Post-Caledonian thermal evolution and crustal uplift in the Eidfjord area, western Norway

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Andriessen, P. A. M. & Bos, A.: Post-Caledonian thermal evolution and crustal uplift in the Eidfjord area, western Norway. *Norsk Geologisk Tidsskrift*, Vol. 66, pp. 243–250. Oslo 1986. ISSN 0029-196X.

Apatite and zircon fission-track ages and biotite Rb-Sr and K-Ar ages in the Eidfjord crustal segment (the Eidfjord Granite and overthrust granitic gneisses) indicate a rapid cooling, from 400°C to 200°C, between 390 and 306 Ma ago. Then the cooling rate slowed considerably: it took until 166 Ma ago before the crustal level, which is now situated at an altitude of 1620 m, had cooled further to 105°C, while the crustal level which lies now at sea level had cooled to this temperature about 10 Ma ago. Assuming a geothermal gradient of 30°C/km, a mean uplift rate of 0.1 mm/a shortly after the Caledonian orogenesis is calculated for the time-span from about 390 to 306 Ma ago. This rapid uplift is stratigraphically reflected in the deposition of the Old Red Sandstone deposits. For the last 300 Ma or so, until the present, the uplift rate was much slower: of the order of 0.02 mm/a. From the mineral age pattern, it is calculated that since about 390 Ma ago, shortly after the termination of Caledonian orogenesis, over 13,000 metres of crustal material have been removed by erosion and denudation, of which some 5000 m within the first 80 Ma or so.

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The Eidfjord area on the western edge of Hardangervidda is located in the Caledonian belt of western Norway (Andresen & Faerseth 1982). The area is underlain by granitic gneisses and gneissose granites, intruded by a granitic mass designated as the Eidfjord Granite. To the east of Eidfjord, at Hardangerjøkulen, the topographically highest part of the region, the Cambro-Ordovician cover of the Precambrian basement is tectonically overlain by granitic gneisses (Barkey 1965) variously attributed to the Hardangervidda-Ryfylke Nappe System (Andresen et al. 1974) or to the Jotun Nappe (Andresen & Faerseth 1982).

A Rb-Sr whole-rock investigation of a suite of samples from the Eidfjord Granite by Priem et al. (1976) yielded an isochron age of 911 ± 35 Ma (recalculated with the ^{87}Rb decay constant of $1.42.10^{-11}a^{-1}$). A similar U-Pb age of 905^{+86}_{-47} Ma is indicated by the upper intercept of the discordia through a suite of zircons (N. A. I. M. Boelrijk, private communication). It is thus clear that the intrusion of the Eidfjord Granite must be related to the Sveconorwegian orogenic episode. For two biotites from the Eidfjord Granite, Priem et al. (1976) reported Rb-Sr ages of 381 Ma and 386 Ma (recalculated with the above ^{87}Rb constant) and K-Ar ages of 410 Ma and 436 Ma. The latter age was interpreted as being too high, so the age of the biotite can be taken as approximately 390 Ma; this age is related to the closure of the biotite Rb-Sr and K-Ar systems following the termination of the Caledonian metamorphism.

Similar ages have been reported for the Hestbrepiggan Granite in the Jotunheimen area, north of Eidfjord. Here the Rb-Sr whole-rock data point to an age of 942 ± 35 Ma, while biotite reveals Rb-Sr and K-Ar ages of 381 Ma and 396 Ma, respectively (Priem et al. 1973; Rb-Sr ages recalculated with the above ^{87}Rb constant).

It has been postulated that the Caledonian orogenesis in south-west Norway took place in two phases. An early phase of tectonism and igneous activity between 550 and 500 Ma which is related to the Finnmarkian phase in northern Norway and the Grampian event in the Scottish Caledonides (Roberts & Gale 1978; Sturt et al. 1978; Andresen & Faerseth 1982) should have been followed by a period of uplift and erosion (Faerseth 1982). Then, in the Middle and Late Silurian, about 410 Ma ago, a second phase of Caledonian deformation should have taken place, the Scandian event (Roberts & Sturt 1980; Roddick & Jorde 1981). Petrological and mineralogical in-

