Interpretation of the lead isotopic composition from sulfide mineralizations in the Proterozoic Sjangeli area, northern Sweden

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Cu-, Fe-, and Zn-mineralizations occur in the Lower Proterozoic supracrustal rocks of the Sjangeli area, a tectonic basement window in the Caledonides of northern Sweden and Norway. The supracrustal belt consists of basic and ultrabasic metavolcanites which are intercalated with quartz-mica schists and banded carbonate-silicate rocks. The metavolcanites consist of tuffs, lava flows, and pillow lavas. Cu-mineralizations, which form three different types, occur within basic metatuffs. Two types represent epigenetic remobilizations of Cu-mineralizations, while the third type is characterized by stratiform bornite-chalcocite-quartz-magnetite layers of possible syngenetic origin. The Fe- and Zn-mineralizations occur in quartz-mica schists and form stratiform lenses.

At about 430 My a disturbance of the lead isotope system occurred, which changed the $^{238}U/^{206}Pb$, $^{235}U/^{207}Pb$, and $^{232}Th/^{208}Pb$ ratios. Radiogenic lead was added to the mineralizations during the 430 My disturbance. The added lead component was different for the various deposits. Further, during the 430 My event, U was mobile, which is illustrated by the significantly increased $^{238}U/^{206}Pb$ ratios of the massive pyrrhotite mineralizations.

Lead lines from the recalculated lead composition (430 My) indicate mixing model ages of 1546 My to 2112 My. Since the mixed lead components were non-cogenetic, the model age variation of the mixing lines indicates differences of the mixing component, rather than real age differences. However, a lead source of Middle Proterozoic age can be deduced.

The Middle Proterozoic disturbance resulted in the formation of vein mineralizations which have predominantly high-uranogenic, low-thorogenic lead. The in-situ lead growth indicates high $^{232}Th/^{207}Pb$ ratios and implies that U had also been mobile during the Proterozoic metamorphism.

Model ages for the least radiogenic lead (Koppasen) yield a minimum age of 2050 My.

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Introduction

In the southwestward younging Baltic Shield of northern Scandinavia (Sköld 1986, 1987, Gaal & Gorbatschev 1987), the Rombak-Sjangeli basement window forms an accretory zone of the Archean craton of probable island arc setting (Korneliussen et al. 1986). The Rombak-Sjangeli window is characterized by N-S striking volcano-sedimentary belts. The volcanites vary from basic and ultrabasic composition at the eastern border to intermediate and acidic composition in the west (e.g. Korneliussen et al. 1986). The Sjangeli area represents the easternmost of these Lower Proterozoic supracrustal belts and consists of basic volcanites (tuffs and lavas) and carbonate and pelitic sediments.

These Proterozoic supracrustal belts of the Rombak-Sjangeli window became intruded by acid to basic rocks at about 1700–1800 My ago. This is illustrated by Rb-Sr ages deduced from whole-rock isochrons from granites, e.g. $1691 \pm 90$ My (Heier & Compston 1969) and $1780 \pm 85$ My (Gunner 1981). Subsequently, the Proterozoic rocks were eroded and were covered by discontinuous metaconglomerates and arkosic quartzites (cf. Kulling 1964, Tull et al. 1985), presumably during the Cambrian (Bergström & Gee 1985). The quartzites generally form thin flat-lying veneers, but locally they are bent upright along steep faults which remobilized basement blocks (Bax 1986, 1989). The basement reactivation is related to the Caledonian overthrusting, which in the Rombak-Sjangeli area probably started about 450 My ago (e.g. Tull et al. 1985). The overriding of the Caledonian nappes over the basement resulted in a low-grade metamorphism reaching greenschist facies (e.g. Bryhni & Andre-
asson 1985, Tull et al. 1985) in the eastern part of the basement windows, which includes the Sjangeli area.

Sulfide mineralizations occur within the supercrustal belts of the Proterozoic basement. The lead isotopic composition of sulfide minerals from some of these mineralizations (cf. Figs. 1 & 2) has been determined. The bulk of these analyses comes from the Cu-mineralizations of the Sjangeli area and their wall rocks.

