SKB TECHNICAL REPORT 95-18

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November 1995

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Information on SKB technical reports from1977-1978 (TR 121), 1979 (TR 79-28), 1980 (TR 80-26), 1981 (TR 81-17), 1982 (TR 82-28), 1983 (TR 83-77), 1984 (TR 85-01), 1985 (TR 85-20), 1986 (TR 86-31), 1987 (TR 87-33), 1988 (TR 88-32), 1989 (TR 89-40), 1990 (TR 90-46), 1991 (TR 91-64), 1992 (TR 92-46), 1993 (TR 93-34) and 1994 (TR 94-33) is available through SKB.

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Gråbo, November 1995

Keywords: foreland basin, Caledonian, stable isotopes, limestones

ABSTRACT

The Phanerozoic rocks present on the Fennoscandian Shield are dominantly of Cambrian to Silurian age. They represent a relatively thin sedimentary cover (less than 500 metres). The question is: why do we not see any remnants of younger sedimentary rocks? Did they ever exist, or have they been eroded, transported and redeposited elsewhere?

$\delta^{18}O$ and $\delta^{13}C$ analyses of Ordovician limestones from different places in Sweden and from the Oslo region in Norway show modification of their original marine signature according to the $\delta^{18}O$–values, while the $\delta^{13}C$–values generally are typical for marine limestones. In some cases the modifications can be explained by intrusions of dykes or by metamorphic events, but in most areas the redistribution of the oxygen isotopes indicates burial diagenesis. From a number of published investigations, raised temperatures at the present surface during the Late Palaeozoic, are indicated by different temperature indicators. We suggest that these increased temperatures were due to a sedimentary cover of mainly Devonian sediments deposited on top of the Cambrian–Silurian sequence. This palaeo–cover caused raised temperatures at the present rock surface, as shown by $\delta^{18}O$ values, as well as vitrinite reflectance, conodont colour alteration index (CAI) and by illite crystallinity in the Cambro–Silurian rocks.

In the Proterozoic basement, annealing of fission tracks in apatite and mobility of radiogenic lead also give evidence of increased temperatures. A model where the thickness of the Upper Palaeozoic cover of the Caledonian foreland is 2–4 kilometres thick is suggested. This cover mainly consisted of Late Silurian–Devonian erosional products from the Caledonides the latter which formed during a Silurian continent–continent collision. A major Permian to Triassic uplift and erosion reduced the cover significantly.

SAMMANFATTNING

De rester av fanerozoiska bergarter som finns bevarade på den Fennoskandiska skölden är av kambrisk till silurisk ålder och representerar ett relativt tunt sediment täcke (<500 m). Varför ser vi inga rester av yngre sedimentbergarter på skölden? Är de fullständigt borteroderade och avsatta på annan plats eller har de aldrig existerat?

$\delta^{18}O$/ $\delta^{13}C$–analyser av marina ordoviciska kalkstenar från olika platser i Sverige och från Oslo–området i Norge, indikerar en post–depositionell omvandling vilken kan vara resultatet av att de varit överlagrade av ett tjockt sedimenttäcke. För att klargöra om denna teori är riktig eller ej har vi gjort en litteratur–sammanfattning av resultaten från olika studier av palaeotemperaturindikatorer.

Denna sammanfattning omfattar således egna $\delta^{18}O$/ $\delta^{13}C$ analyser samt publicerade arbeten om vitrinitet, "conodont alteration index" (CAI), illitkristallinitet och organisk mognad hos de kambro–siluriska sedimenten samt fissionspårsanalyser och mobilitet av radiogent bly i urberget. Vi finner starka indikationer på att paleotemperaturer i intervallet 80 till 125°C rådde under sen–palaeozoisk tid och tolkar detta som resultatet av ett flera km tjockt täcke av sen–siluriska och devonska sediment. Dessa sediment var huvudsakligen erosionsprodukter från den kaledonska bergskedjan, vilken bildades under silurisk tid då den Nordeuropeiska kontinenten kolliderade med den Nordamerikanska. Erosionsprodukterna bildade ett 2–4 km tjock täcke av molassediment på det kaldonska förlandet. En huvudsakligen permisk–triasisk erosion/"uplift" reducerde sedimenttäckets tjocklek avsevärt, och under tertiär eroderades det sannolikt bort helt och hållet.

CONTENTS

1 INTRODUCTION

It is generally accepted that most of the crystalline basement of the Fennoscandian Shield has been covered by Lower Palaeozoic deposits. As reconstructed from remnants in Sweden, this sedimentary cover was generally less than 500 m thick there (e.g. Lindström et al. 1991). Upper Palaeozoic sediments have not been documented in the Swedish part of the Fennoscandian shield, although Devonian to Permian sediments are present e.g. in the SE part of the Baltic Sea (Ahlberg 1986; Grigelis 1991). Suggestions about a Phanerozoic cover thicker than 500 metres have been made by Buchardt & Nielsen (1985) concerning the Bornholm Island (southern Baltic Sea), in the Dalsland province (SW Sweden) by Zeck et al. (1988) and in Hunnedalen, SW Norway by Andriessen (1990). Moreover, regional sedimentation of Upper Palaeozoic sediments on the platform can be suspected since vast amounts must have been eroded from the Caledonian mountain range during the Upper Palaeozoic (cf. Andriessen & Bos 1986). Progressive spread of terrestrial Silurian–Devonian "Old Red" molasse sediments southwards across SE Scandinavia was suggested by Basset (1985).

Within the present study Ordovician limestones, from different parts of Sweden and the Oslo area in Norway, have been analysed for their stable isotope composition (oxygen and carbon). Also, carbonates from some Palaeozoic fracture/infillings and breccias from the west coast of Sweden were analysed. This study uses the $\delta^{18}O$ and $\delta^{13}C$ results together with literature documentation of vitrinite reflectance, conodont alteration indices, illite crystallinity, lead mobility and fission track studies of apatites from basement granites, to estimate the thickness and extent of a Phanerozoic palaeo-cover on the Scandinavian part of the Fennoscandian Shield.

2 GEOLOGICAL SETTING

The bedrock of Sweden can be subdivided into three main units; the Precambrian crystalline basement, the Phanerozoic sedimentary cover and the Caledonides.

The Precambrian crystalline basement of the Fennoscandian Shield extends from the Kola peninsula in the NE to the Caledonides in the west and to the Tornquist–Teisseyre Zone in the south (Figure 1). The basement becomes successively younger from the Archean nuclei in the NE to the Meso– to Neo–Proterozoic in the SW. The shield is bounded to the SE by the Phanerozoic cover rocks of the Russian platform.

The Proterozoic orogenic evolution of the Fennoscandian Shield was terminated by the emplacement of Late Sveconorwegian (c. 0.9 Ga) granites in SW Sweden and SE Norway, and was succeeded by Neo–Proterozoic uplift and erosion of the shield. During the Vendian/Cambrian the peneplanation was completed. The crystalline basement was partly covered by debris.

The Lower Palaeozoic sequences preserved in different parts of Sweden typically constitute incomplete columns, with hiatus and conglomerates, indicating intermittent erosional conditions (Lindström et al. 1991). The extensive distribution of clastic dykes of sand– and siltstones of Early Cambrian age and also fractures filled with Middle and Late Cambrian black (alum) shale (e.g. Samuelsson 1975; Bergman 1981; Greiling et al. 1988) indicates a wide distribution of Cambrian deposits.

From Arenig to Caradoc, marine sedimentation took place on the Baltoscandian platform. The sea was generally deepest close to the present Caledonian front to the west and more shallow in Estonia, White Russia and Ukraine (Bruton et al. 1985). The sedimentation environments are divided into three different confacies belts (Jaanusson 1976). The Scanian Belt in the south (including Scania, Halle and Hunneberg in Västergötland and the Oslo region) is characterised by sediments rich in clay. The Central Baltoscandian Confacies Belt (including Kinnekulle and Billingen in Västergötland, Närke, Östergötland, Öland, Siljan and the autochthonous sediments of Jämtland; cf. Figs. 1 and 2), is characterised by a thin limestone cover formed during a very slow deposition of carbonate. The limestones are rich in fossils of the Baltoscandian shelly fauna type. The third belt is the North–Estonian Belt including Gotland. This is characterised by sedimentation of grey limestones with a diverse fauna.

The Early Silurian period is characterised by transgression and marine deposits. Silurian sediments are recorded in many but restricted areas in Sweden and consist mainly of shales and reef limestones. However, the sequences left are incomplete, except for on Gotland (cf. Bassett, 1985).

Late Silurian to early Devonian collision and extensive subduction of Fennoscandia under Greenland/Laurentia caused large-scale stacking and eastward thrusting of nappes, forming a prominant mountain chain, the Scandinavian Caledonides. It represents the result of a complex history of outboard and marginal terrane development, assembly and Early Caledonian (Finnmarkian c. 520–480 Ma) accretion to Fennoscandia. The piling of these terranes onto the margin of Fennoscandia culminated with the Himalayan-type Scandian continent–continent collision (c. 430–390 Ma) and underthrusting of Fennoscandia. Dismembered units of exotic and indigenous terranes were thrusted far eastwards, over the Fennoscandia foreland of crystalline basement and miogeoclinal cover. Seismic reflection profiles show a gently hinterland-dipping décollement, with the detachment mainly of cover rocks in the front. Towards the interior of the orogen the crystalline basement is progressively deeper involved in thrusting and late normal faulting, down to at least mid-crustal levels (Hurich et al. 1988; Palm et al. 1991).

Initiation and progressive spread to the east and southeast of the terrestrial Old Red Sandstone facies across SE Scandinavia occurred during the late Silurian and Early Devonian. This involved molasse sedimentation examplified by the Ringerike sandstone in the Oslo region (Basset 1985). The non-marine fluvial Röde Formation of suggested Wenlock age within the Lower Allochthon of Jämtland was interpreted as distal deposits of detritus from rising Caledonides in the west (Bassett et al. 1982). The Devonian in Sweden is documented only in Scania, e.g. the Öved sandstone, belonging to the Old Red Sandstone deposits. Several remnants of intra-orogenic Devonian extensional basins are found in Norway (Steel et al. 1985). Thick deposits of terrigeneous Devonian rocks are described from a WSW–ENE oriented basin in the Baltic countries and the southeastern Baltic sea (Grigelis 1991; Kurss 1992). In central Lithuania the Devonian rocks reach over 1000 m thickness. Structures and fossils indicate continental, lagoonal and shallow marine environments. The stratigraphy is dominated by crossbedded sandstones, interpreted as mainly fluviatile-deltaic deposits. The maturity of the terrigeneous material generally increases upwards. Dolomites and terrigeneous sand- and siltstones with dolomitic matrix and subordinate clays occur in the middle and upper Devonian. The lower Devonian sediments were deposited under regressive conditions, changing into transgression during the Middle to Upper Devonian. Reconstructed palaeocurrent directions indicate that the main transport was from N or NNW, i.e. from the eastern slopes of the Scandinavian Caledonides (Kurss 1992).