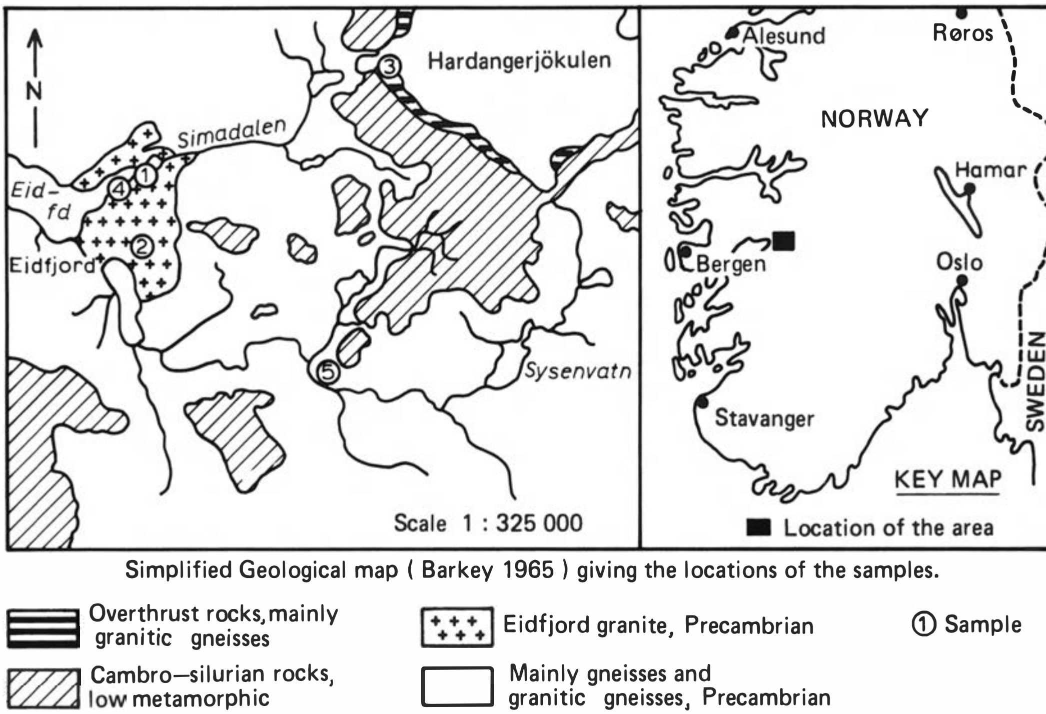

Fig. 1. Geological sketch map of the Eidfjord area (after Barkey 1965), showing the locations of the investigated samples.

vestigations of basement rocks in the Rogaland/Vest-Agder area in south-west Norway also provided convincing evidence for a polyphase Caledonian metamorphism (Sauter et al. 1983). Following this concept, the biotite ages of the Eidfjord and Hestbrepiggan Granites have to be related to the Scandian event, representing cooling ages after the metamorphism.

In this study the post-metamorphic thermal history of the Eidfjord crustal segment is reconstructed on the basis of fission-track ages of apatites and zircons from four samples, two from the Eidfjord Granite, one from a gneiss and one from a granitic gneiss. The locations are shown in Fig. 1. The granite samples, Eid 1 and Eid 4, were taken at sea level (the biotite ages reported by Priem et al. 1976 were obtained from the same samples), while Eid 5, a gneiss from the basement, comes from an elevation of 700 m. The granitic gneiss Eid 3 is a sample from the top of Hardangerjøkulen, with an altitude of 1620 m. This gneiss belongs to a slice of the Precambrian basement, forming part of the overthrust nappe system that is tectonically separated from the underlying Eidfjord Granite. Petrological and structural investigations reveal, however, that these rocks have also been affected by the Caledonian deformation and metamorphism, and that the overthrusting of the nappe took place during the first phase of Caledonian orogenesis, or even earlier. This implies that during the Scandian event the nappe was already in situ (Andresen & Faerseth 1982). The Eidfjord Granite and the overthrust granitic gneiss can thus be conceived as forming part of the same crustal segment.

