Canadian Mineralogist
Vol. 20, pp. 319-332 (1982)

METAMORPHIC EVIDENCE FOR RAPID (2 mm/yr) UPLIFT OF A PORTION OF THE CENTRAL GNEISS COMPLEX, COAST MOUNTAINS, B.C.

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ABSTRACT

The Eocene granulite-facies metamorphic terrane in the Coast Mountains east of Prince Rupert, B.C. was apparently uplifted and eroded at the unusually fast rate of 2 mm/yr. Reactions showing the transition from high-pressure to low-pressure assemblages are the best evidence for this interpretation. These include garnet $\rightarrow$ hypersthene + anorthite, almandine/pyrope + sillimanite $\rightarrow$ cordierite, sillimanite + grossular $\rightarrow$ anorthite, garnet + biotite + sillimanite $\rightarrow$ cordierite + hercynite + orthoclase, and kyanite $\rightarrow$ sillimanite. The occurrences of low-density $CO_2 + H_2O$ fluid inclusions, of naturally decrepitated fluid inclusions, and of andalusite in late discordant veins imply that the rocks were still at moderate (>400°C) temperature when relatively near the surface, which is consistent with a model of rapid uplift. Assuming no additional heat input during uplift, petrological, fluid inclusion, geological and available age data from the Central Gneiss Complex can be fit by a model uplift curve published by Albarède (1976). In the model, the uplift of rocks exposed at present occurred between 62 and 48 Ma, beginning at 35 km and terminating at 5 km.

Keywords: uplift rate, metamorphism, fluid inclusions, Coast Mountains, British Columbia.

SOMMAIRE

Le domaine métamorphique à faciès granulite d'âge éocène de la chaîne côtière, situé à l'est de Prince Rupert (Colombie-britannique), semble avoir été soulevé et érodé au taux anormalement rapide de 2 mm par année. Les réactions qui signalent la transition d'assemblages de haute à basse pression fournissent le meilleur argument à l'appui de notre interprétation: grenat $\rightarrow$ hypersthène + anorthite, almandin/pyrope + sillimanite $\rightarrow$ cordiérite, sillimanite + grossulaire $\rightarrow$ anorthite, grenat + biotite + sillimanite $\rightarrow$ cordiérite + hercynite + orthose, et disthène $\rightarrow$ sillimanite. Les inclusions fluides à $CO_2 + H_2O$ de faible densité, les inclusions fluides naturellement décrépitées, et la présence d'andalousite tardive en filonnets discordants montrent que ces roches étaient encore à température assez élevée (> 400°C) relativement près de la surface; ces observations concordent avec le modèle de soulèvement rapide. En supposant qu'il n'y ait eu aucun apport de chaleur pendant ce soulèvement, on peut interpréter les données pétrologiques, géologiques et radiométriques, ainsi que celles que révèlent les inclusions fluides, par la courbe d'Albarède (1976) décrivant le soulèvement. Selon ce modèle, le soulèvement du complexe central gneissique, qui affleure aujourd'hui, débuta il y a 62 Ma à 35 km de profondeur pour se terminer il y a 48 Ma à 5 km de la surface.

(Traduit par la Rédaction)

Mots-clés: taux de soulèvement, métamorphisme, inclusions fluides, chaîne côtière, Colombie-britannique.

INTRODUCTION

The Central Gneiss Complex (CGC) of the Coast Mountains, B.C. contains unusually young granulite-facies assemblages (Hollister 1975) within the predominantly migmatitic and plutonic terrane that forms the core zone of the Coast Plutonic Complex (Hutchinson 1970, Roddick & Hutchison 1974). Subsequent work in the CGC east of Prince Rupert (Fig. 1) includes a study on the occurrence of CO_2 and CO_2 + H_2O fluid inclusions in the granulite-facies rocks (Hollister & Burruss 1976), a determination of pressure and temperature of metamorphism (5 kbar, 750°C) for one locality within the granulite-facies terrane (Selverstone & Hollister 1980), and descriptions of the melt-producing reactions within the migmatite (Lappin & Hollister 1980, Kenah 1979). Petrological features noted in these studies suggest a post-peak metamorphic P–T path that brought the rocks to shallow depth while still at high temperatures (Hollister 1979).

Age dates published by Armstrong & Runkle (1979) and Harrison *et al.* (1979) show that the Quottoon pluton (Fig. 1) cooled from ~700 to ~400°C in only 3 Ma. Harrison & Clarke (1979) used these results to suggest that the Quottoon pluton cooled rapidly because it intruded into country rocks that already were ~200°C cooler than the pluton at the time of intrusion. However, Kenah (1979) showed that