No other sedimentary left overs from the Devonian, Carboniferous and Permian periods have been found in Sweden. However, Carboniferous microfossils are redeposited in Jurassic sediments in Scania, indicating the former existence of Carboniferous deposits in southern Sweden (Lindström et al. 1991).

The transition between the Carboniferous and the Permian is manifested as a dramatic period in southern Scandinavia. Both extensive volcanism and rifting characterize this period. In the Oslo area extrusive rocks were formed during the time interval 300 Ma to 240 Ma ago (Sundvoll et al. 1990). At this time also N–S trending dykes were formed which can be traced from Oslo in Norway along the Swedish west coast to Göteborg in the south (e.g. Samuelsson 1971). Permian sediments are found in the Oslo Graben. They are fossiliferous, continental deposits with intercalating lava beds of the Oslo Igneous Province. Early Permian dolerite sills (Priem et al. 1968) are found at the tops of the hills of Lower Palaeozoic rocks in the Västergötland province (e.g. Halleberg, Hunneberg, Kinnekulle and Billingen), SW Sweden.

Mesozoic and Tertiary rocks are found only in the down faulted southwestern part of the Scania province, S Sweden, except for Cretaceous sediments which are deposited on the deeply eroded crystalline basement in NE Scania and in S Halland (Figure 1).

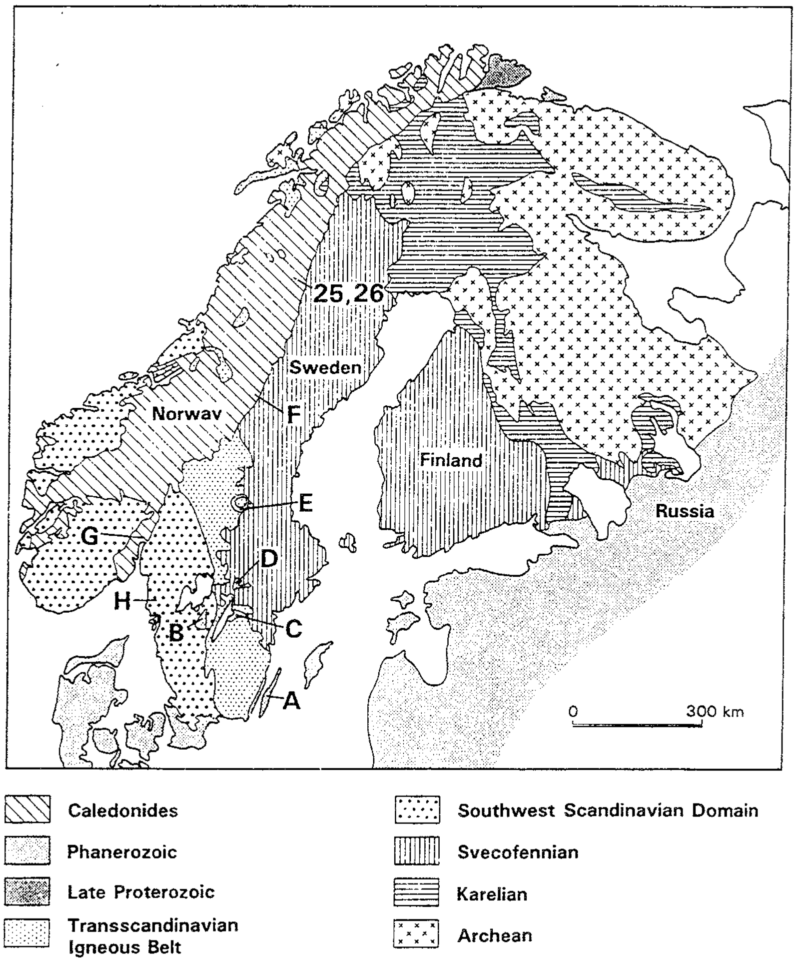

Fig. 1. *Lithological units of the Fennoscandian shield. Symbols A–F and (Fig.2) and G (Fig.4) refer to sampled areas. Localities 25 and 26 refer to Caledonian Upper Allochthon samples "Slätdal" and "Gilliks" resepectively, (Tab.1).*

3 SAMPLE AREAS FOR STABLE ISOTOPE ANALYSES

The samples collected for stable isotope analyses were taken from three different environments. These are authochthonous limestones from the Ordovician strata of the Phanerozoic cover, Ordovician and Silurian limestones deformed during the Caledonian orogeny and Cambrian fracture fillings (Figure 2).

The autochthonous, Ordovician cover rocks were sampled on the island of Öland (Lundegårdh et al. 1985), in the Östergötland (Wikman et al. 1982), Västergötland (Lindström et al. 1991; Wikner et al. 1991) and Närke (Andersson et al. 1982) provinces and from the impact structure infillings at Siljan in the Dalarna province (e.g. Bottomley et al. 1978; Åberg et al. 1989; Wickman, 1980) (Figure 2).

The allochthonous Ordovician and Silurian samples were collected in the Lower Allochthon of the Jämtland province, within the Permian Oslo Rift in Oslo, as well as within the far transported Upper Allochtonous Gilliks and Slätdal Formations (Zachrisson 1969) (Figures 2 and 4).

The sampled Cambrian fissure infillings (Samuelsson 1975, 1985; Greiling et al. 1988) are all from the Bohuslän province, mainly from the vicinity of Göteborg (Figure 2).

Samples for stable isotope analyses are bulk samples of matrix and cement calcite (fossils excluded) except for the samples from the Oslo region where corresponding fossil/matrix–cement samples were prepared.

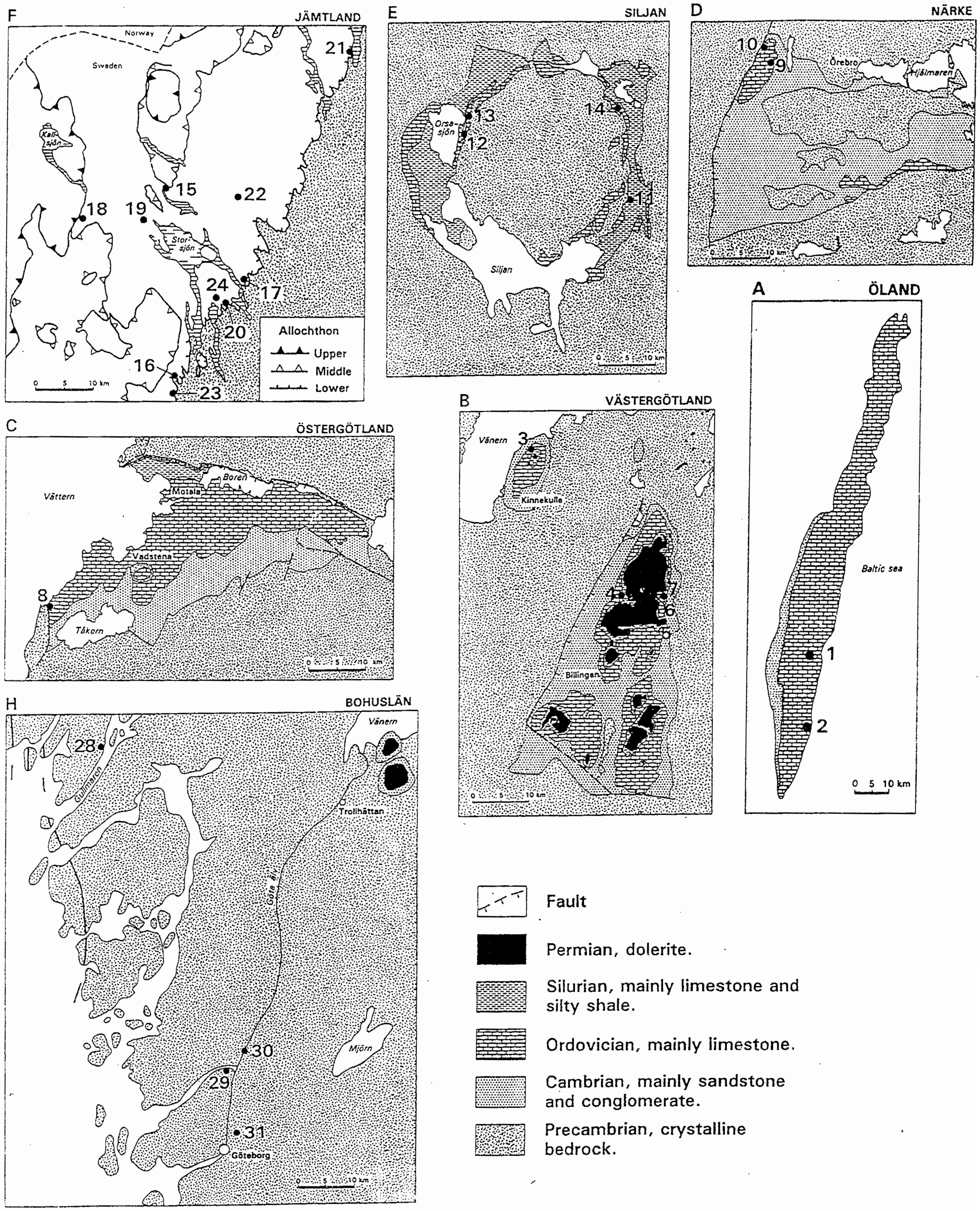

Figure 2. *Sampled areas (cf. Fig.1). The sample sites are marked 1–29 refering to numbers in Table 1.*

4 ANALYSES OF $\delta^{18}O$ AND $\delta^{13}C$ IN PALAEOZOIC LIMESTONES AND FRACTURE INFILLINGS

$\delta^{18}O$ and $\delta^{13}C$ analyses of the carbonates have been carried out at two laboratories; Institutt for Energiteknikk (IFE), Kjeller, Norway and Department of Marine Geology, Göteborg University, Sweden. They use similar conventional techniques and the results have been shown to be comparable.
The results are related to a standard as follows:

$\delta^{13}C(\text{sample}) = [X(\text{sample})-X(\text{stand})]/X(\text{stand}); X={}^{13}C/{}^{12}C.$

The same equation is valid for the ${}^{18}O/{}^{16}O$ fractionation expressed as $\delta^{18}O$. The standard used for the carbon and oxygen analyses was related to the PDB standard. The accuracy of the $\delta^{13}C$ and the $\delta^{18}O$ analyses is ±0.1 o/oo.