The purpose of this paper is (1) to use the lead isotopic composition of the samples to illustrate the effect of the Caledonian and Svecokarelian orogenies on the isotopic composition of the sulfide lead and the whole-rock trace lead; and (2) to use the lead composition as a tracer of the source of the different lead components.

Geology of the Sjangeli area

The Sjangeli supracrustal belt lies at the eastern border of the Rombak-Sjangeli window (Fig. 1) and is correlated with the Kopparåsen and the Gautelis supracrustal rocks (cf. Adamek 1975, Romer 1987). The Sjangeli supracrustal belt, which is about 10 km long and reaches a width of 5 km, strikes approximately N-S and the units dip steeply to the west (70-85°). The belt is characterized by a strike-parallel 'zebra-stripe' pattern consisting of volcanic units which are intercalated with sedimentary and volcano-sedimentary units.

To the east (Fig. 2), the metamorphosed volcanic rocks are lava flows, that are embedded in basic metatuffs, and which commonly are strongly mylonitized. The lava flows occur only in this easternmost volcanic unit, and they do not contain any mineralizations. The waterlain basic metatuffs, hereafter sometimes simply called tuffs, contain thin layers of banded silicate-carbonate rock. The tuffs are the most intensely mineralized part of the Sjangeli area. Most of the stratabound mineralizations are stratiform, but locally discordant quartz-bornite veins also occur, which could represent remobilizations of stratabound mineralizations.
The central volcanic unit consists of high-Mg basalts (12-18% MgO) which contain lenses of serpentinites. The amphibolites consist mainly of layered metatuffs, but locally contain pillow lavas and thin layers of banded carbonate-silicate rock. The high-Mg basalts have high contents of Cr and Ni (Romer 1987) and show platinum group element ratios similar to komatiites (Barnes et al. 1988). The amphibolites do not contain Cu-mineralizations. A metasomatic alteration zone which contains a magnetite layer with bornite occurs at the NE contact between the mica schists and a serpentinite lens.

The western volcanic unit consists of pillow lavas which locally are interlayered with thin discontinuous layers of banded carbonate-silicate rock. This unit does not contain Cu-mineralizations.

Quartz-mica schists and banded silicate-carbonate rocks intercalate with the volcanic and volcano-sedimentary units. At Ruvssot (Fig. 2) the quartz-mica schists contain stratiform Cu-mineralizations in the vicinity of the metasomatized alteration zone of the amphibolites. The quartz-mica schists contain additional stratiform Fe-mineralizations and at Cunojavre (cf. Fig. 2) they contain a stratiform Zn-mineralization.

The amphibolites show Early Proterozoic amphibolite facies mineral assemblages. The other volcanic units have been extensively retrograded and show greenschist facies mineral assemblages.

Two metamorphic events and possibly the spilitization of the volcanic rocks have disturbed the isotope systems. The youngest event, which partially reset the isotope systems, is the Caledonian orogeny. Here, the term Caledonian is used rather loosely and refers to any event in the approximate time range 380 to 480 My (Romer 1989). Rb-Sr whole-rock data yield mixing lines with an apparent age of 488 ± 64 My (Romer 1989). Similarly, mineral Rb-Sr isotope data indicate a partial remobilization of the isotope systems at that time. Since the mixing lines do not yield ages, no precise time constraints for the Caledonian overthrusting can be deduced for these lines. Tull et al. (1985) refer to a dynamometamorphic event prior to c. 450 My ago to the west of the Rombak-Sjangeli area, while mineral ages from Vestvågøy vary from c. 460 My (amphiboles) to c. 350 My for biotites (Tull et al. 1985). For the recalculation of the lead isotope composition, an age of 430 My was arbitrarily chosen.

Whole-rock lead lines indicates ages of 1600 - 1800 My (Romer 1989). These ages coincide with the ages of the Svecokarelian granites, to which the granites and syenites of the Rombak-Sjangeli window belong (cf. Gorbatschev 1985). Similar metamorphic ages were obtained for the Vesterålen area (e.g. Griffin et al. 1978, Jacobsen & Wasserburg 1978).