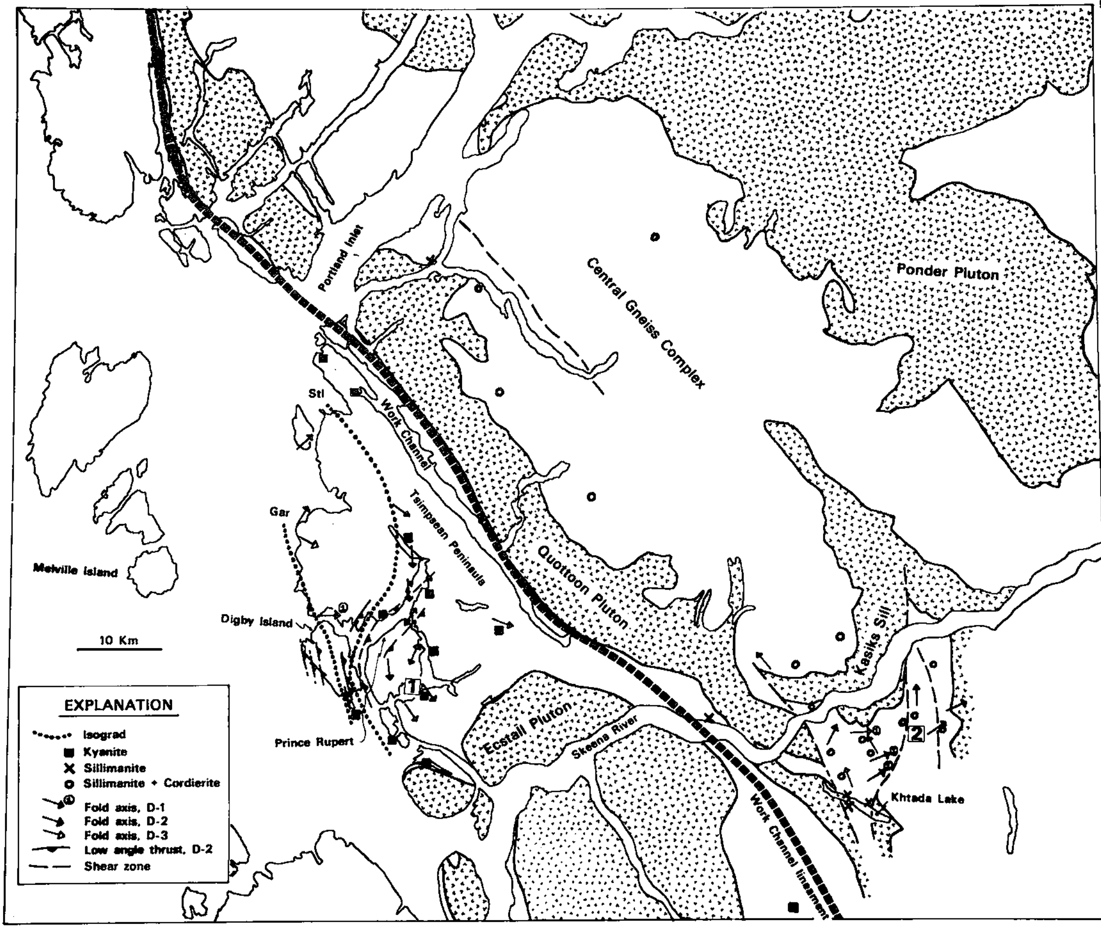

FIG. 1. The geology of the Prince Rupert – Skeena map area, from Hutchison (1971) and Crawford & Hollister (1982). The area south of Kasiks Sill, east of the Quottoon pluton, and including Khtada Lake, is the region referred to here as the Khtada Lake Metamorphic Complex (KLMC). Detailed petrology at locality 1 is given by Crawford *et al.* (1979) and at locality 2, by Selverstone & Hollister (1980).

the Quottoon pluton intruded as a crystalline mush more or less synchronously with the regional development of the granulite-facies metamorphism. Based on the P–T estimate of Selverstone & Hollister (1980), the country rocks were at ~750°C at ~17 km, or 150°C greater than assumed by Harrison & Clarke (1979) when the Quottoon pluton intruded. The geological arguments, therefore, call for another model to explain the rapid cooling of the Quottoon pluton.

The model developed in the present paper is based on an analysis by Albarède (1976), who calculated that rapid uplift (2–5 mm/yr) could account for the development of a sequence of metamorphic assemblages observed in the Massif Central, France, and similar to that found in the CGC. The petrological and fluid-inclusion data for the CGC define a pressure–temperature path during uplift that is strongly convex toward the temperature axis. The cooling dates, used to estimate a portion of the time–temperature history of the area, combined with the pressure–temperature path and the thermal models of Albarède, lead to the conclusion that the petrological, fluid inclusion, and age data are consistent with a rate of uplift of the CGC of 2 mm/yr over a time span of about 15 Ma.

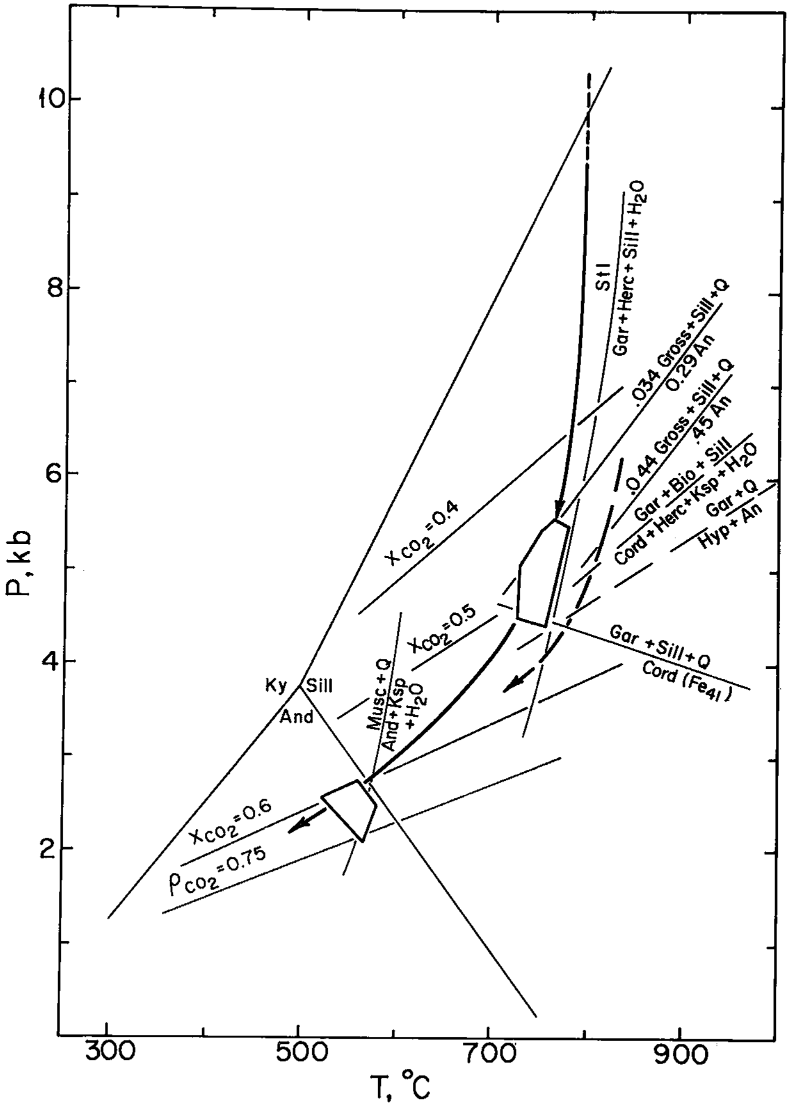

FIG. 2. Phase-equilibria curves pertaining to uplift of the Khtada Lake Metamorphic Complex (KLMC). The P–T region at about 5 kbar applies to locality 2, Figure 1. The P–T region at about 2 kbar, the fluid isochores, and the reactions are discussed in text. Heavy solid line is the model-uplift curve discussed in the text (Fig. 7) for an uplift rate of 2 mm/yr, fitted to the petrological data of locality 2 and the P–T region at 2 kbar, and dashed to show extension into the kyanite field. The heavy dashed curve shows uplift of the higher-temperature metamorphic rocks in the KLMC (Kenah 1979) than for locality 2. Ky (kyanite), And (andalusite), Sil (sillimanite), Stl (staurolite), Gar (garnet), Herc (hercynite), Cord (cordierite), An (mole fraction of anorthite in plagioclase), Gross (mole fraction of grossular in garnet), Q (quartz), Musc (muscovite), Ksp (orthoclase), Hyp (hypersthene), Bio (biotite).