4.1 RESULTS AND DISCUSSION

The results of the oxygen and carbon isotope analyses are shown in Table 1. Duplicate analyses were made for most of the samples which show a spread in $\delta^{18}O$ of maximum 1.0 o/oo and in $\delta^{13}C$ of maximum 0.7 o/oo. Based on chemical analyses (Shaikh et al. 1989), most of the samples are limestones with MgO contents < 1.0%, with a few exceptions within the range 1.0 to 2.0%.

Table 1. $\delta^{18}O$ and $\delta^{13}C$ in Ordovician limestones and dikes.

Sample location	$\delta^{18}O$ o/oo (PDB)	$\delta^{13}C$ o/oo (PDB)
Ordovician limestones		
A: Öland:		
1. (H) Vickleby	−6.0	+0.8
(H) -"-	−6.2	+0.5
2. (H) ' Segerstad (d) 7.70 m	−6.4	+0.6
(L) ' -"- (d) 13 .47 m	−7.0	+0.3

(L) ' –"– (d) 16.70 m	–7.4	+4.2
B: <u>The province of Västergötland:</u>		
3. (He) Hällekis (q)	–6.6	–0.6
(He) –"–	–6.4	–0.8
4. (L) Varnhem	–6.2	+3.1
(L) –"–	–6.4	+3.8
5. (L) Skulltorp	–6.7	- 0.2
(L) –"–	–6.1	- 0.3
6. (D) Vomb (r)	–7.7	+0.5
(D) –"–	–8.1	+0.5
7. (D) Gullhögen (q) upper	–7.5	+0.4
–"–	–7.5	+0.4
(D) Gullhögen (q) middle	–6.6	+0.9
–"–	–7.2	+1.1
(L) Gullhögen (q) lower	–7.8	–1.0
–"–	–6.8	–0.7
C. <u>The province Östergötland:</u>		
8. (L) ' Bårsta (d) 1.08 m	–7.2	–0.2
–"– 2.57 m	–8.0	–0.6
–"– 5.30 m	–7.0	–0.6
D. <u>The province of Närke</u>		
9. (L) ' Örsta (d) 3.52 m	–7.1	+0.3
–"– 6.20 m	–7.3	+0.3
–"– 7.86 m	–8.0	- 0.1
10. (L) ' Filipshyttan (d) 1:26	–7.1	- 0.1
E. <u>The Siljan area:</u>		
11.* Solberga (d) 43 m	–5.2	+2.2
* –"– 142 m	–4.6	+1.3
* –"– 207 m	–6.7	+0.7
12. (H) ' Kårgärde	–6.6	+1.3
(S) –"–	–8.3	+1.0
13. (Fo) ' Enån	–7.0	+1.3
14. (F) ' Fjäckan	–7.2	+0.4
(Sl) ' –"–	–5.8	+0.3
F. <u>The province of Jämtland:</u>		
15. (B) Berge	–8.4	+0.4
16. (F) ' Storhallen	–6.3	+0.8
17. (Fo) ' Lundbomberget	–7.9	+1.1
(S) ' –"–	–8.8	+1.6
18. (B) ' Ristafallet	–8.9	+0.1
19. (B) ' Mattmar	–9.5	+0.9
20. (D) ' Lappgruban	–8.6	–0.4

21. (H) ' Järvsand	−10.9	+0.6
22. (Sl) ' Husås	−10.3	+0.1
23. (F) ' Gammalbodberget	−6.7	+0.9
24. (Sl) ' Stengärde	−10.1	−1.2
The province of Västerbotten:		
25. Slätdal	−16.6	−1.3
26. Gilliks	−13.6	−2.9
G. The Oslo region:		
27. Hovedöya (cement and matrix)		
Dist. from dike 0.1 m	−16.4	−6.5
1.0 m	−15.3	+2.7
5 m	−15.4	+2.9
10 m	−15.0	+3.2
40 m	−13.9	+3.4
Hovedöya (fossile)		
Dist. from dike 0.1 m	−14.7	−1.2
1.0 m	−12.8	+4.0
5 m	−10.4	+2.6
10 m	−8.5	+3.6
−"−	−8.6	+3.8
40 m	−9.5	+3.5

Palaeozoic breccias and dikes:

H. Bohuslän:		
28. Gullmaren breccia	−12.8	−3.0
29. Kungälv breccia	−16.6	−0.3
30. Nol limestone dike	−14.1	−1.7
31. Hjällbo limestone dike	−12.9	−9.4

(d) = sample from drillcores, (q) = sample from quarry, (r) = sample from road cut
' = sample provided by Lars Karis , Geological Survey of Sweden
* = sample from the Deep Earth Gas project.

Different sequences of Ordovician limestones are referred to as follows: Lower Ordovician; (L) = Lanna, (H) = Holen. Middle Ordovician: (S) = Segerstad, (He) = Hällekis, (Fo) = Folkeslunda, (F) = Furulund, (D) = Dalby. Upper Ordovician; (Sl) = Slandrom. Lower Silurian; (B) = Berge

4.1.1 STABLE ISOTOPE COMPOSITION OF A PALAEO-OCEAN

The $\delta^{18}O$ of carbonate minerals is a function of the temperature and the isotopic composition of the water during precipitation.

The stable isotope composition of the Palaeo-ocean has long been debated (e.g. Marshall 1992; Hoffman et al. 1991; Wadleigh & Veizer 1992). Most authors argue for lower $\delta^{18}O$-values, in the Early Palaeozic ocean than in the modern ocean. Thus, based on Early Palaeozoic articulate brachiopods, Wadleigh & Veizer (1992) suggest $\delta^{18}O$-values of -2 to -3 o/oo during the Ordovician. However, the highest values recorded in their study, represents ocean composition close to 0, and are found in Ordovician and Silurian fossils in shales. Shales are known to be relatively impermeable and should thus retain the most original signature. Furthermore, post-depositional processes which result in enriched $\delta^{18}O$-values are unlikely.

Based on analyses of brachiopods and cement of the Boda limestone (infilling of the Siljan impact crater, central Sweden), a drastic change in $\delta^{18}O$ in the Ordovician ocean was interpreted as due to a glaciation (Middleton et al. 1991). Such changes are reported for the Weichselian glaciation when the ocean was enriched in $\delta^{18}O$ with approximately 1 o/oo during its maximum (Fairbanks, 1989). Oceanic fluctuation of approximately 1.5 o/oo $\delta^{18}O$, between times of maximum glaciation and times of total deglaciation, can be expected (Marshall 1992). We therefore suggest the $\delta^{18}O$-values of the Ordovician ocean to be within the 0 to -2 o/oo range (cf. also Gao 1993).

During the Ordovician, Fennoscandia drifted from about 50°S to 20°S (Perroud et al. 1992). We therefore assume a ocean water temperature of 10 to 20°C.

The $\delta^{13}C$-values in the carbonate minerals are a function of dissolved bicarbonate in the water from which they precipitated and in contrast to the oxygen isotopes the fractionation of carbon isotopes is not very sensitive to temperature changes. However, $\delta^{13}C$ can be affected by so called "vital effects". The result can be carbonate enriched in $\delta^{13}C$ (where ^{12}C has been withdrawn from the marine water by biologic activity) or depleted in $\delta^{13}C$ (where organic material has been oxidised and incorporated in the carbonate system). Typical $\delta^{13}C$-values for marine limestones are in the range of -1 to +2.5 o/oo.

Using the fractionation factor given by O'Neil et al. (1969), estimated temperatures, and isotope ratios, the range of isotopic compositions of unaltered Ordovician marine limstones was calculated and is shown in figure 3.

4.1.2 STABLE ISOTOPE RATIOS OF ANALYSED LIMESTONES

All results from the analysed Ordovician limestones are plotted in the diagram of Figure 3. Most of the samples show $\delta^{13}C$ values in accordance with those obtained for marine carbonates. The $\delta^{13}C$ values of the Lower Ordovician limestones typically range between −1.0 to +0.5 o/oo whereas the $\delta^{13}C$ values for Middle and Upper Ordovician samples mostly range between 0.0 to +1.5 o/oo. Oxygen isotopes have been redistributed as exemplified by the samples from Öland, Östergötland and Närke (A, C and D in Figure 2). The isotope values of limestone samples from these areas cluster within an interval of $\delta^{18}O$ = −8 to −6 o/oo (with values of −7 to −6 o/oo for samples from Öland, and values between −7 and −8 o/oo for samples from Östergötland and Närke). The $\delta^{13}C$–values are within the interval −1 to +1 o/oo except for one sample on +4.2 o/oo. As no disturbances like volcanic activity, dyke intrusions or extraterrestrial impacts are known from these areas a regional, low–temperature, post–depositional redistribution of isotope ratios is suggested. If meteoric water circulation caused the redistribution of the oxygen isotopes in the limestones a larger spread in the $\delta^{18}O$–values should be expected due to various isotopic ratios of the groundwaters in different areas and also due to the variation in water/rock ratio. In contrast, the observed values are typical for diagenetic changes during mainly closed to semi–closed conditions (cf. Marshall 1992) i.e. reequilibration at raised temperatures by a mainly marine pore water. This is in accordance with Buchardt & Nielsen (1985) who reported a similar stable isotope pattern from the Lower Palaeozoic marine limestones of Bornholm as that in the Närke, Öland and Östergötland limestones. These are characterised by reequilibrated $\delta^{18}O$–values and more or less well preserved original $\delta^{13}C$–values. They interpreted the reequilibration as due to burial metamorphism, assuming redistribution of oxygen isotopes between limestone and marine pore water and calculated temperatures of about 90°C, representing a post–Silurian cover of at least 3 km. Using the same approach (fractionation factors for calcite/water according to O'Neil (1969) and assuming Ordovician Sea $\delta^{18}O$ = 0 to −2 o/oo SMOW) on the limestones from Öland, Östergötland and Närke results in temperatures from 34° to 58 °C.

Notably, most of the $\delta^{18}O$–values from Västergötland (Billingen and Kinnekulle; B in Figure 2) are similar to those from Öland, Närke and Östergötland. However, the oxygen isotope ratios from Billingen are decreasing when approaching the dolerite sill (which is c. 40 metres away from the nearest sample). The isotope ratio of the farthermost sample corresponds well to the values obtained for samples not affected by Permian intrusions. At Kinnekulle the shortest distance between the sampled limestone and the dolerite sill is even larger (c. 80 metres) which resulted in less significant heating of the limestone. The variation of the $\delta^{13}C$– values in the samples from Billingen (−1.0 to +1.1 o/oo) is probably due to environmental differences during the sedimentation.