Rb-Sr whole-rock data yield a date of 2324 ± 15 My (Romer 1989). Since this date is based on only 3 analyses and as the area has been metamorphosed twice, it probably does not represent a reliable age and is probably too high by 100 to 150 My (cf. Romer 1989). Ages in excess of 2000 My have been reported from the Vesterålen area (Griffin et al. 1978). Where metavolcanites yield ages of about 2200 My.

Mineralization types

In the Sjangeli area, mineralizations of Cu, Fe, and Zn predominately form stratiform lenses and layers. However, Cu-mineralizations also form remobilizations in discordant veins and metasomatized contact zones. In the following discussions, the mineralizations are classified according to their main metal and, in addition, the Cu-mineralizations are classified into types (i), (ii), and (iii), while the Fe-mineralizations are separated into magnetite and pyrrhotite mineralizations.

Cu-mineralizations occur within the basic tuffs at Unna Alakats, Sjangeli, and further to the north at Valfjojokk (Fig. 2). Comparable Cu-mineralizations occur in similar wall-rock at Ruvsojaure and in the quartz-mica schists at Ruvssot. In addition, Cu-mineralizations at Ruvssot are related to a magnetite band in the metasomatic alteration zone between the ultrabasic volcanites and the mica schists (Fig. 2).

The Cu-mineralizations form three main types, that differ in respect to the field occurrence and the main copper sulfide (Romer 1987): Type (i) is the predominate mineralization in the Sjangeli area. It forms massive bands and lenses that contain more than 50% sulfides and oxides, with sulfides dominating over oxides. The bands are up to 5 cm broad and up to 50 m long while the lenses have a length of 20 to 50 cm and are 2 to 5 cm thick. The
bands and lenses contain mainly bornite, chalcocite and magnetite. Epidote and quartz are the main gangue minerals, while Fe-actinolite is minor.

Type (ii), which is only found at Unna Alakats, forms schistosity-parallel quartz veins which are discordant to the bedding. The quartz veins, about 5 cm broad, are paralleled by a sulfide band on one or both sides. The sulfide bands contain bornite and chalcocite together with minor magnetite. These bands were probably formed during the Proterozoic metamorphism and deduced their metal content from preexisting mineralizations of type (i).

Type (iii) occurs at Ruvsojaure and is different from the other two types since bornite and chalcocite are absent and chalcopyrite almost exclusively occurs in small veins. The mineralization is bound to an intensely epidotized and partly actinolitized amphibolite, which resembles the basic metatuffs from Unna Alakats.

In addition to these three main types (i-iii), a 1-3 m thick, 100 m long magnetite zone with abundant bornite occurs at Ruvssot. This zone is part of the metasomatic alteration zone between the serpentinites and the quartz-mica schists (Romer 1987).

Within the mica schists, banded quartz-magnetite mineralizations form stratiform bodies up to 1 m thick and 40 to 60 m long. Polycrystalline 'globules' of coarse pyrite and chalcopyrite, about 1 cm in diameter, occur locally within the magnetite bands. Near Cunojavre a stratiform Zn-mineralization, about 100 m long, forms a thin (5-10 cm) schistosity-concordant 'sheet'. Sphalerite is the main mineral of this massive mineralization, but minor pyrrhotite, chalcopyrite and pyrite also occur. Further, within the amphibolites, a 50 m long, 5-30 cm broad stratiform pyrrhotite mineralization occurs, consisting exclusively of pyrrhotite and quartz.

Methods

All isotope measurements were performed during a research visit at the University of California, Santa Barbara. Ion exchange techniques were used to separate Pb, U, and Th (Grünenfelder et al. 1986). Contents of Pb, U and Th were determined by isotope dilution using a composite $^{206}\text{Pb}/^{238}\text{U}/^{208}\text{Th}$ tracer. Pb was measured using a Re single filament silic-
Sulfides from the Sjangeli area

Cu-mineralizations

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Other mineralizations

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Sulfides from other supracrustal belts in the Rombak-Sjangeli area

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Table 1. Lead composition and contents (in ppm) of Pb, U, and Th of sulfides from the Sjangeli and related areas. For sample location compare Figs. 1 & 2. py = pyrite; po = pyrrhotite; bo = bornite; sph = sphalerite; ga = galena; ccp = chalcopyrite; ct = chalcocite; asp = arsenopyrite.