Petrography and petrology

In all the samples the mineral stilpnomelane occurs, but in none of them have the minerals prehnite or pumpellyite been detected. The biotite is mostly chloritised and shows inclusions of sphene, epidote and colourless mica. The colour of the biotite is green to yellowgreen for Eid 2, Eid 3 and Eid 4, and yellow-grown for Eid 1 and Eid 5. It seems therefore that the metamorphic grade left the prehnite-pumpellyite facies and entered the greenschist facies. It is rather difficult to establish the metamorphic conditions, but some similarities with those in the vicinity of the Caledonian front is SW Norway (Verschure et al. 1980; Sauter et al. 1983) seem obvious. The occurrence of stilpnomelane in the basement is largely restricted to the vicinity of the Caledonian orogenic belt (Sauter et al. 1983), where biotites with a Caledonian age have been reported. The disappearance, or non-occurrence in this case, of prehnite and pumpellyite points to the beginning of the greenschist facies, where temperatures of 350–400°C under moderate load pressures prevailed. The upper stability of stilphomelane indicates, according to experimental data of Nitsch (1970), to a temperature of 445°C ± 10°C and 4 Kb. It seems therefore justified to estimate the metamorphic conditions at around 400°C and moderate pressures during the Caledonian orogeny.

Analytical procedures, constants and calibration

The fission-track analytical procedures are according to Gleadow et al. (1976) and Naeser (1978). Apatite is bedded in epoxy and zircon in FEP-teflon. The external detector method is used for zircon, the population method for apatite. Prior to the irradiation, zircon mounts are etched in a eutectic KOH-NaOH melt at 220°C for about 3 hours. Only a limited number of crystals with a low-U content and showing sharp polishing scratches were suitable for fission-track counting. The apatite concentrate is split into two fractions; the spontaneous fission tracks of one fraction are annealed in an oven at 450°C for 10 hours. Subsequent to the irradiation, the two apatite fractions are mounted and polished and then etched simultaneously in 7% HNO_3 for 30–40 seconds at 20–25°C. Muscovite detectors are etched in 48% HF for 11–15 minutes at 20–25°C.

Recently several articles have been published concerning age calibration of the fission-track dating method (Hurford & Gleadow 1977, Hurford & Green 1981a, 1981b, 1982, 1983). Minerals with a well-known age, like the Fish Canyon zircon (Naeser et al. 1981), are used for the determination of a so-called Zeta (ζ) factor which circumvents the absolute evaluation of neutron flux and decay constant of spontaneous fission of ^{238}U. To obtain the empirical calibration factor for a particular dosimeter glass, one must repeatedly calibrate the dosimeter glass against an age standard. The Zeta value is calculated as follows (Hurford & Green 1983):

$$\zeta = \frac{1}{\lambda_D} \frac{(e^{\lambda_D T} - 1)}{\frac{\varrho_s}{\varrho_i} g \, \varrho_d}$$

Where λ_D = total decay constant of ^{238}U, 1.55125 $10^{-10}a^{-1}$; T = age of the Fish Canyon tuff zircon, 27.9 ± 0.7 Ma (Steven et al. 1981); g = the geometry factor, 0,5; ϱ_d = fission-track density of the dosimeter glass; ϱ_s = density of the spontaneous fission tracks in the Fish Canyon zircon; ϱ_i = density of the induced fission tracks of the irradiated Fish Canyon zircon. With the Zeta value the age of the unknown sample can be calculated as follows:

$$T_{unk} = \frac{1}{\lambda_D} \ln \left(1 + \lambda_D Z \frac{\varrho_s}{\varrho_i} \varrho_d\right)$$

Table 1. Fish Canyon Tuff zircon: T = 27.9 Ma

Number of crystals	N_s[1]	ϱ_s[2] (10^6 t cm^{-2})	N_i[1]	ϱ_i[2] (10^6 t cm^{-2})	N_s/N_i	ζ_{962}	$\varrho_{d(962)}$[2] (10^5 t cm^{-2})
6	881	3.77	965	8.26	0.456	288.2	2.125
6	901	3.58	1626	12.94	0.276	303.0	3.335
6	565	4.18	982	14.5	0.288	290.8	3.335
6	1049	3.99	1003	7.60	0.525	295.3	1.804
6	929	4.35	1034	9.68	0.449	332.8	1.870
6	995	4.49	1826'	16.5	0.272	309.9	3.315
6	970	5.41	975	10.9	0.496	294.5	1.913
6	868	4.32	1212	12.1	0.317	338.0	2.611
6	695	4.17	546	6.55	0.636	320.5	1.371

[1] N_s and N_i: number of spontaneous and induced fission tracks, respectively

[2] ϱ_s and ϱ_i: density of spontaneous and induced fission tracks; ϱ_d: density of the induced fission tracks of the dosimeter glass of NBS_{962}.