The eleven limestone samples from sediments of the Lower Allochthon in Jämtland (F in figure 2) show relatively large variations concerning oxygen isotopes ($\delta^{18}O$ = – 10.9 to – 6.3 o/oo; Figure 3). The carbon isotope values are in the interval $\delta^{13}C$ = –0.4 to +1.6 o/oo thus correspond to values expected for marine limestones. Six out of ten samples cluster around –9.5 to –7.9 o/oo in $\delta^{18}O$–values (corresponding to a temperature of 44–76°C if equilibrated with a Palaeozoic ocean water). Two samples show higher values (–6.3 and –6.7 o/oo). These are from the southern part of the area (Storhallen and Gammelbodberget). The three samples from Huså, Stengärde and Järvsand show lower $\delta^{18}O$–values (–10.1, –10.3 and –10.9 o/oo). All Jämtland limestones were most probably covered by the Middle and Upper Allochthons which caused metamorphic overprinting and redistribution of the stable isotopes.

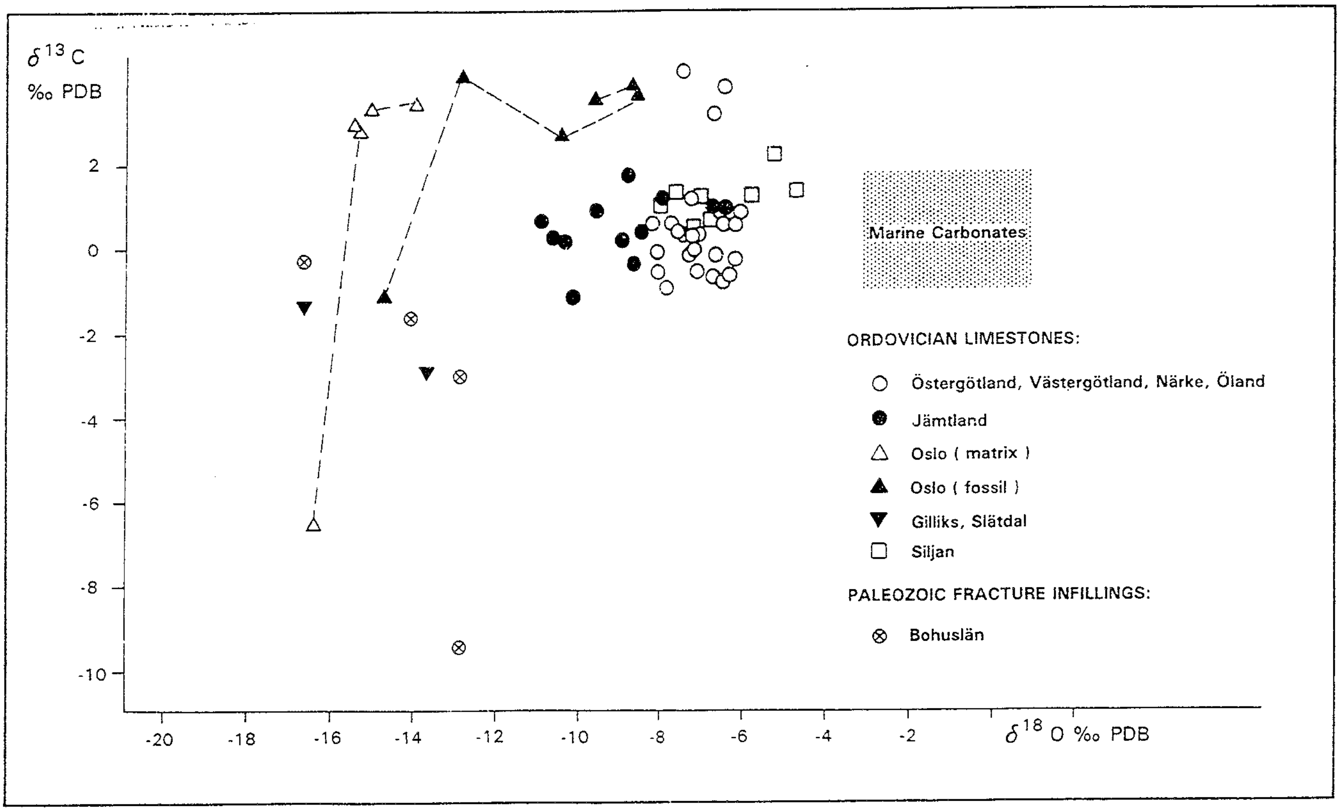

Figure 3. *$\delta^{18}O/\delta^{13}C$ plot of all samples analysed. The values representing unaltered marine limestones are marked with a hatched square. For details, see text.*

The samples from the Upper Allochthon in the Västerbotten province (Gilliks and Slätdal; 25 and 26 in Figure 1) show relatively low oxygen isotope values (−13.6 and −16.6 o/oo respectively). The carbon isotope values are slightly lower than for all the other limestone samples, except for the Oslo samples, although, a marine signature is still indicated. The Gilliks and Slätdal samples belong to the exotic terranes of the Upper Allochthon and are indeed involved in the processes of the Caledonian orogeny. These were transported from an oceanic setting to the west into their present positions and had an overburden of at least the Uppermost Allochthon. They are now in greenschist metamorphic grade. A higher grade of oxygen isotope redistribution in these samples is expected from their tectonic position as compared to the autochthonous limestone sediments on the platform. This is in full agreeement with our results.

The eight samples of Ordovician limestones from the Siljan impact structure (E in Figure 2) show $\delta^{18}O$ ranging from −8.3 to −4.6 o/oo (Figure 3). The relatively large spread in oxygen isotope values can not be related to depth of overburden or geographical position. Instead, differences in water/rock ratios are indicated. Furthermore, input of meteoric water in the tectonised parts, like in steeply dipping fractures and also along bedding planes should be considered. Notably, the redistribution of the stable isotopes within the Siljan impact structure is within the same range or less than observed in the other places investigated. We conclude that the heating caused by the impact event did not significantly affect the stable isotopes. The redistribution of the oxygen isotopes in some of the samples (Solberga 2 samples and Fjäckan 1 sample) is significantly less than in all the other samples analysed. From a number of samples of Boda limestone (Upper Ordovician) Middleton et al. (1991) report $\delta^{18}O$–values of cement with even larger range (−3 to −12 o/oo) of which they attribute the lower values to diagenetic changes and the higher values to be more or less original.

Samples from the Ordovician limestone at Hovedöya, Oslo, which stratigraphically are referred to as the Lower Allochthon, show a significant redistribution of oxygen isotopes. All analyses show $\delta^{18}O$–values which deviate from normal marine signatures. There is a clear relation to the distance from a crosscutting Permian dyke (Figure 4). The decrease in $\delta^{18}O$ and $\delta^{13}C$ close to the dyke is interpreted to be due to circulation of fluids, probably meteoric water, driven by the dyke intrusion. Since the matrix has a higher porosity (higher water/rock ratio) than the fossils it was influenced to a higher degree by fluids. The more extensive redistribution of the oxygen isotopes compared to the carbon isotopes indicates that the former reequilibrates at lower water/rock ratios than carbon isotopes do (cf. Sverjensky 1981; Marshall 1992). It is suggested that the influence of the Permian dyke is negligable already ca 10 metres from the dyke and that the values of c. 9 o/oo at 10 and 40 metres rather reflect the thermal influence from nappes and/or Permian lavas.