rocks of the Sjangeli area. The whole-rock lead isotope data for the tufts, the amphibolites, the lava flows, and the granites is presented in Romer (1987, 1989). Since most samples contained only trace lead (0.6-62 ppm Pb) and also contained traces of U and Th (up to 2.9 ppm U and 0.33 ppm Th; see Table 1), the lead compositions
have been altered by in-situ radiogenic lead generation. To obtain the lead composition at the time of the last disturbance, the lead has to be recalculated based on the assumption that the contents of U, Th and Pb remained unchanged after the last event. The time for recalculatio was arbitrarily set to 430 My (see also the section on the geology of the Sjange Li area). The recalculation eliminates the effect of in-situ radiogenic lead addition, and mixing relationships should become more obvious. Note that the recalculation of the isotopic composition of the lead is based on the assumption of no gain/loss of U, Th, and Pb.

The isotopic composition of the sulfide lead at 430 My

The sulfide lead isotopic compositions, shown in Fig. 3, form a rather well defined linear array in the $^{206}\text{Pb}/^{204}\text{Pb} - {^{207}\text{Pb}}/{^{204}\text{Pb}}$ plot. However, in the $^{206}\text{Pb}/^{204}\text{Pb} - {^{208}\text{Pb}}/{^{204}\text{Pb}}$ diagram, the samples scatter, and there are samples with high $^{206}\text{Pb}/^{204}\text{Pb}$ ratios and anomalously low $^{208}\text{Pb}/^{204}\text{Pb}$ ratios (e.g. H7, M9). The lead isotopic composition of the Fe-mineralizations and the Cu-mineralizations differs (cf. Fig. 6). There are also differences between the various Cu-mineralizations and, further, between the magnetite and the pyrrhotite Fe-mineralizations.

The Cu-mineralizations define a single lead line in the $^{206}\text{Pb}/^{204}\text{Pb} - {^{207}\text{Pb}}/{^{204}\text{Pb}}$ diagram, while in the $^{206}\text{Pb}/^{204}\text{Pb} - {^{208}\text{Pb}}/{^{204}\text{Pb}}$ diagram two different lead trends are obvious. One trend, defined by most Mx samples and Sja-1 shows a $^{208}\text{Pb}/^{204}\text{Pb}$ ratio of about 0.75, while the second trend defined by Hx and M9 falls clearly below this trend (cf. Fig. 3).

Samples H7 and H9 belong to the mineralizations of type (iii), while M9 belongs to mineralization type (i). Lead from these Cu-mineralizations indicates a lead growth in a high $^{238}\text{U}/^{206}\text{Pb}$, low $^{232}\text{Th}/^{206}\text{Pb}$ environment.

In contrast to these samples, samples from mineralization type (i) (except M9) show a highly variable lead composition, with low radiogenic lead for massive mineralizations and with more radiogenic lead in low-grade (usually disseminated) mineralizations. The radiogenity of the lead composition of this type correlates negatively with the lead contents (cf. Fig. 4) and therefore the lead line of the type (i) mineralizations could be the result of isotopic mix-
Interpretation of the lead isotopic composition from sulfide

Fig. 5. Lead isotope ratios from different units from the Sjangeli area (recalculated for 430 My). Data from Table 1 and Romer (1987). For discussion see text.

The lead composition from the Fe-mineralizations differs widely. Sulfide lead from the magnetite mineralizations follows the same trend as the Cu-mineralization type (i) lead. However, the lead from the magnetite mineralizations has a higher $^{206}\text{Pb}/^{204}\text{Pb}$ ratio for a given $^{208}\text{Pb}/^{204}\text{Pb}$ ratio (cf. Fig. 6). The pyrrhotite lead from the pyrrhotite mineralization is characterized by a large in-situ decay correction, since pyrrhotite has large contents of U. Note that the pyrrhotite has rather low Th contents. The lead line through the recalculated samples has a flatter slope in the $^{208}\text{Pb}/^{204}\text{Pb}$ - $^{206}\text{Pb}/^{204}\text{Pb}$ and the $^{207}\text{Pb}/^{204}\text{Pb}$ - $^{206}\text{Pb}/^{204}\text{Pb}$ diagram than the type (i) Cu-mineralizations and the magnetite mineralizations.