TABLE 1. MINERAL FORMULAE

Sample	C2U				K12C						TPR 30D			TPR 30D-A		
Mineral	Gar(12)	Plag(8)	Opx(6)	Hbl(22)	Gar(12)	Plag(8)	Ksp(8)	Cord(18)	Bio(20)	Herc(4)	Gar(12)	Bio(20)	Herc(4)	Gar(12)	Herc(4)	Stl(46)
Si	3.01	2.12	1.96	6.67	2.98	2.66	2.97	4.99	5.38	.00	3.00	5.48	.00	3.00	.00	6.79
Al	1.93	1.87	.03	1.27	2.05	1.38	1.06	4.04	3.46	1.97	2.01	3.34	1.99	1.99	1.99	18.72
Fe	1.99	.004	1.21	1.85	2.38	.00	.00	.67	2.47	.80	2.07	1.86	.59	2.18	.70	2.94
Mg	.28	.00	.74	2.46	.48	.00	.00	1.29	2.10	.19	.73	3.00	.26	.60	.23	.81
Mn	.23	.00	.04	.03	.02	.00	.00	.00	.00	.00	.07	.00	.00	.07	.00	.02
Ca	.60	.91	.04	1.75	.09	.33	.02	.00	.00	.00	.11	.00	.00	.16	.00	.00
Na	.00	.10	.00	.23	.00	.54	.07	.00	.00	.00	.00	.04	.00	.00	.00	.00
K	.00	.001	.00	.13	.00	.01	.85	.00	1.65	.00	.00	.86	.00	.00	.00	.00
Ti	.004	.00	.003	.21	.00	.00	.00	.00	.33	.00	.00	.24	.00	.00	.00	.25
Zn	.00	.00	.00	.00	.00	.00	.00	.00	.00	.06	.00	.00	.14	.00	.09	.07

Numbers in parentheses are numbers of oxygen atoms per formula unit used to calculate cation proportions.

Petrological Data Suggesting Rapid Uplift

Metamorphic mineral assemblages and textures from the Khtada Lake Metamorphic Complex (KLMC) within the CGC (Fig. 1) suggest that the rocks contain evidence of metamorphic reactions arrested during a natural quenching process. Phase assemblages characteristic of high pressures and temperatures are partially replaced by assemblages indicating lower pressures. Although an interpretation of isobaric temperature increase could account for some of the textures, pressure decrease without significant temperature change is the most reasonable overall interpretation. After an initial rapid, nearly isothermal decrease in pressure, the temperature must have dropped sufficiently rapidly to preserve the relics of the high-pressure assemblages and the low-pressure reaction products.

Selverstone & Hollister (1980) defined the P–T conditions at locality 2, Figure 1; these P–T values are represented by the six-sided region shown on Figure 2. Based on other data from the KLMC, metamorphic conditions within the KLMC vary from these conditions ($\sim$5 kbar and $\sim$750°C) to 800°C or higher (Kenah 1979), at about the same pressure, based on occurrences of garnet + hercynite + sillimanite in quartz-free assemblages, which must have formed above the staurolite breakdown curve (Fig 2). This assemblage is common in the western portion of the KLMC. In one sample (TPR30D–A, Table 1), staurolite, hercynite and sillimanite occur as an assemblage totally enclosed within garnet. This assemblage probably formed on the staurolite breakdown curve (Fig. 2), which is multivariant with respect to Mn, Zn, Mg and $P(H_2O)$. Thus, a range of temperature conditions from those of locality 2 to above the staurolite breakdown curve in quartz-free assemblages existed across the KLMC. In the subsequent discussions, rocks that crystallized over this range of temperature, at about 5 kbar, are referred to as sillimanite–cordierite gneisses. However, for simplicity, the uplift model is only presented with respect to the P–T conditions at locality 2; petrological evidence for the uplift, on the other hand, comes from rocks that reached temperatures ranging from 750 to over 800°C.

The four-sided P–T region centred at about 2.5 kbar and 550°C is based on occurrences of andalusite-bearing veins that cross-cut the fabric of the sillimanite–cordierite gneisses. The assemblage in the veins is andalusite–fibrolite–muscovite–orthoclase–quartz–plagioclase, which, assuming equilibrium crystallization, constrains the P–T conditions of formation of the veins to lie along the muscovite + quartz = potassium feldspar + andalusite (or sillimanite) reaction curve for the appropriate $P(H_2O)$. The position of the reaction boundary is also dependent on additional components in the phases, but analyses show these to be too low in concentration to significantly offset the positions of the reaction curves. The position of the curve shown in Figure 2, for $P(H_2O) = \frac{1}{2} P_{total}$ (Kerrick 1972), was chosen because it crosses the andalusite–sillimanite boundary of Holdaway (1971) at the intersection with the isochore for commonly occurring secondary $CO_2 + H_2O$ fluid inclusions that have a mole fraction CO_2 of about 0.6 and a molar volume of 43 cm^3/mole. The isochore for typical low density ($\rho = 0.75$ g/cm^3) pure CO_2 secondary fluid inclusions found in the CGC is also used to help constrain the P–T conditions of formation of the andalusite veins. Fluid inclusions found within andalusite are now mixtures of $CO_2 + CH_4$ ($X(CH_4)$ $\sim$0.01), suggesting a low $P(H_2O)$ when the andalusite grew. If a higher-temperature position for the andalusite–sillimanite boundary were used, such as the one proposed by Greenwood (1976), which has the field of andalusite overlapping that of granite melt, the P–T conditions of the andalusite veins would be at somewhat higher temperature and pressure. Discussion on the use of fluid inclusions in support of the uplift model is given in the next section.

Several textural examples of lower-pressure assemblages that have partially replaced higher pressure assemblages have previously been described. Hollister (1977) illustrated and discussed the occurrence of kyanite inclusions in garnet from samples where the Al_2SiO_5 polymorph in the matrix is sillimanite. Selverstone & Hollister (1980) described cordierite-bearing low-pressure assemblages that replaced higher-pressure garnet + sillimanite + quartz assemblages at locality 2; following the procedure of Holdaway & Lee (1977), they calculated the position of this curve in P–T space [Gar + Sill + Q = Cord(Fe_{41}); Fig. 2]. Figure 2 also shows the position of the curve for the reaction garnet (0.034 Grossular) + sillimanite + quartz = plagioclase (0.29 anorthite), calculated according to the procedure of Ghent *et al.* (1979) for locality 2 (Selverstone & Hollister 1980). Curves based on other data from the KLMC are bracketed by this curve and the one for 0.044 grossular and 0.45 anorthite (Fig. 2), which is taken from Kenah (1979).

Two additional textural examples of apparent lower-pressure assemblages that have partially replaced higher-pressure assemblages have been identified in the KLMC. These are garnet + quartz = hypersthene + anorthite, and garnet + biotite + sillimanite = cordierite + hercynite + orthoclase. Both are shown as dashed curves on Figure 2 because their locations in P–T space are not well constrained. They are shown to pass through the estimated P–T conditions with slopes calculated as discussed below. The occurrences are also described in the following paragraphs.