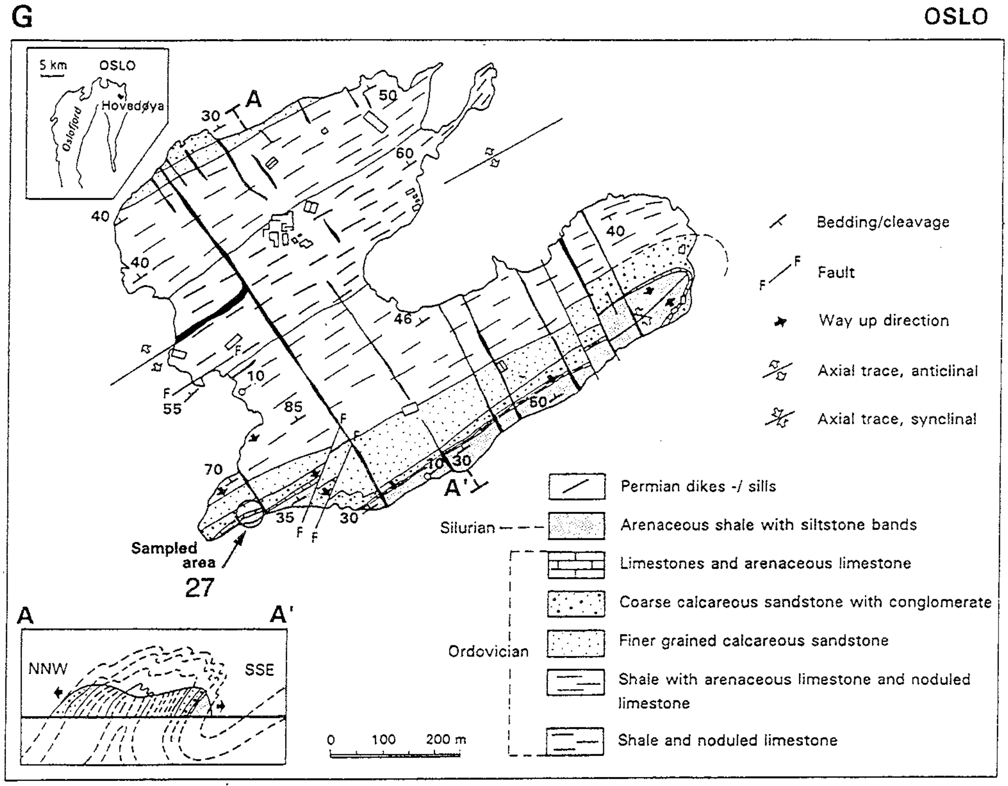

Figure 4a. *Bedrock map of Hovedöya, Oslo and the sample site.*

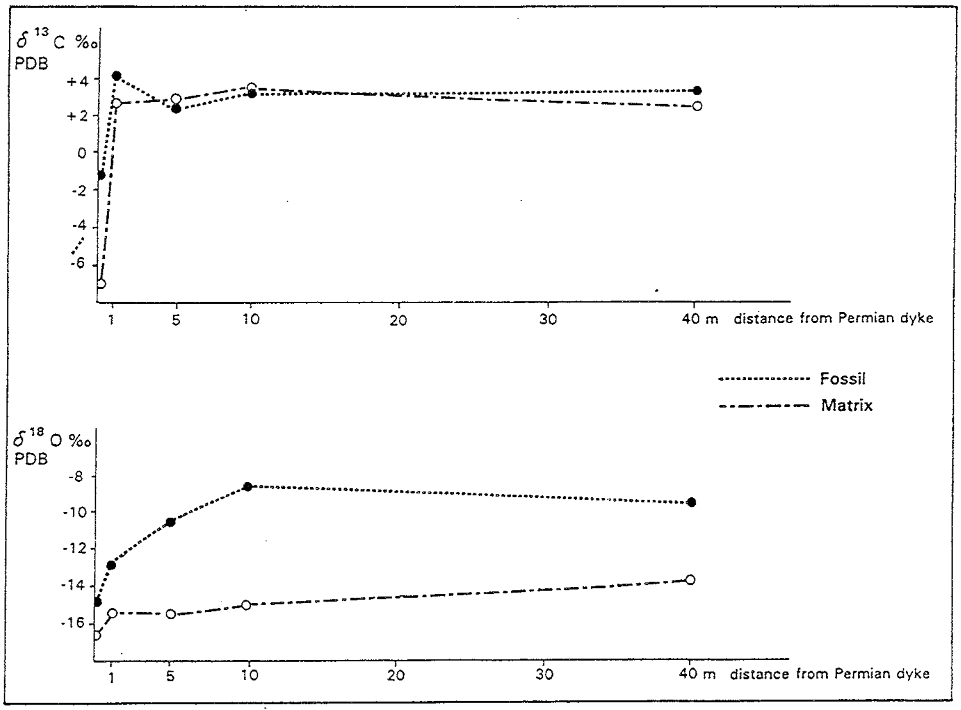

Figure 4b. *$\delta^{18}O$ and $\delta^{13}C$ for samples at different distances from the Permian dike.*

The four samples of Cambrian fracture infillings and breccias (H in Figure 2) are shown to deviate from most of the limestone samples by having lower $\delta^{18}O$ and also lower $\delta^{13}C$ (Figure 3). Of these, two samples (limestone fragment of the Kungälv breccia and the calcite cemented sandstone fracture infilling from Nol) show marine signatures concerning the carbon isotope values although the original $\delta^{18}O$– values are markedly lowered. The Gullmaren breccia, where crystalline rock fragments have been sealed by limestone/Alum shale cement, show similar oxygen isotope signature as that of the Kungälv and Nol samples but has somewhat redistributed carbon isotopes. It can be concluded that none of these samples have retained their original signature. The sample from a fracture filled with sandstone at Hjällbo, Göteborg, shows a typical "meteoric" signature concerning both oxygen and carbon isotopes. The low $\delta^{13}C$–values indicate input of organic carbon, i.e. near–surface conditions.

The temperatures calculated from the $\delta^{18}O$–values of the limestone samples in equilibrium with a marine pore water (assuming a large water/rock ratio) is shown in figure 5. If the water/rock ratio was low and insufficient to reequilibrate the oxygen isotopes in the limestones e.g. due to limited pore volumes, the calculated temperatures will be too low. Furthermore, succeeding low temperature oxygen isotope redistribution caused by a meteoric water may also have affected the isotope ratios. However, despite the large uncertainties in the interpretation of data from bulk samples, the $\delta^{18}O$–analyses all show modification of the their original values. In some areas this is explained by shorter thermal events or allochtonous overburden. However, in other areas this modification must have an alternative explanation. We suggest that burial metamorphism is the probable explanation for this oxygen isotope redistribution.

In order to verify this interpretation and to make more realistic estimates of the maximum temperatures, the obtained data are compared with other temperature indicators. A review, of relevant temperature indicators reported about, is given below.

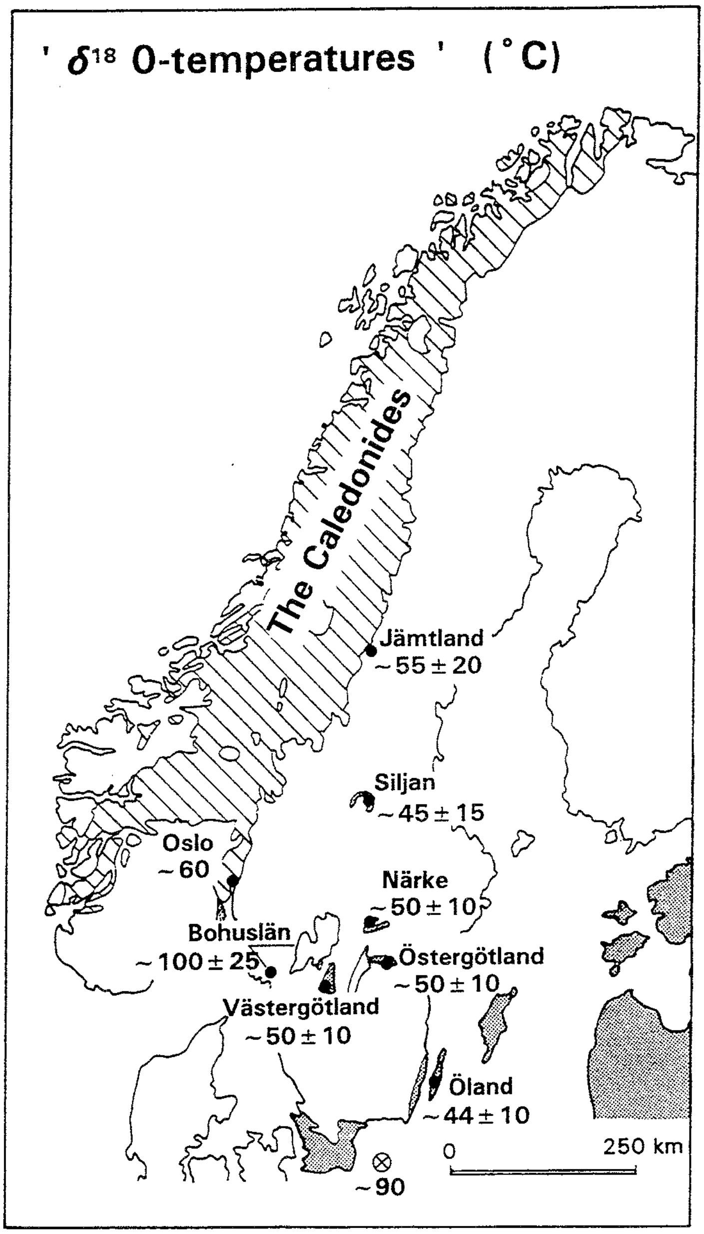

Figure 5. *Calculated temperatures based on $\delta^{18}O$–values of analysed limestone samples assuming equilibrium with a marine porewater of 0 to 2 o/oo (SMOW), (fractionation factor according to O'Neil et al., (1969)).*

5 PALAEOZOIC-MESOZOIC PALAEOTEMPERATURES - LITERATURE DOCUMENTATION

5.1 CONODONT ALTERATION INDICES (CAI)

Temperature induced colour changes due to low grade metamorphism of conodonts were studied by Bergström (1980) (Figure 6a). He determined Conodont Alteration Indices (CAI) according to Epstein et al. (1977) on samples of Ordovician limestones from 110 localities in Scandinavia and the British Isles. The results from Sweden show that sediments from the Gotska Sandön island, Lake Tvären in SE Sweden, the Östergötland province, the Närke province, Bothnian Sea (close to the Swedish east coast), and the Siljan impact structure all show CAI–values of 1 - 1.5, which means that they have not been heated above 90°C. The lack of strong thermal imprint on the samples from the Siljan impact structure is in accordance with results obtained at similar structures in the U.S. (Bergström 1980).

Samples from the Västergötland province show a wide range of CAI–values; 1.5 to 5 (<90° to >400 °C) for Billingen/Kinnekulle and 6–7 (>400°C) for Halleberg. This is interpreted as due to the heating from the intruded dolerite sill. At Billingen this influence can be recorded only down to a few tens of metres beneath the dolerite (Middle Ordovician). Whereas the Lower Ordovician limestones show no significant increase in CAI–values, varying between 1 to 2 (<140°C) (Bergström 1980).

Samples of the Ordovician limestones from the Caledonian front in Jämtland are also included in the study of Bergström (1980). These samples show large variations of CAI–values (3 - 5), indicating maximum temperatures of 110° to >300°C. It was concluded that these rocks have been burried to a depth of 3–7 km either by autochthonous overburden or by a pile of nappes from the Caledonides. He also suggests a third possibility of regional heating ("hot spot") during the Caledonian Orogeny.

CAI–values of 4.5 to 5 were recorded from the Oslo region, suggesting a temperature of 300°C. Permian volcanic activity was extensive in this region explaining the high CAI–values according to Bergström (1980).

5.2 ILLITE/SMECTITE RATIOS AND ILLITE CRYSTALLINITY

Maturity and mineralogy of the black shales from the Caledonian front in Jämtland has been investigated by Snäll (1988). Based on illite crystallinity, XRD–peak ratios and polymorph criteria, he concluded that the shales have suffered burial metamorphism not exceeding 250–280°C. Based on the rank of the organic matter in the shales (e.g. H/C ratios and vitrinity reflectance) a temperature above 90°C and probably around 200° was suggested (Snäll 1988) (Figure 6b). This is in accordance with the results from an investigation of illite crystallinity and vitrinity reflectance carried out by Kisch (1980) on samples from the same area. Both authors argued for a cover of Caledonian Nappes as the reason for metamorphism. The possibility of an autochthonous cover of more than 4 km of Devonian sediments was also discussed. The "hot spot" theory suggested by Bergström (1980) was, by these authors, regarded as less probable.

Analyses of illite/smectite (I/S) ratios (determined by XRD) in samples from 100 to 450 m depth in a drillcore at Grötlingebo, Gotland (Figure 6b) gave ratios of 0.5 to 0.8, i.e. illite constitutes 50 to 80% of the analysed samples. Snäll (in Push & Karnland 1988) interpreted this as due to heating to at least 110°C corresponding to a depth of 2.5 km or temperatures above 170°C if the depth was only 1 km. Snäll (1988) also investigated crystallisation index (CI) and sharpness ratio of the 10 Å peak of the illite samples using XRD. These results together with traces of oil in the sediments indicate that these samples are within the "oil window" and that they have suffered prolonged heating of 110 to 120°C, i.e. suggesting an overburden of at least 2.5 to 3 km of probably Devonian sediments.

