in suspension and possibly added some abrasion to the development. However, no *true abrasion forms* (i.e. rock-mills), which could prove a dominant abrasional effect from bedload exist within the cave. Moreover, fragile schist flakes protrude into the passage in places, suggesting that abrasion was of minor importance. Rather, the conduits possess characteristics of a dominantly soluble origin, formed through the corrosional action of groundwater on the marble. This corrosion was symmetrical around the guiding fracture (Fig. 3).

Guiding fractures

More than 80% of the guiding fractures (i.e. the joints or voids from which the conduits devel-

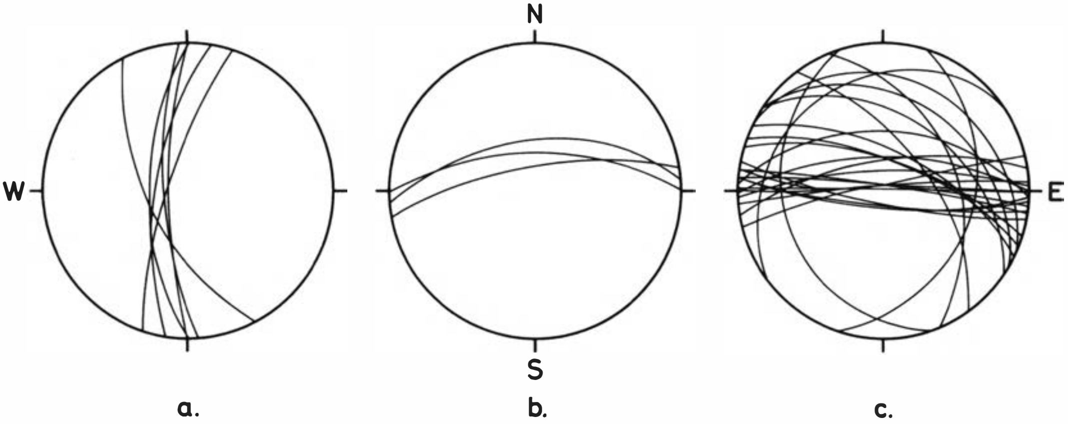

Fig. 4. Stereographic (Wulff) projections of structural elements in Kvithola.Lower hemisphere. a) Foliation planes of the marble, as observed within the cave. b) Sheeting fractures on the surface (valley side). c) Guiding fractures of the passages in Kvithola. More than 80% of the guiding fractures are parallel with the valley side, and situated less than 10 m from the surface.

oped through solutional enlargement) are parallel with the fjord wall and with sheeting fractures on the surface (Fig. 4). This fracture-set deviates from the larger scale sets of joints and photolineaments as seen on aerial photographs or satellite imagery (Gabrielsen et al. 1981), which have a NNE-SSW trending strike. As Kvithola runs very shallowly beneath the hillside, it is most likely that the sheeting as seen on the surface and the cave guiding joints belong to the same set. This means that the cave as a system of conduits developed from pre-existing sheet-fractures. Sheet-fractures are formed by release of stress from overburden as a response to denudation (Jahns 1943, Harland 1956). The joints are closely related to the landform and could not have been formed or at least opened for hydraulic conductivity unless the valley profile had already been established. Hence, the local base-level of erosion (bedrock of the valley floor) was below the level of the cave when it was formed.

Conditions for phreatic development

Passage cross-sections and scallops demonstrate that the cave is totally phreatic and conveyed water *downwards* along the valley wall. The paleocurrent discharge indicates that it functioned under a piezometric surface with a gradient that was considerably less than the surface slope. This piezometric surface was situated above the explored parts of the cave, at least 200 m a.s.l. Since the cave is formed from sheet-fractures which are younger than the valley itself, these hydraulic conditions could only have occured when the valley was filled with ice, either in a deep subglacial situation, or within a proglacial (ice-marginal) situation. It is, however, more difficult to be certain whether the cave formed as a result of deep subglacial (Ford 1977) or ice-marginal speleogenesis. The latter corresponds to the case first outlined by Oxaal (1914), and the former to the mechanism suggested by Horn (1935).

Glacio- and geohydrologic interaction

Subglacial conduits appear as two principally different types (Paterson 1981). N- (Nye) channels are formed along natural channels in the bedrock on the glacier sole, where only the roof and/or one of the walls consist of ice. R- (Røthlisberger) channels are formed entirely of ice and may therefore also exist englacially. N- channels are the much more stable of the two. Because of the plasticity of the ice, R- channels are prone to close in periods of decreasing discharge or to shift along the glacier bed as the ice mass moves. A cave conduit in bedrock adjacent to the glacier bed can be viewed as an N-channel of extreme stability (Lauritzen 1983, Smart 1984). Kvithola would have been just such an effective subglacial channel (Fig. 5).