A representative sample of a symplectite of hypersthene + anorthite partially replacing garnet is illustrated in Figure 3. The reaction giving this association, garnet + quartz = hypersthene + anorthite, was described by Perkins & Newton (1981) as one that could be used to calculate pressure of metamorphism because the slope of the reaction curve in P–T space is relatively flat. Using the thermodynamic data, including activity coefficients, given by Perkins & Newton (1981) and the composition data for sample C2U (Table 1), the reaction for this sample would pass through 3.2 kbar, 800°C with a slope of –0.7 bar / °C. However, this calculation involves a substantial extrapolation

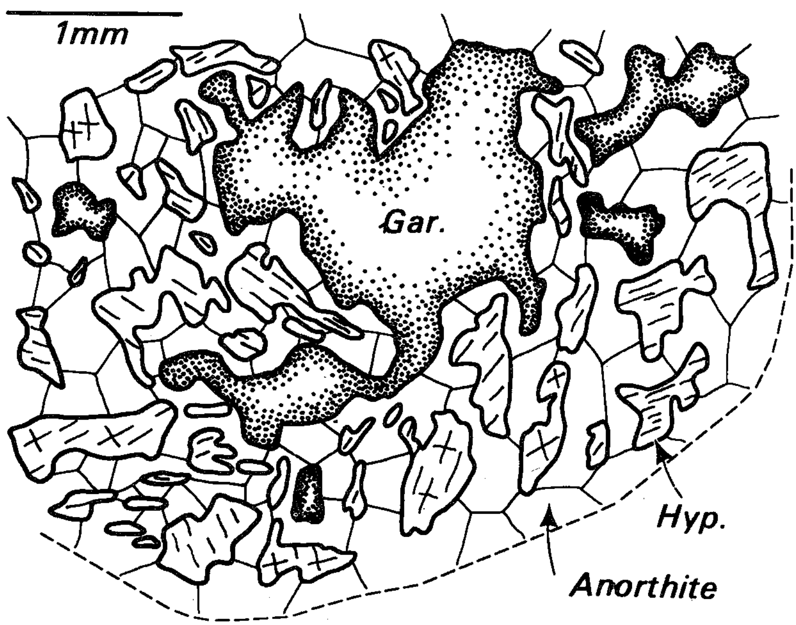

FIG. 3. Sketch of a symplectite of anorthite (clear) and hypersthene (shown with cleavage) replacing garnet. Dashed line shows the former limit of garnet grain. Balanced reaction is given in Table 2. Sketch after Kenah (1979).

TABLE 2. CALCULATED SLOPES OF EQUILIBRIA AT 5 KBAR AND 800°C

Reaction A		
Reactants		
2.0 $Fe_3Al_2Si_3O_{12}$	115.28	
1.0 $Ca_3Al_2Si_3O_{12}$	125.3	
3.0 SiO_2	23.72	$\Delta S_r \approx 11.46$ cal/°C
		$\Delta V_r = 73.06$ cm^3
Products		$\frac{dp}{dt} \approx 9.4$ bars/°C
6.0 $FeSiO_3$	32.952	
3.0 $CaAl_2Si_2O_8$	100.79	
Reaction B		
Reactants		
0.9 $Fe_3Al_2Si_3O_{12}$	115.28	
1.0 $KMg_{1.5}Fe_{1.5}AlSi_3O_{10}(OH)_2$	151.99	
4.8 Al_2SiO_5	49.9	$\Delta S_r \approx 22.29$ cal/°C
		$\Delta V_r = 99.33$ cm^3
Products		$\frac{dp}{dt} \approx 9.4$ bars/°C
1.5 $MgFeAl_4Si_5O_{18}$	233.22	
2.7 $FeAl_2O_4$	40.75	
1.0 $KAlSi_3O_8$	109.01	
1.0 H_2O	25.73	

from the pure Mg system; the Mn content of the garnet sample C2U exceeds the limit recommended by Perkins & Newton for use of their thermodynamic data to calculate the position of the reaction curve. The slope shown on Figure 2 was estimated from the Clapeyron curve (reaction A, Table 2) using ΔS of reaction calculated from the ΔS of the change from 6-co-ordinated Al to 4-co-ordinated Al [80 dj/°atom (Albee 1965)], and using ΔV of reaction from molar volumes given in Helgeson *et al.* (1978). This slope (6.6 bar / °C) is nearly the same as that (6.4 bar / °C) obtained from the data of Perkins & Newton (1981) for the reaction grossular + 2 pyrope + quartz = enstatite + anorthite, assuming ideal mixing of grossular and pyrope.

The texture showing the reaction garnet + biotite + sillimanite = hercynite + cordierite + orthoclase is illustrated in Figure 4. The composition of the minerals are given in Table 1 (sample K12C). Data for a sample (TPR30D) with the quartz-free assemblage hercynite + garnet + sillimanite are also given in Table 1. In calculating the Clapeyron curve for the reaction (B, Table 2), the same value of ΔS for co-ordination change of Al was used as for reaction A, cordierite was assumed to be anhydrous, ΔS of dehydration at 800°C and 5 kbar was interpolated from Fyfe *et al.* (1958), and

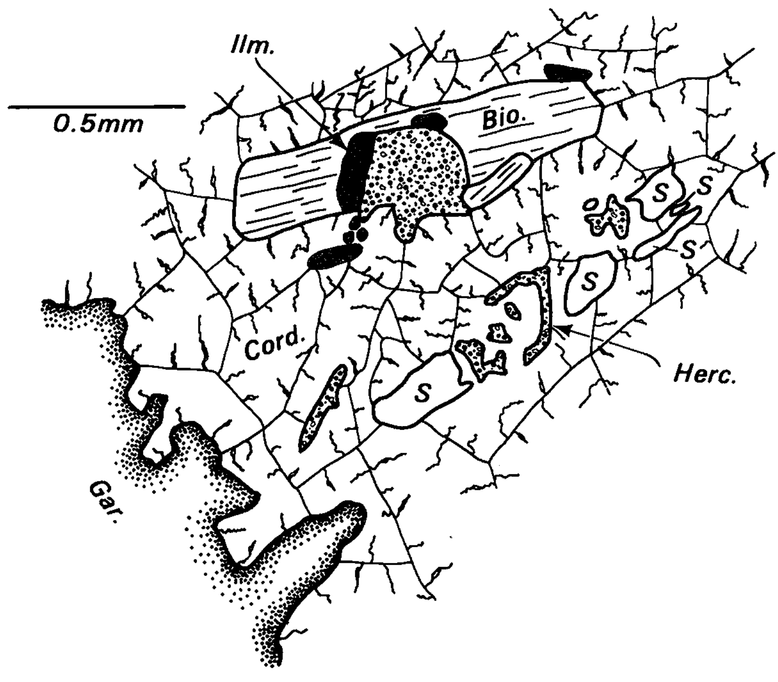

FIG. 4. Sketch of texture illustrating reaction Bio + Sill + Gar = Cord + Herc + Ilmenite (Ilm) + Ksp + H_2O. The sillimanite grains (S) are in optical continuity. Other abbreviations are as in Figure 2. Biotite with cleavage, ilmenite black, sillimanite clear, hercynite speckled, cordierite with microcracks from grain boundaries. Balanced reaction is given in Table 2. Sketch after Kenah (1979).

the molar volume of hercynite was taken from Robie *et al.* (1978). Although the calculated slope (9.4 bar / °C) is less well constrained than the slopes of the other reactions shown on Figure 2, it is reasonable to conclude that the reaction boundary has a relatively shallow slope.