Where λ_D = total decay constant of ^{238}U; Z = calibration factor for a particular dosimeter glass against known age standards; ϱ_d = fission-track density of the dosimeter glass; ϱ_s = spontaneous fission-track density of the mineral; ϱ_i = induced fission-track density of the mineral.

Table 1 shows the data of nine irradiated Fish Canyon zircon mounts. From these results, a mean Zeta value of 308.1 is calculated for the dosimeter glass 962 of NBS. Using the combination of $\lambda_F = 7.03\ 10^{-17}a^{-1}$ and the neutron flux determination via NBS glass 962 and the copper calibration, the repeated Fish Canyon zircon age determination yields a mean age of 28.8 ± 2.5 Ma (95% C.L.). The widely accepted mean age of this tuff formation is 27.9 ± 0.7 Ma (Steven et al. 1976). The agreement between the recommended value and the determined one supports the choice of the above-mentioned combination of λ_F and neutron flux determination. The ages of the Eid zircons have been calculated according to the Zeta factor approach, whereas the ages of the Eid apatites are calculated with a λ_F value of 7.03 $10^{-17}a^{-1}$ and the neutron flux determination via NBS glass 963 and the copper calibration. The errors for the ages are calculated according to a Poisson distribution of the fission tracks.

Results and discussion

Zircon fission-track ages

The fission-track data of the zircons from the granites Eid 1 and 4 and the granitic gneiss Eid 3 are listed in Table 2, along with the calculated ages. Within the error limits all ages are concordant between 294 and 316 Ma, with a mean age of 306 ± 22 Ma (2σ).

Fission-track ages record the last time the rock cooled through the annealing temperature of the dated mineral. For zircon this temperature is commonly taken between 175° and 225°C (Harrison et al. 1980). This implies that the crustal segment under discussion cooled through the 175°–225°C isotherm approximately 305 Ma ago. Within the limits of error, no difference in age is apparent between zircons from sea level and zircons from an altitude of 1620. This indicates that there cannot have been a substantial difference between both crustal levels with regard to the time that they cooled through the 175°–225°C isotherm, which implies a fairly rapid uplift. It also confirms the conclusion on the basis of petrological and structural evidence that the nappe at Hardangerjøkulen was already in its present position before the termination of Caledonian orogenesis.

Concordancy between the zircon fission-track and U-Pb lower-intercept ages

The zircon fission-track ages are roughly concordant with the U-Pb lower-intercept age of 270^{+181}_{-208} Ma reported for a suite of zircons from the Eidfjord Granite (N.A.I.M. Boelrijk, private communication). Similar lower-intercept ages, between 350 and 330 Ma, have also been obtained from the Øye and Hafslo Granites, which belong to the younger intrusives into the Precambrian basement of the Jotun Nappe; the upper intercepts of these suites of zircons likewise correspond to Sveconorwegian ages of about 1000 Ma (Corfu 1978). A number of lower-intercept ages between 380 and 330 Ma have also been reported from elsewhere in the same general area, both in the autochthonous basement and in the overlying Paleozoic sedimentary rocks (Corfu 1979, 1980). In the Precambrian basement of southern Norway most suites of zircons likewise display lower intercepts corresponding to ages between about 400 and 300 Ma (Swainbank 1969, Pasteels & Michot 1975, Wielens et al. 1979). Lower-intercept ages in this range are thus widespread all over southern and south-western Norway, both within the Caledonian belt and in the Precambrian basement in front of it. In the latter terrain, this phenomenon has been related by Wielens et al. (1979) and Verschure (1981) to the influence of Caledonian orogenesis, along with the transition of the pumpellyite-prehnite facies to greenschist facies (temperature conditions roughly between 400° and 350°C).