K–Ar datings on autogenic, illite dominated, mixed–layer clay samples from fracture zones in the Äspö tunnel, SE Sweden gave ages around 300 Ma (Maddock et al. 1993). Illite/smectite ratios determined by XRD (by S. Snäll, Geological Survey of Sweden) of clay minerals from the same set of NE trending fracture zones indicate a formation temperature in excess of 100°C and a pressure corresponding to at least one kilometre of overburden (cf. Velde 1985).

A discrepancy between illite/smectite ratios and organic maturation, where the XRD determined I/S ratios indicate higher temperatures (higher I/S ratios) than the maturation data, have been reported by e.g. Hurst and Irwin (1982).

Chemical and physical properties of Middle Ordovician potassium bentonites from Kinnekulle, S Sweden were presented by Brusewitz (1986). She determined the illite/smectite ratios by using K_{fix} estimated from chemical analyses and structural formula (K fixed per $O_{10}(OH)$) and compared these values with the illite/smectite ratios estimated by XRD. The XRD determined illite/smectite ratios were higher than those estimated by the K_{fix} method. The different ratios were compared with the CEC (Cation exchange capacity)–values for the same samples and the value for the K_{fix} method turned out to be the most reliable (Brusewitz 1986).

It is thus reasonable to consider the temperatures estimated from the I/S ratio as maximum temperatures.

5.3 ORGANIC MATURITY

Several studies of organic maturity have been carried out on the Cambrian–Ordovician remnants in Sweden (Figure 6c).

Buchardt et al. (1986) reported a study of thermal maturity of Alum Shales in southern Scandinavia. The maturity was rated by HI–OI index, T_{max}, PI and "vitrinite" reflectance values (R_R %) on liptinite organic matter. (True vitrinite reflectance is measured on altered remains of vascular land plants which are not present in Cambrian and Ordovician rocks). A ranking of the thermal maturity of the shales was carried out. Samples from Närke were ranked as immature or early mature with respect to oil generation. Samples from Öland showed a southward change going from immature in the north to early mature in the south as a result of increasing burial depth to the south. Samples from central Västergötland (Billingen) were within the oil window but showed large local variations of R_R% values from 0.9 to 1.3. The higher values were attributed to sampling near a Permian dolerite sill. Analyses of samples from Oslo showed that they are post-mature with respect to oil generation.

Maturity ranking by means of e.g. reflectance of vitrinite-like macerals, from black shales in Sweden has also been reported by Lewan and Buchardt (1989). In their study they recorded R_R%– values for samples from Öland, Östergötland, Närke and Kinnekulle (Västergötland) between 0.44 to 0.53 (except for one sample of 0.32 from Öland). Values less than 0.5 R_R% are indicative of thermal immaturity (Wright 1980). The temperatures corresponding to a maturation ranking within the oil window are c. 65° to 120°C according to e.g. Heroux et al. (1979). However, the authors state that these temperatures are very preliminary. Wright (1980) stated that the temperature for oil maturity varies from 60° to 150°C in different basins but that a reasonable average for the window can be 80° to 120°C.

In conclusion most of the analyses of Buchardt et al. (1986) and Lewan and Buchardt (1989) show that the samples from Östergötland, Kinnekulle (Västergötland), Närke and Öland are close to the threshold for oil maturity (immature– early mature). Samples from the Oslo region and Billingen (Västergötland) were influenced by Permian volcanism why they show higher thermal maturity.

Kisch (1980) carried out vitrinite reflectance measurements on the black shales from the Caledonian front of Jämtland. He found these to be postmature with respect to oil generation, i.e. temperatures had exceeded 120°C. This is in accordance with the results reported by Snäll (1988) on samples from the same area.

Lower palaeozoic shales and limestones of the Siljan impact structure were analysed with respect to organic richness, type and maturity (Vlierboom et al. 1986). They concluded that all the analysed organic–rich rocks are on the verge of oil generation. However, oil seeps in the area are more mature (within the oil window). The reason for this can not be explained by known thicknesses (Thorslund 1960) of Palaeozoic sediments in the area.

5.4 ORGANIC PRECIPITATES

Findings of asphaltite in fracture zones of the Precambrian basement rocks in Sweden are relatively frequent. A number of samples ranging from different mines from Boliden, N Sweden to Kvarntorp, S Sweden were presented by Welin (1966). From C/H, $\delta^{13}C$ and ^{14}C analyses he concluded the carbonaceous material to be of plant or animal origin. U/Pb analyses of associated thucholite samples gave no precise ages although a Phanerozoic age for the formation was suggested.

Radionuclide mobility in thucholite samples from fractures in Pre–Svecokarelian quartzite of the Västervik region, SE Sweden has been studied by Åberg et al. (1985a). They reported uranium–lead apparent ages for the crystallisation of the thucholite to be inconsistent, indicating radionuclide remobilisation. However, ages of 300–400 Ma from thucholite in drillcore samples taken from depths exceeding 20 m were suggested to better reflect ages of mineralisation.

5.5 FISSION TRACK AGES ON APATITES FROM BASEMENT GRANITES

Several fission track datings on apatite are reported from Sweden (Figure 6d). Annealing temperatures of apatite are suggested to be approximately 125°C +/–25° (e.g Naeser, 1979; 1981).

The sub–Cambrian peneplane is identified over large areas of Sweden (Rudberg 1970; Lidmar–Bergström 1988). This means that apatites at the present surface (except for the area influenced by the Caledonian orogen) should record Vendian or older fission track ages, if not influenced by Phanerozoic heating. However, all ages recorded are generally Late Carboniferous to Late Triassic (299 to 220 Ma) except for two. One of these (486 Ma) was excluded by Koark et al. (1978) as less reliable. The other age (375 Ma) was reported from the Siljan impact structure (Andriessen & Verschure 1991).

Resetting in the province of Dalsland was interpreted by Zeck et al. (1988) as caused by overburden of 3 – 4 km of Silurian and possibly Devonian sediments. The track length distributions in apatite indicated slow cooling from 125°C to ambient surface temperatures. This corresponds to a "slow uplift" by erosion without significant tectonic disturbances. Titanite fission tracks indicate ages of 680 +/- 50 Ma (Zeck et al., 1988). As the annealing temperature of titanite is 250 +/-50°, these temperatures were not reached during the Phanerozoic.

Persson (1986) and Hansen et al. (1989) reports apatite fission track ages from 228+/-38 Ma to 299+/-50 Ma for granitoids from the western part of the Värmland province. Corresponding lower concordia intercepts for U-Pb zircon datings cluster around 400 Ma.

Andriessen & Verschure (1991) report fission track ages of 375 Ma on apatites from the centre of the Siljan impact structure, which is the approximate age otherwise recorded for the impact (Bottomley et al. 1978; Åberg et al. 1989). Notably fission track datings from the Dala granite outside the impact structure yielded apatite ages of 250 Ma, which is in accordance with fission track apatite ages obtained from other investigated basement granites in Sweden.

Koark et al. (1978) recorded fission track apatite ages varying from 191+/-48 Ma to 281+/-58 Ma from the northernmost parts of Sweden and 283+/-73 Ma from Grängesberg, central Sweden. These results were interpreted to indicate an overburden of cover rocks.

Based on fission track studies Andriessen & Bos (1986) calculated a mean uplift rate of 0.1 mm/year for SW Norway during the first 80 Ma following the Caledonian orogeny. This was slowed down to 0.02 mm/year from then to the present. They concluded that during the first 80 Ma c. 5000 metres of crustal material was eroded and that altogether 13000 metres have been eroded since the Caledonian orogeny (Andriesen & Bos 1986). We suggest that these sediments were in part deposited on the foreland to fill a foredeep basin.

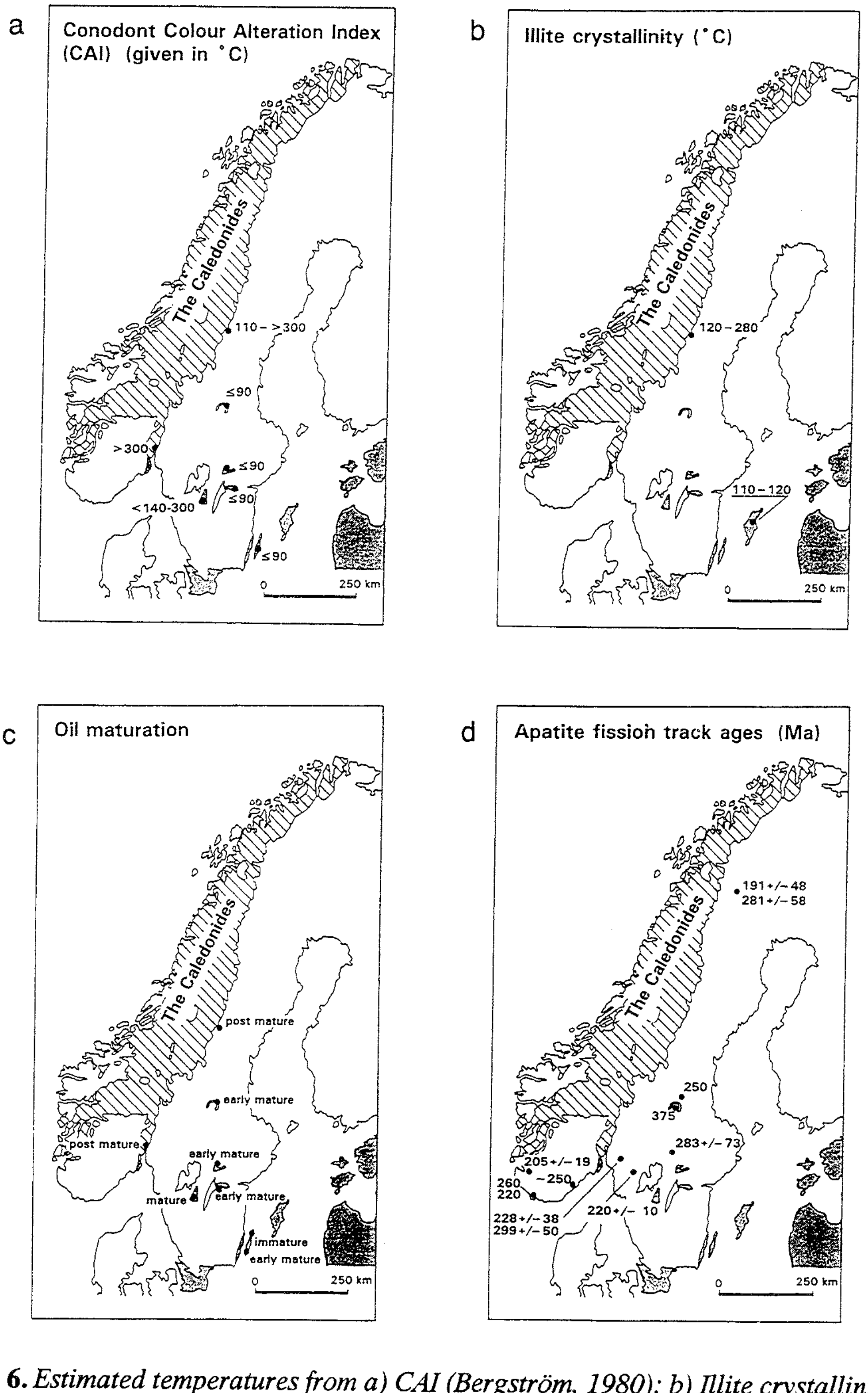

Figure 6. *Estimated temperatures from a) CAI (Bergström, 1980); b) Illite crystallinity (Kisch, 1980; Snäll, 1988; Push & Karnland, 1988); c) Oil maturation (Kisch, 1980; Buchardt et al., 1986; Vlierboom et al., 1986; Snäll, 1988; Lewan & Buchardt, 1989); d) Apatite fission track (van den Haute, 1977; Koark et al., 1978; Zeck et al., 1988; Hansen et al., 1989; Andriessen, 1990; Andriessen & Verschure, 1991).*