The recalculations of two isotopically different lead compositions. This would explain the constant $\kappa$ ratio ($^{232}\text{Th}^{238}\text{U}$) necessary to obtain a linear trend in the $^{206}\text{Pb}/^{204}\text{Pb}$ - $^{208}\text{Pb}/^{204}\text{Pb}$ diagram. Further, there would be no need for a negative correlation between the isotopic ratios and the lead contents, as an in-situ lead growth interpretation would imply (since lead content and lead radiogeny correlate negatively, cf. Fig. 4). The slope of the tie-line of samples M2 and M3 reflects a higher $^{206}\text{Pb}/^{204}\text{Pb}$ ratio than all other trends (Fig. 3). This indicates that if there had been mixing, the mixing lead composition cannot have been completely homogenous. The added lead composition, which could have been introduced at different times near 430 My, must have had local variations in its isotopic composition. Sample M11, which is the only sample from Cu-mineralization type (ii), falls on the trend defined by samples from Cu-mineralization type (i).
Fig. 6. Sulfide trace lead isotope ratios from mineralizations from the Rombak-Sjangeli basement window (Fig. 1) and the Nasafjall window. The ratios are recalculated for 430 My, except for the open symbols. For comparison, the field defined by the lead compositions of samples from the Sjangeli area (Fig. 5) are shown.

The lead composition of the Zn-mineralization is highly radiogenic although both samples (N12 and N13) come from a massive mineralization and have high Pb contents. In addition, the tie-line of the two samples N12 and N13 has a flat, negative $^{206}\text{Pb}/^{204}\text{Pb}$ slope. This suggests that the lead of the Zn-mineralizations represents a mixture of two non-cogenetic lead components.

The isotopic composition and source of the 'contamination' lead
The recalculated lead compositions of sulfides from the different mineralizations fall on different linear trends (Fig. 3; $^{206}\text{Pb}/^{204}\text{Pb} - ^{207}\text{Pb}/^{204}\text{Pb}$ diagram). If these lines represent mixing lines, this demonstrates that there is no common mixing component. Since these mineralization types are geographically separated and some of them are related to different lithologies (cf. Fig. 2), it is possible that the lead of the host-rocks has contributed to the mixing lead, which would give the mixing lead a local modification and variation.

For example, the lead composition of the type (i) Cu-mineralizations and their host-rocks, the tuffs, plot in fields which overlap (cf. Fig. 5). However, for the same $^{206}\text{Pb}/^{204}\text{Pb}$ ratio the tuffs have a wider variation of the $^{207}\text{Pb}/^{204}\text{Pb}$ ratio and usually also have higher $^{206}\text{Pb}/^{204}\text{Pb}$ ratios.

The sulfide lead composition could be explained with a simple two component mixing process. One lead component would have a low radiogenic lead composition, similar to sample M3 or the Koppåsen galena, while the other lead component has to be more radiogenic than sample M4 or M6. Setting
$^{206}\text{Pb}/^{204}\text{Pb}$ arbitrarily to 38.0, the other ratios are fixed at $^{207}\text{Pb}/^{204}\text{Pb} = 18.06$ and $^{208}\text{Pb}/^{204}\text{Pb} = 50.0$ (cf. Fig. 3). Such an end member could also partially explain the variation of the lead composition of the tuff lead composition. The remaining lead variability of the tuff lead composition would be due to another lead component, which formed in a low $\kappa$ environment, or would be due to in-situ lead growth in a low $\kappa$ environment. The $^{206}\text{Pb}/^{204}\text{Pb}$ ratio of such a lead component would indicate a potential source with an average Th/U atomic ratio of about 2.85 for the time interval from 1800 My to 430 My.