The genetic relationship between topography, guiding fractures, cave morphology and paleo-

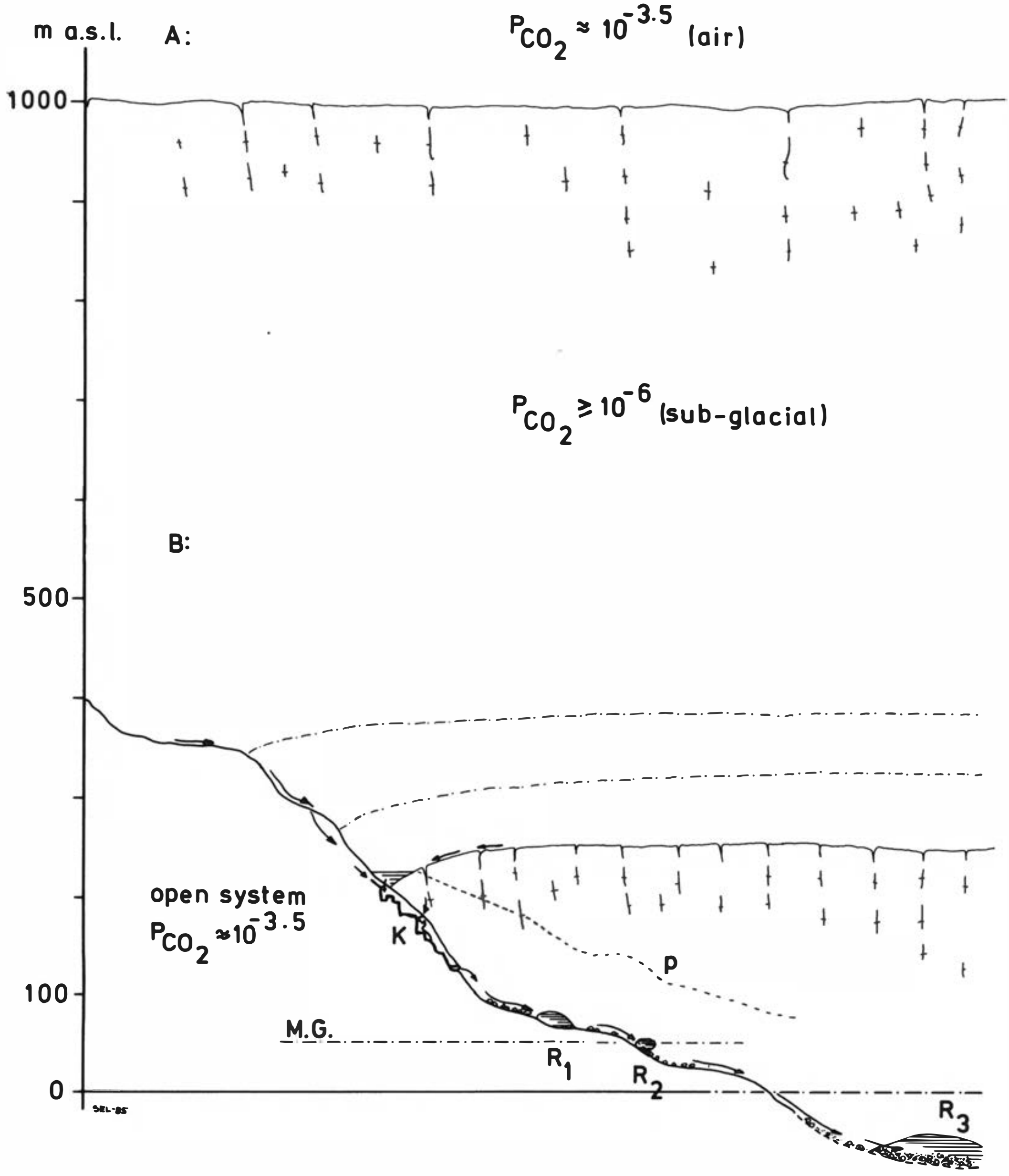

Fig. 5. Ice-contact speleogenesis of Kvithola, showing two alternative situations of icecover:
A. (Upper). Deep subglacial condition, ice surface arbitrarily chosen at 1000 m a.s.l. Under a thick ice cover, supraglacial water will have very limited access to the glacier sole. Any liquid water will originate from basal or englacial melt. Ice has a low CO_2 content and, combined with stagnant hydraulic conditions, will make corrosion very sluggish.
B. (Lower). Ice-marginal situation. Supraglacial runoff with an open system pCO_2 of high corrosion capacity may enter the cave (a very stable N-channel), draining towards R-channels (R_{1-3}) at the glacier base. Hydraulic gradients are steep enough to provide a high steady-state corrosion rate. M. G.: Late Weichselian upper marine limit; p: hypothetical piezometric surface; K: Kvithola. For comparison, Pco_2's are given at one atmosphere total pressure.

hydraulics demonstrates that speleogenesis took place below an englacially or proglacially controlled piezometric surface.

Conditions for passage growth

The rate of passage enlargement is a function of the acidity, or CO_2 content of the water, and the rate of flow. High CO_2 concentrations originate from biological activity (respiration) in soil; this may raise the Pco_2 of the water one or two orders of magnitude above that of air ($Pco_2 = 10^{-3}$atm). A proglacial environment will have a very low biotic activity, and, like high alpine environments, the dominant CO_2 source will be atmospheric air.

A high rate of flow may maintain a highly effective (steady state) rate of solution, because fresh water will always be in contact with the rock surfaces. The average flow velocity in the passages ($\bar{u} = 0.5$ m s^{-1}, Table 1), suggests that the time of residence of the conduit water within the entire hillside (300 m) would be less than 10 min. This is a very short time related to the time required for karst conduit water to attain equilibrium and hence an insignificant rate of dissolution (several days). Thus, we can expect that corrosion rates would have been similar in all parts of the Kvithola cave.

The CO_2 concentration of the deep subglacial environment is distinctly different from supra- and englacial sites which form an open system to the atmospheric CO_2 reservoir. Deep subglacial waters are depleted with respect to CO_2, and combined with a closed system condition, they have a very limited solutional capacity (Ford 1971, Ek 1974). Water of the deep subglacial environment, originating from basal or englacial melt, is often more than 1 order of