5.6 PHANEROZOIC MOBILISATION OF LEAD

Ore lead crystallized as galena is well known from many areas in Sweden (e.g. Johansson & Rickard 1984). Some of these occurances are considered to have crystallized during the Palaeozic. The most prominent ore body from this period is the Laisvall ore consisting of disseminated galena in a Vendian sandstone located along the eastern border of the Caledonides (e.g. Grip & Frietsch 1973; Johansson & Rickard 1984).

Other important galena ores are known from Dorotea and Idre/Vassbo, all along the Swedish Caledonian front, within quartz sandstone (or quartzite) of Vendian/Cambrian age. The mineralizations are all influenced by tectonic structures. Grip & Frietsch (1973) suggest that remobilization of lead and precipitation of galena in connection to shear and fracture zones took place during the final stage of the Caledonian orogeny. Roedder (1968) showed that the ore–forming solutions at Laisvall have a high salt content and were formed at temperatures of more than 100°C.

Fracture fillings of galena in dykes are also known from within the Scandinavian Caledonides (Grip 1967). They seem to be connected to basement cored windows with associated Vendian sandstones.

East of the Caledonian front occurences of galena are known within N–S to 20°E trending shear and fracture zones (Romer 1990, 1991 a,b). They represent listric faults, similar to those recorded in the Rombak window by Bax (1989), which were mobile during or subsequent to the Caledonian orogeny. The galena is crystallized in quartz and/or calcite veins and is characterized by an anomalously high content of radiogenic lead. According to Romer (1991a) the lead was leached from Proterozoic rocks c. 400 Ma ago. This type of veins can be traced at least 100 km to the east of the present Caledonian front.

Other lead bearing fractures of Phanerozoic age are also known well outside of the Caledonides. They are frequent in the Lower Cambrian sandstones in the province of Scania, south Sweden. Some of these are fillings in fault breccias.

A regional survey of the isotopic compositions of ore lead from 23 vein and impregnation deposits in the Swedish segment of the Fennoscandian Shield was carried out by Johansson & Rickard (1984). They interpreted the galena crystallization as due to leaching from Precambrian basement rocks. Variation in isotopic ratios because of selective leaching of the rocks and mixing with less radiogenic Caledonian lead (for galena ores along or close to the Caledonian front) was recorded. A model where 1.8+/–0.15 Ga old basement lead was leached and crystallized at 0.4+/–0.15 Ga was advocated. The galena in limestones within the Siljan impact structure was considered to be due to mobilization of radiogenic lead during burial of post–Silurian sediments (Johansson & Rickard 1984).

Other examples of galena bearing fractures are the Cambrian sandstone fracture infillings (Mattsson 1962; Nordenskiöld 1944; Samuelsson 1975; Bergman & Lindberg 1979) found in different parts of the shield. The Meso–Proterozoic Götemar granite on the Swedish east coast and the Rapakivi granite in Åland are examples of rocks that host such dykes. In both cases the galena shows excess of radiogenic lead (Sundblad 1991; Vaasjoki 1981) resulting from remobilization.

Fission tracks in the Götemar granite reveal that uranium is mainly contained in zircon and monazite (Kresten & Chyssler 1976). These minerals are the main carriers of U in most granitoids implying zircon as a major source for leaching of radiogenic lead. Smellie & Stuckless (1985) showed that post–magmatic, hydrothermal alterations may have taken place within the time range 420+/–171 M.a. This corresponds to the lower discordia intercept age. A similar lower discordia intercept for the Götemar granite is also reported by Åberg et al. (1985b). Smellie & Stuckless (1985) also concluded that the major disturbance of the U–Th–Pb system was almost certainly dominated by lead mobility.

The galena fillings at Åland is considered to be younger than 500 Ma, as they cut across already lithified Late Cambrian sandstone fracture infillings. U–Pb zircon datings of Rapakivi granite (Suominen 1991) show Mesozoic lower discordia intercepts with large errors.

Lower discordia intercepts for Swedish U–Pb zircon datings cluster with a maximum at 300 to 400 Ma. This is shown in Figure 7 where only datings with errors less than 100 Ma are used. SW Swedish zircon data are omitted, due to the polymetamorphic character of this terrane and strong indications that Sveco–norwegian thermal and tectonic events have influenced the U–Pb system.

Similar lower intercept ages are reported from S Norway but somewhat lower ages are recorded from SW Finland (Suominen 1991). We speculate that mobilization of lead and precipitation of galena during the Phanerozoic was due to a prolonged thermal event which is indicated by discordia lower intercepts. This is dated to be 200 to 400 Ma and could represent heating due to burial by Palaeozoic sediments and along the Caledonian margin also by allochthonous rocks.

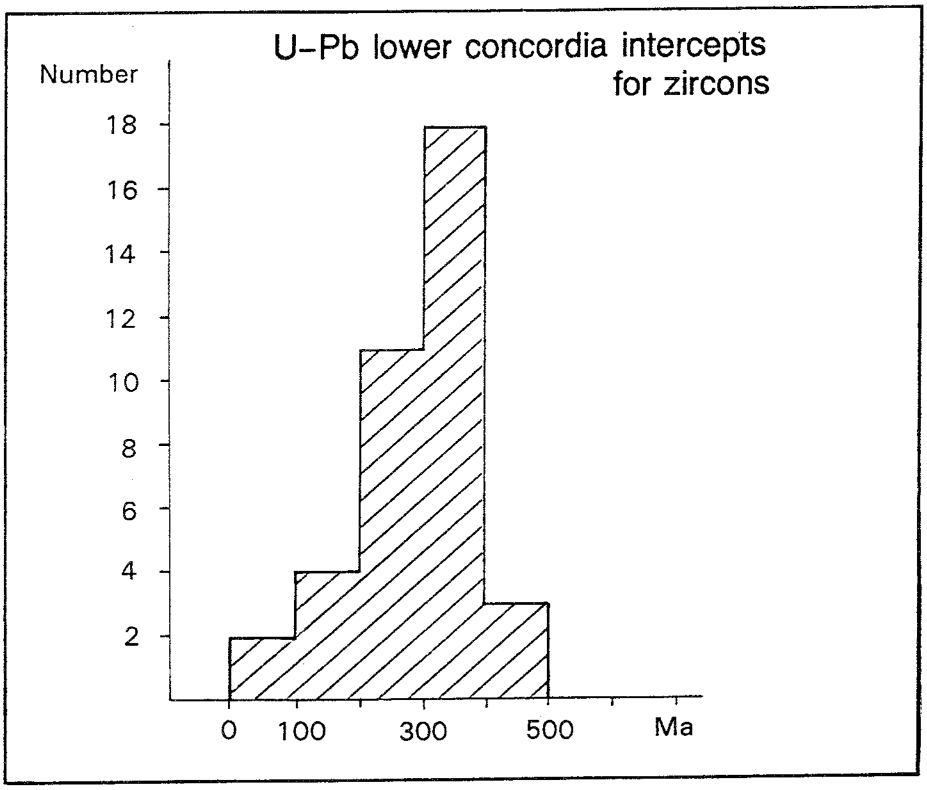

Figure 8. *Histogram for zircon lower intercept ages from Swedish Svecokarelian/ Svecofennian rocks. (Reference list on request from the authors).*

6 MODELLING THE PHANEROZOIC COVER

Thermal events like intrusions of Permian dykes, extraterrestrial impacts and overthrusting nappes can only explain the observed temperature changes in a few places. If we take all the results here reviewed into account, there is an overwelming support for a very low grade metamorphism of the Palaeozoic sediments in Sweden. In those areas where indicators have been refered to other than from the Lower Palaeozoic, we know that the samples originate from a crustal level close to the sub-Cambrian peneplane, which is possible to trace over large parts of Sweden (Lidmar-Bergström 1994). Although the basement has been eroded where younger denudation surfaces than this have developed, the erosion must have been relatively small. We conclude that burial under a sedimentary and/or tectonic cover is the most plausible explanation for the observed temperature records.

Estimated temperatures by different indicators are shown in fig. 8. There is a difference in reliability for different temperature indicators. Generally the temperatures indicated by $\delta^{18}O$ are the lowest and those indicated by CAI and illite/smectite ratios are the highest. Organic maturity (e.g. vitrinite reflectance) usually support a temperature significantly higher than those indicated by the $\delta^{18}O$–values and less than, or equal to those from the clay mineralogy and the CAI values. These relations are in accordance with earlier observations (cf. Hurst & Irwing 1981; Marshall 1992). Temperatures of approximately 90 to 100 °C during a period of at least 100 Ma are recorded for large parts of Sweden and even higher temperatures are indicated close to the Caledonian front.

In order to estimate the thickness of a cover corresponding to the temperature indications, several assumptions must be made for the properties of the sedimentary cover that has been eroded. Heat production and thermal conductivity of the sedimentary pile as well as heat flow from the basement determines the temperature gradient of the cover. The heat is here assumed to be transported by thermal conduction.

A comparision of the apatite reset age obtained inside the Siljan impact structure (375 Ma) with the apatite age obtained from the Dala granite outside the structure (250 Ma) suggests that Devonian sediments dominated the pile. That is, at least the major part of the sediments was deposited before the Late Devonian rebound in the crater area. We infer that the sedimentary pile was dominated by sandstones as the conditions during Late Silurian to Early Devonian in S Sweden changed from shallow marine to continental (Lindström et al. 1991). This is further supported by the occurence of more than 1 km thick deposits of Devonian sandstones in Estonia and Latvia (Grigelis 1991; Kurss 1992), and the Röde Sandstone in Jämtland, Sweden, all deposited in a fluvial regime .