Alternatively, the lead variation of type (i) Cu-mineralizations could be due to accumulation of lead which was leached from the immediate wall rock. If the wall rock of all type (i) mineralizations had the same ratio, then the mixing with a highly variable lead extracted from the wall rock still yield a linear sulfide lead trend. If sulfides and the wall rock once had the same isotopic composition, then the mixing line would represent the time of isotopic homogenization of sulfides and wall rock, i.e., it would represent the time of formation or the time of a metamorphic homogenization. The anomalous widening of the metatuff lead composition field for more radiogenic lead in the $^{206}\text{Pb}/^{204}\text{Pb} - ^{207}\text{Pb}/^{204}\text{Pb}$ diagram could be due to the addition of low $\kappa$ lead, or a negative correlation of $\mu$ and $\kappa$.

This is compatible with the observation that the whole-rock lead composition usually is more radiogenic than the lead composition of the coexisting sulfides. The sulfides could represent a mixture of a low radiogenic lead component and a radiogenic lead component which was derived from wall rock leaching at about 430 My.

These two models can be tested in mixing diagrams (fig. 4). The sulfide samples from the type (i) Cu-mineralizations yield a good linear alignment in both mixing diagrams, $^{208}\text{Pb}/^{204}\text{Pb} - ^{1/208}\text{Pb}$ [ppm$^{-1}$] and $^{208}\text{Pb}/^{204}\text{Pb} - ^{1/208}\text{Pb}$ [ppm$^{-1}$]. The sulfide data could therefore be explained with simple mixing.

However, the tuff lead composition is not compatible with such a simple mixing. The mixing trend in the $^{208}\text{Pb}/^{204}\text{Pb} - ^{1/208}\text{Pb}$ [ppm$^{-1}$] diagram is scattered and delineates a field rather than a straight line. This field shows a negative correlation between the $^{208}\text{Pb}/^{204}\text{Pb}$ ratio and the $^{208}\text{Pb}$ content. In the $^{208}\text{Pb}/^{204}\text{Pb} - 1/^{208}\text{Pb}$ [ppm$^{-1}$] diagram, the tuffs form a wide sub-horizontal field (Fig. 4). The $^{208}\text{Pb}/^{204}\text{Pb}$ ratio does not vary systematically with the $^{208}\text{Pb}$ content of the samples, and therefore, there has been no mixing or the mixing component did not contain $^{208}\text{Pb}$. Alternatively, if all the metatuffs had been enriched in U for a long time, then the variation in radiogenity of the tuffs is mainly due to variable $\mu$ ratios for the different samples. However, to obtain a linear field (in Fig. 4), there should be at least a weak correlation between $\mu$ and $^{208}\text{Pb}$ ppm. Assuming a homogeneous lead composition of the tuffs at the time of infiltration and no external lead addition, the slope of the tuff field in the $^{208}\text{Pb}/^{204}\text{Pb} - ^{207}\text{Pb}/^{204}\text{Pb}$ diagram (Fig. 3) indicates a model age of 1910 My.

The obvious contrast between the sulfide lead and the tuff lead in the mixing diagrams (cf. Fig. 4) probably indicates a differential acceptance of externally derived lead in the sulfides and in the whole-rock sample. If only the sulfides received externally derived lead, while the silicates of the whole-rock samples did not accept additional lead, then the sulfide lead would show mixing lines, while the silicate lead would show in-situ lead growth. Since the tufts also contain a minor amount of sulfides, they would, according to the above reasoning, also have obtained an external lead addition, although only a minor one. However, for this model, the metatuff lead line has no strict age meaning, since there is an external, probably non-cogenetic lead component present.

The Fe-mineralizations also yield linear trends which could be interpreted as mixing lines. However, they received an isotopically different lead component (cf. Fig. 3).

**Lead from a low $\kappa$ source**

If an oxidized fluid infiltrates a permeable rock which reduces the fluid, then the U in the fluid would precipitate. Such a process could have caused the U enrichment of the tufts, the selective U enrichment of the pyrrhotite mineralizations and the pyrites of sample H7 (cf. flat 0-430 My tie lines in the $^{208}\text{Pb}/^{204}\text{Pb} - ^{207}\text{Pb}/^{204}\text{Pb}$ diagram Fig. 3).