The common interpretation of such consistent lower-intercepts of zircon discordia trajectories in a particular area is to attribute them to an episodic radiogenic lead loss. In cases that such an isotopic disturbance cannot be related to a metamorphic event reflected by other mineral ages, the episodic lead loss is usually explained in terms of Goldich & Mudrey's (1972) uplift-dilatancy model. Similary, the consistent range of lower-intercept ages of suites of zircons all over southern and south-western Norway may be related to the uplift and concomitant erosion of the crustal block following Caledonian orogenesis, leading to relaxation of pressure on the crystals and the action of hydrothermal solutions. The fission-track ages indicate that the ambient tem-

Table 2. Fission-track analytical data

Sample (altitude)	Mineral[a]	Fossil[b] tracks $\times 10^6/cm^2$	Induced[b] tracks $\times 10^6/cm^2$	Neutron[c] dose	Age $\pm 2\sigma$ (Ma)	Number of grains	r, $\bar{S}$[d]	U ppm
EID 3 (1620 m)	ap	0.30 (264)	0.23 (205)	2.18	166±31	50	0.09	3.5
EID 5 (700)	ap	4.61 (1809)	4.44 (1740)	2.18	134± 8	50	0.04	66.8
EID 4 (0)	ap	0.14 (306)	0.17 (362)	2.18	110±18	70	0.13	2.6
EID 3 (1620)	zir	12.48 (844)	2.31 (78)	0.96	294±70	7	0.96	70[e]
EID 1 (0)	zir	8.52 (1211)	1.46 (104)	0.96	316±66	6	0.96	44[e]
EID 4 (0)	zir	13.50 (939)	2.38 (83)	0.96	307±72	6	0.84	72[e]

a: ap = apatite; zir = zircon.
b: in parentheses the number of tracks actually counted is given.
c: $\times 10^{15}$ neutrons/cm^2.
d: r = correlations; $\bar{S}$ = relative standard error of the mean for the induced-track count.
e: this is an estimated value that applies only to the studied surfaces of a limited number of selected crystals with low-U content.

perature in the Eidfjord crustal segment at that time has been between 175° and 225°C.

Apatite fission-track ages

The fission-track data of the apatites from the granite Eid 4, the gneiss Eid 5 and the granitic gneiss Eid 3 are listed in Table 2, along with the calculated ages. Contrary to the zircons, the apatites give discordant ages. The ages increase as a function of topographical altitude: from approximately 110 Ma at sea level to approximately 134 Ma at 700 m and approximately 166 Ma at 1620 m. Such a pattern is commonly interpreted in terms of an uplift and cooling history of the crustal segment involved (Wagner et al. 1977, Naeser 1979, Zeitler et al. 1982).

Cooling and uplift history of the Eidfjord crustal segment

All ages available from the Eidfjord area, both the isotopic and fission-track data, are listed in Table 3. Except for the Rb-Sr whole-rock isochron and the U-Pb zircon upper-intercept ages which date the formation of the Sveconorwegian granite, all ages are interpreted to signal stages in the cooling history following Caledonian orogenesis. This thermal evolution is related to the post-Caledonian uplift of the crustal segment.

For each of the investigated samples a mean uplift rate U can be calculated on the basis of the ages of mineral pairs with different closure or annealing temperatures. The uplift rate follows from the equation

$$U = \frac{Tc(1) - Tc(2)}{A_1 - A_2} \cdot \left(\frac{\Delta T}{\Delta Z}\right)^{-1}$$

where $\Delta T/\Delta Z$ is the geothermal gradient, Tc(1) and Tc(2) are the annealing or closure temperatures of the minerals, and A_1 and A_2 are the mineral ages. The calculated uplift rates for four min-

Table 3. Age data from the Eidfjord area

			Age (Ma)
	whole-rocks	Rb-Sr isochron	911±35
	zircon	U-Pb upper intercept	905^{+86}_{-47}
		U-Pb lower intercept	207^{+181}_{-208}
EID 4 (0 m)	biotite	K-Ar	423±12
		Rb-Sr	386±10
	zircon	fission-track	307±72
	apatite	fission-track	110±18
EID 1 (0 m)	biotite	K-Ar	410±12
		Rb-Sr	381±10
	zircon	fission-track	316±66
EID 5 (700 m)	apatite	fission-track	134± 8
EID 3 (1620 m)	zircon	fission-track	294±70
	apatite	fission-track	166±23

Table 4. Uplift rates calculated from mineral pairs and the present-day surface temperature