A thermal conductivity of approximately 2 W/m°C is proposed (based on measurements of Early Palaeozoic and Mesozoic sediments by Sundberg et al. (1985)).

The maximum heat production, through decay of U, Th and K, in the sediments, was probably similar to that in typical granites or somewhat lower (max. 2.5 $\mu W/m^3$ for a granite with 4 ppm U, 15 ppm Th and 3.55 % K (Fowler 1990). Basal heat flow between 45 to 75 mW/m^2 can be estimated based on studies presented by Landström et al. (1980). A detailed evaluation of the heat flow based on measurements from the deep well Gravberg–1 (Siljan) yielded values in the order 60–65 mW/m^2 (Balling et al. 1992). This is probably somewhat high concerning Swedish conditions (Landström et al. 1980). Although the Devonian heat flow probably was slightly higher than at present we have used 55mW/m^2 as the basal heat flow in our calculations.

The position of Fennoscandia during the Devonian to Permian period is assumed to have been 0 to 30°N (cf. Pesonen et al. 1991; Thorsvik et al. 1990) and the average surface temperature was probably in the range of 15 to 20°C. These values were used in the formula below to calculate the temperature beneath a 3 km thick sedimentary cover:

$T= -Az^2/2k + [(Q+A)/k]z$ where
T = temperature in °C at the depth z
A = radioactive heat generation
d = depth of the rock column
Q = heat flow from below into the column
k = the thermal conductivity of the column

Using the input data given above temperatures of 103 to 108°C and a temperature gradient of c. 30°C/km were obtained. The heat generation of the sedimentary pile implies only marginal effects on temperature at the bedrock/sediment interface (c. 5°C lower temperature when decreasing the value to 0.5 instead of 2.5 $\mu W/m^3$). It is also noticed that the relatively thin, less than 10 metres, U–rich part of the Cambrian black shale (e.g. Andersson et al. 1982) does not affect the temperature gradient significantly. In contrast, the conductivity of the sediment has a major effect on the temperature; higher conductivity means lower temperatures at depth (a conductivity of 2.5 instead of 2.0 W/m°C will decrease the temperature at 3 km depth with 16°C). Also, the heat flow from the crystalline basement has a great influence on the temperature. A change of 10 mW/m^2 in heat flow will change the temperature with 15°C. Since basal heat flow is known to vary (e.g. Landström et al. 1980), local variations in temperatures can be expected. Calculated temperatures are consistent with a sedimentary cover of approximately 3 +/– 1 km for large parts of the shield, which is considerably thicker (allochthonous or autochthonous) along the Caledonian border. Snäll (1988) suggested thicknesses of overburden in the Jämtland region in the order of 4–6 km. This was suggested to be due to either a sedimentary cover (autochthonous) or nappes. Supporting evidence for the nappe tectonics to the east of the present Caledonian front is presented by Hossack & Cooper (1986) and Palm et al. (1991).

Based on apatite annealing ages the sedimentary cover remained several km thick until the Permo–Triassic period. Morphological studies (Lidmar–Bergström 1991) show that in places where the sub–Cambrian peneplane is preserved, e.g. in south–central Sweden (Västergötland and Östergötland) the bedrock surface generally was covered by at least a thin sedimentary layer until the Early Tertiary. However, on the west coast of Sweden a Permo–Triassic denudation surface is distinguished by Lidmar–Bergström (1992) and furthermore Devonian as well as Carboniferous terrestrial fossils are recycled into Jurrasic sediments in N Scania (Dorothy Guy–Olsson, Stockholm, pers. com.). These observations are in accordance with the resetting ages of the apatites (Figure 6d) and indicate extensive uplift/erosion during the Late Palaeozoic and Early Mesozoic. A tentative model of the sedimentary thicknesses during the Phanerozoic is suggested in Figure 8.

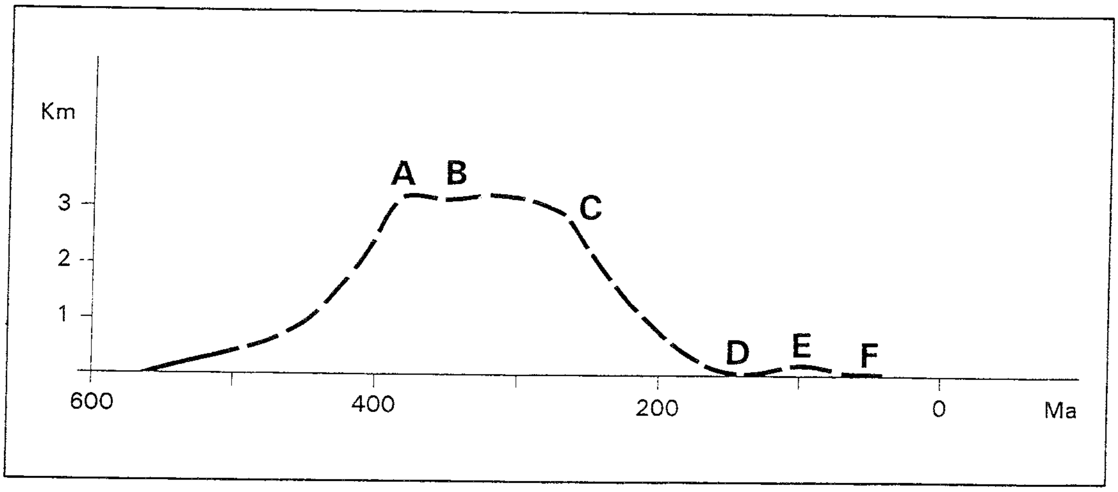

Figure 8. *Estimated thickness of the Phanerozoic sedimentary cover in south–central Fennoscandia. Full line represents the thickness of the Lower Palaeozoic sedimentary cover as reconstructed from existing outcrops. Broken line represents thickness estimated from references and analyses in this paper.*
A: The Devonian cover rocks were already of considerable thickness (ca. 3 km) when the Siljan impact structure was formed. B: This cover can not have been much thicker during the succeeding period as recorded by thermal indicators (see text). Local erosion/sedimentation has probably influenced the cover thickness during this period C: The basement was cooled to a temperature allowing apatite fission tracks to be formed (thickness < 3km). D: The Mesozoic denudation surface was formed in parts of S. Fennoscandia. E: Cretaceous sedimentation in parts of S. Fennoscandia. F: Tertiary (and Quartenary) glaciations unroofed the basement.

7 CONCLUSION

Thermal indicators including $\delta^{18}O/\delta^{13}C$, conodont alteration indices (CAI), illite/smectite ratios and illite crystallinity, oil maturation in Lower Palaeozoic sediments, as well as apatite fission tracks and lead mobility in basement rocks all indicate raised temperatures during the Late Palaeozoic, as high as 125°C at the present level of erosion. These temperatures persisted for at least 100 Ma in south central Sweden. Calculations indicate a sedimentary cover thickness of 3+/−1 km, a thickness which probably varied geographically. We propose that this sedimentary cover was largely composed of molasse, which was eroded from the Caledonides, and deposited into a Caledonian foreland basin. In southern Sweden this basin probably would have interfered with a basin developed to the north of the Danish–German arm of the Caledonides. Regional variations in sedimentary thicknesses may be explained by flexing of the lithosphere and the temporary existence of a foreland bulge (Figure 9). A dynamic interaction between the load/erosion of the Caledonian nappe stack and coupled changing geometry of the foreland basin and bulge is compatible with the stratigraphic, petrographic and fluvial directions of Devonian sediments (Kurss 1992) remaining in the Baltic Sea and States. In southern Scandinavia the Permian to Triassic uplift and erosion reduced the cover significantly.

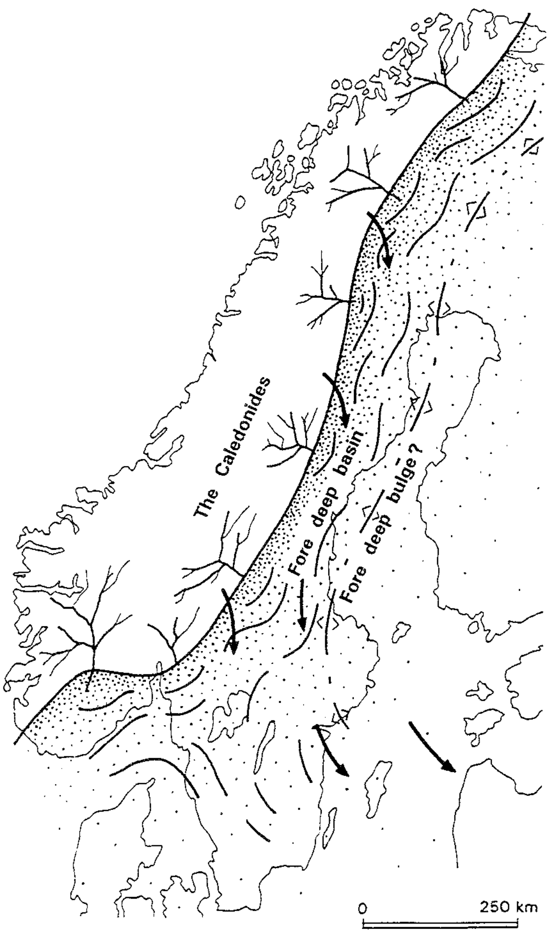

Figure 9. *Tentative sketch of a Caledonian foredeep basin. The sedimentation was not contemporaneous over the entire basin. During the Late Silurian terrestrial facies prograded towards SE (Basset 1985), a direction which was confirmed also during the Devonian in the Baltic Sea area (Kurss 1992).*

ACKNOWLEDGEMENTS

Lars Karis at SGU, Uppsala provided several limestone samples and generous information about the Swedish Lower Palaeozoic. Charles Naeser, USGS, Denver, Mike Middleton and Rodney Stevens, Göteborg, scrutinized early versions of the manuscript and gave us valuable critizism. John Dewey, Oxford University, reviewed the manuscript and provided good advices. Gustaf Lind, Göteborg and Ove Landström, Trosa contributed with their knowledge about thermal flow. Karl-Axel Kornfält, Hugo Wikman, SGU, Lund and Matti Vaasjokki, Geological Survey of Finland, Espoo, generously gave us access to unpublished, radiometric dating results. Piret Plink, Göteborg, helped us with translations of Russian publications. Kristina Kling and Kerstin Berndtsson, Göteborg, made the drawings. The Swedish Natural Science Council and the Swedish Nuclear Fuel and Waste Management Co provided financial support.

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