For all five reactions discussed above, the direction of reaction was probably from higher to lower pressures. Several of the reactions could have resulted from an isobaric increase of temperature, but not the reaction garnet + sillimanite + garnet = cordierite; it is unlikely that the reaction garnet + quartz = hypersthene + anorthite resulted solely from a temperature increase because of its very low slope (–0.7 to +6.6 bar / °C).

If a model of monotonic decrease of pressure without additional heat input to the system is invoked to explain the mineral assemblages and the textures, the decompression path would begin in the kyanite field at about 10 kbar, cross the reaction curves as shown by the heavy solid and dashed lines on Figure 2 between about 725 and 800°C and continue with cooling to pass through the estimated P–T conditions of the andalusite-bearing veins.

A second simple model could account for the petrological data from the KLMC. A P–T path could begin in the kyanite field at lower P and T (for example, 5.5 kbar and 500°C), pass more or less isobarically to about 800°C as a result of input of heat from intrusion of tonalite to gabbro plutons and sills, and then cool, along with a drop in pressure, through the conditions of formation of the andalusite-bearing veins. This model seems less likely for two reasons. First, pelitic assemblages virtually everywhere in the region of the CGC shown in Figure 1 typically contain sillimanite + biotite + orthoclase, indicative of regional temperatures above at least 650°C. Second, to the west of Work Channel lineament (at locality 1, Fig. 1), kyanite-bearing rocks were metamorphosed at pressures above 9 kbar (Crawford *et al.* 1979), indicating that at least some rocks presently exposed in the Coast Plutonic Complex had been at the depths proposed for the beginning of the decreasing pressure path illustrated in Figure 2 (see also Crawford & Hollister 1982).

Fluid-Inclusion Data Supporting Uplift

The typical fluid inclusions found in quartz for several rock associations in the area have shapes characteristic of autodecrepitation. This happens when a fluid entrapped at a given P and T is taken to a new P–T regime and the internal pressure within the inclusion exceeds lithostatic pressure by about 1–2 kbar (Naumov *et al.* 1966, Leroy 1979). This phenomenon was first described by Lemmlein (1956). Bilal & Touret (1976) discussed it for the specific case of xenoliths of granulite-facies rocks carried from depth to the surface by volcanic eruption; Hollister *et al.* (1979) described it for the case of paths of metamorphic uplift in general.

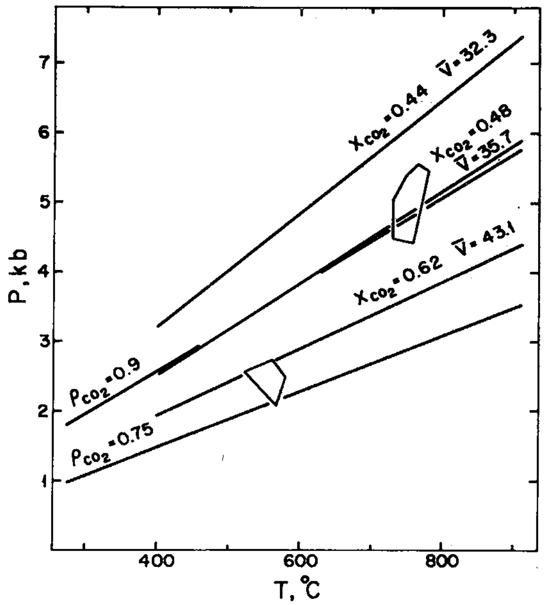

FIG. 5. Isochores for fluids found in inclusions from the Khtada Lake Metamorphic Complex. The two P–T regions of Figure 2 are shown for reference. The parameters density ρ, specific volume $\bar{V}$ and mole fraction X are written above each isochore.

Figure 5 shows isochores for several populations of fluid inclusions found in the Khtada Lake Metamorphic Complex. For each composition, the fluid must have equilibrated at P–T conditions lying on its isochore. For the CO_2+H_2O fluids that were trapped as homogeneous fluids, the temperature at entrapment must have been above the solvus for CO_2 + H_2O + dissolved salt. This temperature is approximately 400°C for the typical salt content observed in fluid inclusions from the Khtada Lake Metamorphic Complex (Hollister & Burruss 1976); Hendel & Hollister (1981) and Sisson *et al.* (1982) discuss further the CO_2+H_2O solvus in natural systems. For the pure CO_2 fluids, one generally cannot limit the P–T conditions to a line segment.

The most dense inclusions do have isochores passing through and above the estimated P–T conditions of final mineral equilibration for locality 2 (Figs. 1,2,5); but many of these inclusions have shapes suggesting natural decrepitation, which implies that they formed by decrepitation of denser (earlier) inclusions when internal pressures had exceeded 1–2 kbar. In other words, the rock was experiencing changes in P and T similar to those deduced from the mineral-reaction textures when the inclusion fluids were trapped.

The rocks must also have experienced pressures and temperatures that lie along the isochores for the least dense fluids found. These fluid inclusions are, in fact, the most abundant in the area and typically occur along healed fractures. For the inclusions with $X(CO_2) = 0.6$, the rocks must have been on the isochore above about 400°C. A path involving a drop in pressure must also be invoked to account for the low-density pure CO_2 inclusions.

The fact that the low-density isochores pass through the region for estimated P–T conditions of formation of the andalusite-bearing veins (Fig. 5) should not be taken as unique evidence confirming these conditions, because the isochores were used to constrain the P–T conditions shown on Figure 2. The P–T conditions near 5 kbar, however, were obtained independently (Selverstone & Hollister 1980) of the isochores for the densest fluids.