For the metatuffs, such an U infiltration would have happened about 1600 - 1800 My ago, probably in connection with the Svecokarelian metamorphism and the intrusion of the Proterozoic granites and syenites into the Rombak-Sjangeli supracrustal rocks.
The infiltration of U into the pyrrhotite mineralization, however, happened about 430 My ago (cf. Fig. 3 and section on 430 My lead composition). The pyrrhotite line in Fig. 3 could represent a mixing line similar to the magnetite mineralization line and the type (i) Cu-mineralization line. However, in a mixing diagram (Fig. 4) the pyrrhotite points do not form a line. Since the primary lead contents of samples could have been different, this does not necessarily rule out mixing. The pyrrhotite line yields a model age for cogenetic lead of 1550 My and an apparent $\kappa$ ratio of 1.22. The slope of the pyrrhotite line in the $^{206}\text{Pb}/^{204}\text{Pb} - ^{207}\text{Pb}/^{204}\text{Pb}$ diagram (Fig. 3) indicates lead growth in an environment with a higher ratio than after the U addition.

Cu-mineralizations of type (ii) and type (iii) are characterized by low $\kappa$ lead. Since these mineralizations do not contain high $\kappa$ lead and the possible mixing components, which have been recognized, have developed in a high $\kappa$ environment, it is suggested that these mineralizations are older than the mixing event at about 430 My. These mineralization types (ii and iii) could have formed during the same event which caused the U infiltration into the tuffs. If they had acquired a high $\mu$ ratio at the same time, their low $\kappa$ lead could be due to in situ lead growth.

The isotopic composition and source of the 'primary' lead

The most primitive lead composition found in the Sjangeli area was found in sample M3. However, in the northward continuation of the Sjangeli area, the Kopparasen area, galena with a less radiogenic isotopic composition has been found. This galena lead falls on the same lead line as the type (i) Cu-mineralization lead, and probably represents an uncontaminated lead composition. Alternatively, the more radiogenic character of M3 could be explained by a complete or partial homogenization of the sulfide lead with its surrounding whole-rock at about 1600 - 1800 My.

Model ages for the galena lead composition vary from 1950 My (Stacey & Kramers 1975) to 2050 My (Zartman & Doe 1981). In a Zartman & Doe (1981) model, the galena lead would represent a mixture of lower crust lead with upper crust lead at about 2050 My. However, the galena composition could also be modelled with a mixture of an old lower crust lead and a younger upper crust lead. In this case, the older component could have preserved its low radiogenic composition as galena, from the time of separation to the time of mixing. At the time of mixing, this old lead would have mixed with a younger lead component. Using a upper crust lead of 1600 My age, the corresponding lower crust lead would have an age of more than 2200 My. For this model, the galena lead composition would give a minimum model age of the supracrustal rocks. Note that such model ages rely on the correctness of the model, and could yield model ages considerably higher than indicated by other methods (e.g. Vaasjoki 1981, Sundblad & Resholt 1986).

The lead isotope constraints on the alteration of the Sjangeli mineralizations

From the mixing lines and mineral tie lines (0 - 430 My) in Fig. 3, model ages can be calculated. These model ages are based on the assumption of cogenetic mixing components. The same assumption also allows for the calculation of an apparent ratio for the time from the separation of the two lead reservoirs to the time of mixing. The mineral tie lines reflect the in-situ $\kappa$ ratio which also can be calculated from the U and Th contents in Table 1. These model parameters are summarized for the various mineralization types in Table 2.

The variation of the model ages represents the effect of a variable non-cogenetic mixing component instead of different ages. However, the consistency of the model ages indicates that the source rocks probably were of Early to Middel Proterozoic age. Alternatively, there could have been an Early to Middle Proterozoic 'homogenization' event, during which the lead of the mineralizations and the source rocks of the 430 My lead homogenized. Both explanations are compatible with a Svecokarelian orogeny with the intrusion of granites and syenites (cf. Heier & Compston 1969, Gunner 1981, Gorbatschev 1985) and related metamorphism.