Sample	method[a]	uplift rate[b] (mm/a)	temp. traject (°C)	time interval (Ma)
EID 4	bio-zir	0.1	400–200	386[c]–317
	zir-ap	0.02	200–105	307–110
	ap-surf T	0.03	105– 10	110–present
EID 4	bio-zir	0.1	400–200	396[d]–326
	zir-surf T	0.02	200– 10	316–present
EID 3	zir-ap	0.02	200–105	294–166
	ap-surf T	0.02	105– 10	166–present
EID 5	ap-surf T	0.02	105– 10	134–present

a: bio = biotite; zir = zircon; ap = apatite; surf T = surface temperature.
b: assumed geothermal gradient is 30°C/km.
c: Rb-Sr age of the biotite.
d: mean value of the biotite Rb-Sr and K-Ar ages.

eral pairs from three samples (Eid 1, 3 and 4), covering the temperature intervals 400°–200°C and 200°–105°C, are listed in Table 4. For the calculations it is assumed that (1) the geothermal gradient was 30°C/km, and that (2) the geothermal gradient has remained constant since approximately 390 Ma ago. Several temperatures at which the isotopic Rb-Sr and K-Ar systems of biotite behave like a closed system are known from the literature; they range from about 225°–400°C (Turner & Forbes 1976, Verschure et al. 1980), with a commonly used value of about 300°C (Jäger et al. 1967). For the Eidfjord area there are several reasons to use the temperature of 400°C as proposed by Verschure et al. (1980). The metamorphic- and petrological conditions show similarities to those described by Verschure et al. (1980) in SW Norway, indicating that recrystallisation of the biotite played an important role. The biotite ages fall within the Caledonian age range, with K-Ar ages higher than the corresponding Rb-Sr ages. The same phenomenon has also been observed in SW Norway near the Caledonian front. The annealing temperatures of zircon and apatite are taken at 200°C and 105°C, respectively. Such a relatively high temperature is adopted for the zircon in view of the concordancy between zircon ages from samples with a difference in topographical altitude as high as 1620 m. This points to a high cooling rate at the zircon annealing temperature, which may be related to the more rapid cooling under conditions of higher ambient temperature in the beginning of the post-Caledonian thermal history. The an-

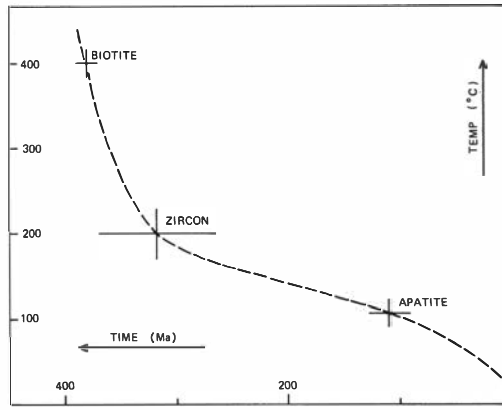

Fig. 2. Time-temperature diagram of granite sample Eid 4 (now at sea level).

nealing temperature of apatite is well established: it ranges from about 150°C to 105°C, depending on the time-span during which the conditions of high temperature prevail (Wagner 1968, Naeser 1981, Gleadow & Duddy 1980). For a slow cooling rate, as is to be expected in the case of crustal cooling in relation to uplift, an annealing temperature of 105°C is commonly taken for apatite.

The mean uplift rate U_{ap} since the rocks have cooled through the 105°C isotherm is calculated by means of the equation

$$U_{ap} = \frac{105°C - P}{A_{ap}} \cdot \left(\frac{\Delta T}{\Delta Z}\right)^{-1}$$

where 105°C is the annealing temperature for apatite, A_{ap} is the age of the apatite, and P is the present-day mean annual surface temperature, taken at 10°C. Table 4 lists the uplift values for the temperature interval 105°C–10°C, calculated for the samples Eid 3, 4 and 5.

Fig. 2 shows a plot of the biotite, zircon and apatite ages against the closure or annealing temperature for granite Eid 4, a sample taken at sea level. The curve through the data-points and the present-day 10°C point represents the cooling history of this sample. A plot of the apatite ages against the altitude of the host sample is shown in Fig. 3. This plot illustrates the mean uplift rate U_{alt} for the time interval from 166 until 110 Ma ago. This rate follows simply from the equation