The fluid-inclusion data, therefore, are consistent with the model of a path of decreasing pressure suggested by the mineral textures and compositions. Naturally decrepitated inclusions imply a P–T path that crossed isochores towards lower-density fluids, and the occurrences of mixed CO_2–H_2O compositions in these inclusions imply that the decrepitation events recorded by the inclusions occurred above the CO_2–H_2O solvus, which is at about 400°C. The preservation of the densest inclusions restricts the P–T path to temperatures below ~600°C at low pressures (*e.g.*, ~2 kbar); otherwise, they too would have decrepitated. Finally, the P–T path must have passed across or along the lowest-density isochores to account for the existence of these inclusions. The P–T path shown in Figure 2 meets these constraints as well as the petrological constraints. Again, however, it must be mentioned that a model of isobaric heating at about 5 kbar, followed by decrease of both P and T, would also be consistent with the fluid-inclusion data.

Age Constraints

According to Kenah (1978, 1979), the Quottoon pluton intruded while the country rock along its eastern flank was at the metamorphic conditions of the sillimanite–cordierite gneisses. No thermal gradient is recorded by the mineral assemblages in the country rock, nor is there evidence of secondary deuteric alteration related to the pluton. The country rocks near the pluton have migmatitic textures similar to those well away from the contact. Kenah (1979) concluded that the country rocks were already in a partly molten state when the pluton intruded and that, at a deeper level than that presently exposed, this partial melt had contributed to the mass of the pluton. The pluton is well foliated in many places and cut by shear zones that postdate the metamorphism (Fig. 1).

Harrison *et al.* (1979) used fission track, K–Ar dates and a Rb–Sr whole-rock isochron obtained by Armstrong & Runkle (1979) to determine the cooling history of the Quottoon pluton at a locality south of the area shown in Figure 1. These dates are shown on Figure 6. Because the Quottoon pluton intruded when the country rocks were at the P–T conditions of the sillimanite–cordierite gneisses, these cooling dates are considered applicable to the KLMC as a whole.

The Rb–Sr whole-rock date on the Quottoon pluton is inferred to have been set when the rocks had cooled sufficiently that the partial melt in the gneisses and a substantial portion of the Quottoon pluton had crystallized. Thus, this date was set during cooling and, based on the petrological evidence for decrease of pressure at high temperature, during uplift. Lappin & Hollister (1980) determined the reactions and estimated the conditions for partial melting in a zone of migmatites on the west flank of the Quottoon pluton along the Skeena River (Fig. 1). They pointed out that, in the presence of aqueous vapor, partial melting of biotite–quartz–plagioclase gneiss would begin at about 700°C, between 3 and 8 kbar, with hornblende formed as a reaction product. This agrees with the experiments of Büsch *et al.* (1974). Kenah (1979) described the same melt-producing reaction in the migmatite to the east of the Quottoon pluton. Final crystallization of vapor-saturated melt in the CGC would thus also be at about 700°C. It is therefore concluded that the Rb–Sr whole-rock age (51 ± 2 Ma) was set at about 700°C when the melt of the

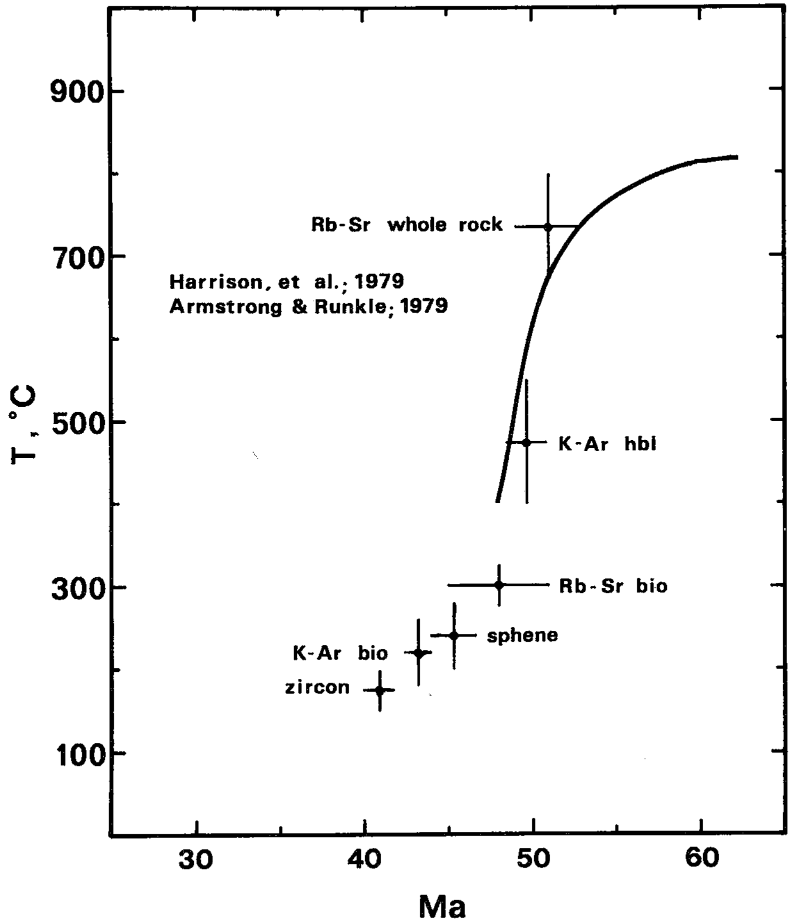

FIG. 6. Cooling dates obtained by Armstrong & Runkle (1979) and Harrison *et al.* (1979) for the Quottoon pluton. The solid line curve is the cooling curve (Fig. 9) from Albarède (1976) placed to fit the age data. The error bars shown are from Harrison *et al.* (1979).

Quottoon pluton and of the migmatite of the CGC crystallized during uplift.

The date of the last thermal event during the Eocene may be deduced from the K–Ar ages of plutonic rocks reported by Smith & Diggles (1981) from the CGC in southeast Alaska just north of the area shown in Figure 1. The youngest dates in the report of Smith & Diggles also have the smallest difference between K–Ar dates on biotite and hornblende and, therefore, are the closest to being concordant. They differ by less than 2.7 Ma; biotite ages of five samples are 46.5 – 47.5 Ma, and hornblende ages are 47.6 – 49.8 Ma. The average of the five hornblende ages is 48.8 Ma, considered the data for the last Eocene igneous activity in the CGC. Note that the hornblende date is close to that (49.6) obtained by Harrison *et al.* (1979) on the Quottoon pluton and suggests that the CGC as a whole had cooled to the hornblende-blocking temperature (~500°C) when the last igneous activity occurred.