magnitude less in total CO_2 content than surficial glacial waters. The corresponding reaction rates and effective penetration distances under closed system conditions and high ambient pressure are very unfavourable for speleogenesis sensu stricto (Lauritzen 1986a). In addition, large ice-sheets with a low surface gradient would minimize local subglacial hydraulic gradients and cause stagnant conditions (Ford 1977). Stagnant conditions would inhibit the dissolution rate, as waters become saturated during long contact with marble and were not replaced by fresh flows, as well as causing prolific deposition of silt and clay-sized sediments. The combination of these factors would slow down the rate of cave enlargement considerably.

The maximum (steady-state) rate of solution under open system conditions may theoretically approach 10^{-7} mMoles cm^{-2} sec^{-1} at $Pco_2 = 10^{-3.5}$ atm and t = 0 °C. This corresponds to a wall retreat rate of about 1 mm a^{-1} (Lauritzen 1986a) However, direct measurements of wall retreat rates in active caves of to-day do suggest about an order of magnitude less (i.e. 0.1 mm a^{-1}, Lauritzen 1986b). Assuming that the flow rates and hydraulic gradients represented by the scallops may also apply to earlier stages in the growth history of the passages, we may estimate the probable growth duration time for the cave. The conduits of Kvithola would then have gained much of their size in about 750–7500 years of continuous flow. The real time that would be required for passage growth would also depend on the contribution from abrasion as discussed above. In terms of glacial advances or retreats (Fig. 5B), the optimistic of these two estimates suggests that only a few of these sequences might have been sufficient to create the cave.

Conclusion

1. Kvithola is so far the only known example of a hanging, relict phreatic tubular cave which can be conclusively demonstrated as of subglacial origin (i.e. deep subglacial or ice marginal), and for which other genetic modes (i.e. pre- or interglacial) may be excluded (Fig. 5).

2. It is proposed that Kvithola was formed at the margin of, but well below the surface of a local glacier. This glacier filled the Fauske-Øvrevatn fjord completely, i.e. *at least* up to 200 m a.s.l. This configuration also provides a favourable glaciohydraulic situation, where the cave formed an extremely stable N-channel, injecting water towards R-channels at the base of the glacier, which in turn conveyed meltwater towards the sea (Fig. 5B). The situation is compatible with Younger Dryas and Preboreal ice extents, but it is uncertain whether one or more such events were necessary to create the cave.

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