$$U_{alt} = \frac{\Delta \text{altitude}}{A_1 - A_2}$$

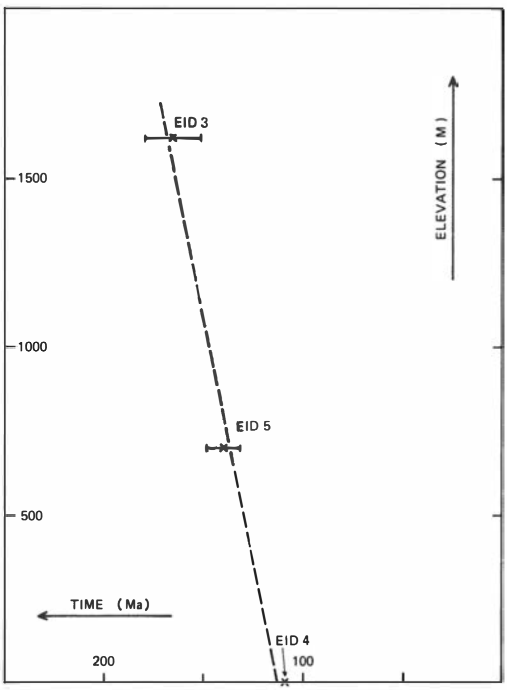

Fig. 3. Age-elevation diagram of the apatites.

where Δaltitude is the difference in altitude between two samples with ages A_1 and A_2. It is again assumed that the geothermal gradient remained constant during the time interval $A_1 - A_2$, as well as that the absolute difference in altitude between the two samples has not been changed (for example, through tectonic processes). The uplift rates calculated in this way are listed in Table 5. They are very similar to the uplift rates calculated from the mineral pairs.

The data listed in Tables 3 and 4, along with the time-temperature diagram of Fig. 2, show that uplift and erosion of the Eidfjord crustal segment were rather rapid during the first 80 Ma or so following the termination of Caledonian orogenesis, with a mean uplift rate of 0.1 mm/a. This rapid post-orogenic uplift is also reflected in the thick deposits of coarse conglomerates and sandstones in the stratigraphic record of the Devonian, the Old Red Sandstone (e.g., Holtedahl 1960). Then the uplift rate slowed to a mean value of 0.02 mm/a for the last 300 Ma or so, until the present. It is interesting to note that similar uplift rates have been reported for two Caledonian granitic intrusions in East Greenland (Gleadow & Brooks 1979).

Table 5. Mean uplift rates between about 166 Ma and 110 Ma ago, calculated from apatites at different topographic altitude

Samples	elevation difference (m)	uplift rate (mm/a)
EID 3–EID 5	1620–700	0.03
EID 5–EID 4	700– 0	0.03

Tectonic implications

The post-Caledonian uplift of the Eidfjord crustal segment was accompanied by denudation and peneplanation at the surface. The thickness of the crustal layer that has been removed by erosion can be estimated on the basis of the fission-track age pattern of the zircons and apatites. Sample Eid 5, now located at an altitude of 700 m, must have been situated some 3000 m below sea level 134 Ma ago, implying that since that time (Late Jurassic) approximately 3700 metres of crustal material have been eroded. The apatite age of sample Eid 3, now situated at an altitude of 1620 m, indicates that since 166 Ma ago (Late Jurassic) some 4600 m have been removed. Similarly, it can be estimated from the zircon ages that since about 305 Ma ago (Late Carboniferous) some 8000 m of crustal material have been eroded, and from the biotite ages that since about 390 Ma ago (shortly after the termination of Caledonian orogenesis) over 13,000 m have been eroded. All these calculations are based on a geothermal gradient of 30°C/km, assuming that it has remained constant over the last 390 Ma.

Acknowledgments. – The authors are much indebted to N. A. I. M. Boelrijk, E. H. Hebeda, H. N. A. Priem, R. H. Verschure, all of the Z. W. O. Laboratory of Isotope Geology, Amsterdam, for valuable discussions. H. N. A. Priem and R. H. Verschure also provided the samples and geological information. The zircon U-Pb data were provided by N. A. I. M. Boelrijk. An earlier version of the manuscript benefited from a thorough review by H. N. A. Priem. The irradiation of the zircons and apatites was performed in E. C. N., Petten, using the Z.W.O./W.A.C.O.M. facilities. This work forms part of the research programme of the 'Stichting voor Isotopen-Geologisch Onderzoek', supported by the Netherlands Organisation for the Advancement of Pure Research (Z.W.O.).

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