This last igneous activity is correlated by Smith & Diggles (1981) with the intrusion of the Ponder pluton, which, here also, is correlated with formation of the cross-cutting andalusite veins. Neither these veins nor the Ponder pluton are deformed; both are, therefore, considered to postdate the last folding event in the CGC that was synchronous with the granulite-facies metamorphism (Crawford & Hollister 1982). The Ponder pluton is considered a relatively shallow-level intrusive body because, where it intrudes previously unmetamorphosed

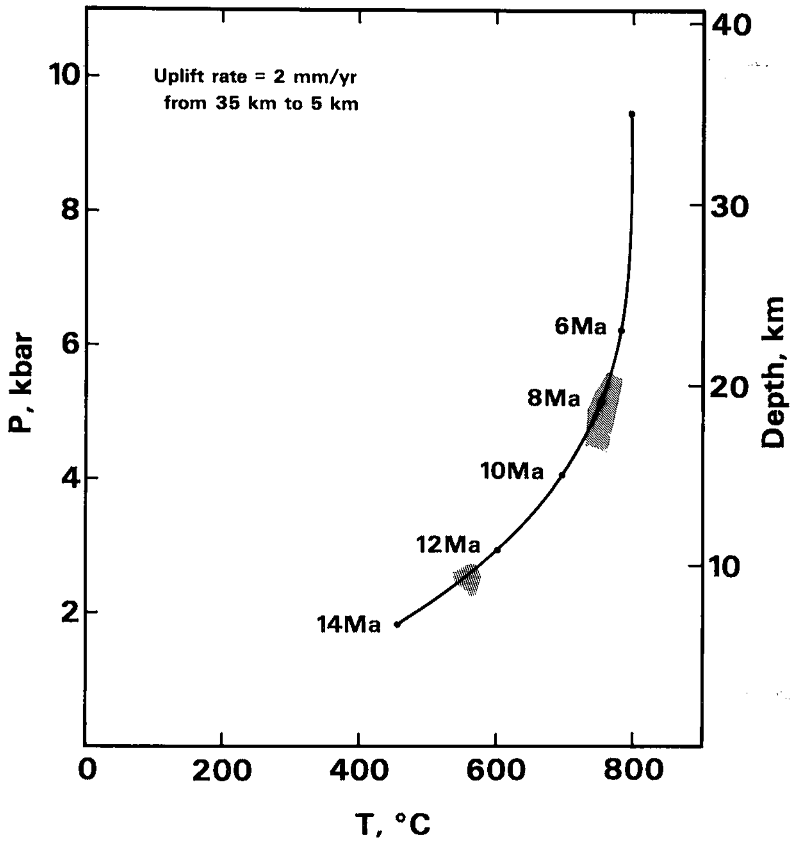

FIG. 7. Uplift curve from Albarède (1979, model B; his Fig. 9) that most closely fits the petrological and age data of the Khtada Lake Metamorphic Complex. Time intervals shown begin at the time uplift commenced. The P–T regions (hatched areas) are from Figure 2.

sediment along its eastern contact, andalusite occurs in a narrow contact-aureole (Hutchison 1970).

UPLIFT MODEL

Albarède (1976) calculated possible paths of uplift in order to interpret petrological data from the Massif Central similar to those from the Khtada Lake Metamorphic Complex. Several of his calculated uplift curves could be used directly to interpret the results summarized above. He showed, using reasonable assumptions regarding heat flow and the content of heat-producing elements in the metamorphic rocks, that an uplift path could pass through two petrologically constrained P–T points in the Massif Central, if uplift of presently exposed rocks commenced from near the base of the crust and terminated when they were about 5 km from the surface. Later and slower uplift unroofed the remaining 5 km.

The uplift curve calculated by Albarède (1976) that best fits the petrological and fluid-inclusion data for the CGC is shown in Figure 7. It assumes a constant rate of uplift of 2 mm/yr beginning at 35 km and terminating at 5 km. The ages show the time required to pass through specific P–T conditions starting at inception of uplift, which is assumed in Albarède's model, to be at 35 km and 800°C. This starting point is consistent with the petrological constraint that the granulite-facies rocks of the KLMC were initially in the kyanite stability-field (Fig. 2), although, strictly speaking, at 800°C uplift would have begun from deeper than 35 km.

Figure 8 shows the assumed initial geothermal gradient and the final geothermal gradient after achieving steady state at infinite time, used in Albarède's calculations to determine the uplift curve in Figure 7. A rather high (40°C/km) assumed initial gradient at the surface was necessary in order to have temperatures at the base of the crust (~35 km) slightly higher (~800°C) than the temperature of the sillimanite–cordierite gneiss at locality 2 (750°C). This would be reasonable for an active plate margin, perhaps in the root zone of an island arc. An initial distribution of radioactive elements that has a high heat productivity near the surface decreasing exponentially to very little at the base of the crust is also an essential assumption but one that is reasonable (Blackwell 1971, Lachenbruch 1970). The lower steady-state thermal gradient in the model after uplift is based on the assumption that the radioactive heat sources were removed by erosion during uplift. It is consistent with measured thermal productivity of presently exposed rocks in the CGC (~1.5 heat generation units: W. Gosnold, written comm., 1975), and with present-day heat flow in the Sierra Nevada batholithic complex.

The constraint of later, slower uplift from 5 km to the surface, used in Albarède's models, is an appropriate boundary condition to apply to the Coast Mountains. Fission-track data obtained on apatite and zircon by Parrish (1980) demonstrate a post-Miocene uplift of about 4 km at a rate of 0.25 mm/yr, which is consistent with geological arguments for an episode of post-Miocene uplift (Monger *et al.* 1972).

The curve on Figure 6 shows the best fit of the uplift curve to the age dates on the Quottoon pluton. Uplift would have commenced at ~62 Ma, given the 35 km and 800°C at inception of uplift, in order for temperatures to have cooled to the time of setting of the Rb–Sr whole-rock age, and the K–Ar blocking age of hornblende. Cooling below ~400°C (at ~48 Ma: Fig. 8) would be nearly isobaric at ~4 km to ~45°C and must have passed through the younger dates obtained by Harrison *et al.* (1979).

There is good internal consistency between the inferred ages of thermal events, the petrological and fluid-inclusion constraints, and the model uplift curve, assuming that uplift of the rocks presently exposed began at 62 Ma from 35 km, 800°C and terminated at ~5 km during the Eocene. The rocks would have cooled to 700°C by 10 Ma after uplift began; according to the model, this would be at 52 Ma, within the range of error for the 51 Ma Rb–Sr whole-rock date. The andalusite veins would have formed 13.5 Ma after uplift began, which would imply their age to be 48.5 Ma. This is remarkably close to, and within error of, the 48.8 Ma date for late plutons in southeastern Alaska that were correlated with the andalusite veins and the Ponder pluton. This correlation is independent of the cooling dates, excluding the Rb–Sr whole-rock date, shown on Figure 8.

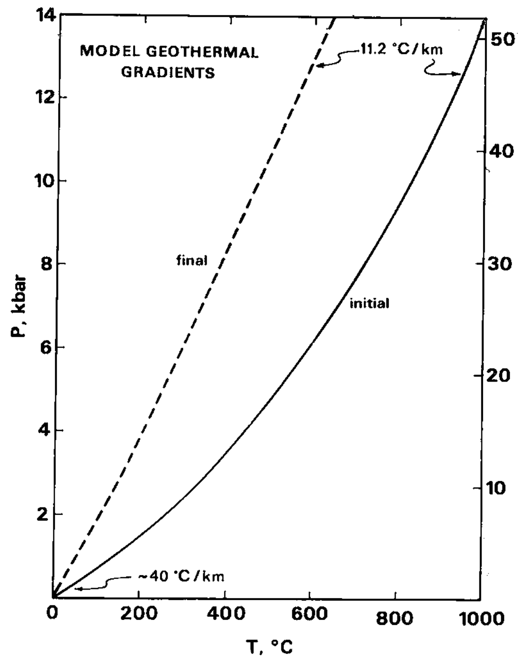

FIG. 8. Geothermal gradients used in uplift model (from Albarède 1976, model B; his Fig. 3). The initial steady-state geothermal gradient assumes that the concentration of heat-producing elements decreases with depth from the surface. The final gradient, steady-state after infinite time, assumes that these higher near-surface concentrations were removed by erosion during uplift. Both gradients assume the same heat contribution from the mantle. The initial gradient is also constrained to pass through 800°C at 35 km, the assumed initial conditions prior to uplift.

DISCUSSION

It is useful to point out and review some of the uncertainties in arriving at the proposed uplift model. First, there is a combined error of approximately ± 1 kbar in the location of the CO_2–H_2O isochores based on the assumptions used in calculating isochores for CO_2+H_2O fluids (Holloway 1981) and on uncertainties in the determination of the molar volumes of the fluids in inclusions (Burruss 1981). There are comparable uncertainties in placing the garnet–

plagioclase and garnet–cordierite equilibria curves in P–T space (Ghent *et al.* 1979, Selverstone & Hollister 1980). The thermal models of Harrison & Clarke (1979) and Albarède (1976) differ significantly in assumed initial and final thermal constants; clearly, the models are sensitive to choice of the initial thermal constants. The ranges of blocking temperatures for the isotopic systems shown in Figure 6 are subject to change. The arguments that the CGC began its uplift history from great depth (> 35 km) are based only on geological extrapolation across a major tectonic discontinuity (Crawford & Hollister 1982) and the occurrence of the kyanite inclusions in garnet. The age dates used to compare the uplift curve to the petrological and fluid-inclusion data also involve extrapolation from localities north and south of the study area, where similar rock-associations to those within the study area occur. More dates from the CGC are needed to support the interpretation that the cooling history of the CGC is similar to that of the Quottoon pluton. The andalusite veins could be directly dated; this would be one test of the proposed model.

If one excludes the occurrences of high-pressure metamorphic rocks west of Work Channel lineament (Fig. 1), a model of heating the CGC at 5–6 kbar from a P–T point in the stability field of kyanite to the 5 kbar, ~800°C metamorphic conditions by multiple intrusion of tonalite to gabbro plutons, dykes and sills accompanied by uplift could be accommodated to fit the data. This could be tested by studying the metamorphic history of a portion of the CGC that does not have substantial late Eocene magmatic intrusion. Such an area may occur in the north-central portion of the CGC shown in Figure 1, and is presently being studied by M.L. Hill (Princeton Univ.).

The uplift model used here differs from the model of Harrison & Clarke (1979) for interpreting the same age data. They interpreted the cooling dates (Fig. 6) of the Quottoon pluton to be the result of cooling by conduction into country rock that had cooled from the granulite-facies metamorphic conditions prior to intrusion of the pluton. This model is not supported by the petrological and fluid-inclusion data summarized in the present paper; these data were not available to Harrison & Clarke (1979) when they published their interpretation.

The uplift rate of 2 mm/yr proposed for the Coast Mountains, involving unroofing and erosion of ~30 km of rock between 62 and 48 Ma, is faster than the maximum uplift rate proposed for the Alps (1 mm/yr: Clark & Jäger 1969) and for the Himalayas (0.8 mm/yr: Sharma *et al.* 1980) However, it is comparable to the 5.5 mm/yr post-Pliocene uplift rate proposed for Taiwan (Li 1976), and to the 1.5 to 22 mm/yr Recent rates proposed for South Island, New Zealand (Adams 1980). The tectonic and petrological consequences for so much uplift in such a short period of time are profound, not least of which is the question of the process leading to this uplift and the question of what now makes up the ~30 km of continental crust believed to underlie the present Coast Mountains (Roddick & Hutchison 1974). These two questions may be related, but any answer at the present time would be speculative. Note that within the area of study, there is a single 5-km-thick hypersthene-bearing gabbroic sill (the Kasiks Sill: Fig. 1); several smaller sills mapped above this one have an aggregate thickness of at least 2 km. It is tempting to speculate that underplating of the crust by magma of basaltic or andesitic composition, coming from the mantle, may have led to the uplift and to the added continental crust. Intrusion of basaltic magma may also have contributed some of the heat required for formation of the high-temperature metamorphic rocks occurring in the area. If this were the case, the P–T–time curve could have begun from a lower temperature than the 800°C used in the calculations, and temperature would have increased to the granulite-facies conditions during uplift.

The proposed uplift would have produced a minimum of 3 x 10^5 km^3 of clastic sediment during the Paleocene to Eocene (Hollister 1979). This sediment would have been deposited as a clastic wedge onto the Pacific Ocean floor, because there is little clastic sediment of this age to the east of the Coast Mountains. Post-Eocene movement of the Pacific plate would have transported the sediment to southern Alaska. At present, deformed Paleocene to Eocene clastic sediments that were not derived locally, do occur in southern Alaska, in the Orca Group (Helwig & Emmet 1981) and in Yakutat Bay (Plafker *et al.* 1980). Clearly the proposed uplift area could be the source of these sediments, as suggested by Hollister (1979).

Acknowledgements

This study was supported by NSF grants DES75–03259, EAR80–07454, and EAR78–05887; substantial logistic support provided by the Geological Survey of Canada is gratefully acknowledged. R. Burruss, C. Kenah, C. Kulick, J. Selverstone and J. Sisson assisted with the collection and reduction of data. M. Sherwood helped with the testing of the applicability of Albarède's calculations to the Central Gneiss Complex. J. Moore discovered the andalusite in the late veins during a visit to the area. I am especially grateful to M.L. Crawford and V.B. Sisson for thorough and constructive reviews. E. Ghent and R. Tracy also provided helpful comments on an early version of the manuscript.

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Received July 1981, revised manuscript accepted March 1982.