A comparison of the pre-mixing and post-mixing ratios shows a strong decrease for the pyrrhotite and probably the type (ii) Cu-mineralizations. This indicates that for these minerali-
Interpretation of the lead isotopic composition from sulfide

<table>
<thead>
<tr>
<th>model mixing age</th>
<th>pre-mixing</th>
<th>post 430 My</th>
<th>model time of Ul/Th separation</th>
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<td>2.05</td>
<td>0.27-3.02</td>
</tr>
<tr>
<td>Cu-type (ii)</td>
<td>1630(?)</td>
<td>2.05(?)</td>
<td>0.91</td>
</tr>
<tr>
<td>Cu-type (iii)</td>
<td>(2040)</td>
<td>(0.17)</td>
<td>0.056</td>
</tr>
<tr>
<td>Fe magnetite</td>
<td>2110</td>
<td>2.15</td>
<td>0.56-1.15</td>
</tr>
<tr>
<td>Fe pyrrhotite</td>
<td>1550</td>
<td>1.22</td>
<td>0.078-0.245</td>
</tr>
<tr>
<td>Zn</td>
<td>1700</td>
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</tr>
</tbody>
</table>

Table 2. Model mixing ages and apparent pre-mixing k (for a homogenization time of 1700 My) calculated from the data in Table 1 and fig. 3. Note that the model mixing ages and the apparent k ratios assume cogenetic mixing lead components. The post-mixing (430 My) ratios are calculated from Table 1. The model time of Ul/Th separation is estimated from changes in the lead growth trends at c. 430 My and c. 1800 My respectively. Values in parentheses (type iii) represent maximum model ages and minimum pre-mixing ratios (only one sample was recalculated). The type (ii) Cu-mineralization data in parentheses are inferred from type (i) Cu-mineralization data.

zations a separation of U and Th occurred at approximately 430 Ma. The contrast of solubility of U and Th in oxidized fluids (e.g. Boyle 1982) suggests a hydrothermal transport. The pre-mixing k ratio indicates an upper crustal lead source, since other reservoirs have higher k ratios, e.g. bulk Earth 4.2 (Allègre et al. 1986), lower crust ≈5.9 - 10 (Heier & Thoresen 1971, Dostal & Capedri 1978, Zartman & Doe 1981) and mantle ≈3 - 4 (Zartman & Doe 1981).

The anomalously low thorogenic lead composition at 430 My of the type (iii) Cu-mineralizations and the tuffs (cf. Figs. 3 & 5) indicates that there had been an earlier separation of U and Th, which was possibly related to the Sveckoarelian metamorphism. This 1600 - 1800 My metamorphism probably represents the time of formation of the remobilized Cu-mineralizations of type (ii) and type (iii).

Indications that the metamorphic fluids were oxidized and could have transported U come from the chemical composition of epidotes from the tuffs and lava flows (cf. Romer 1987). Epidotes from the Sjangeli area have a Xpistacite ≈33, and indicate that they have equilibrated with an oxidized fluid (Bird & Helgeson 1981). In such a fluid, U is predominantly in the 6+ state (Langmuir 1978, Remberger 1984), and therefore mobile.

The lead compositions of the Cu-mineralizations and the Fe-mineralizations differ (Fig. 6). Since these two mineralization types are related to different lithologies, the isotopic difference may indicate that there was some wall rock leaching or that the lead composition of the fluid was different in the various mylonitic zones.

No isotopic constraints can be given for the genesis of type (i) Cu-mineralizations, since there probably had been a rather extensive isotopic homogenization during the Svecoarelian orogeny. However, the stratiform nature and their selective occurrence in the tuffs indicates a possible syngeneric origin in a subaerial shallow-water depositional environment. Such an environment is indicated by the occurrence of tuffs (Romer 1987) and lapilli (Adamek 1975) in the Kopparåsen area (see Fig. 1).

Comparison of other sulfide lead compositions with the Sjangeli sulfide lead

There are only a few lead isotope analyses available from other supracrustal units of the Rombak window. Samples from mineralizations have been provided by Are Korneliussen (NGU, Trondheim) from Sildvik and Sørådal and by Frank-Dieter Priesemann (Folldal Verk A/S, Folldal) from Gautelis. The other data consist of analyses from galena mineralizations from Svanåreivo (Johansson 1983) and from Narsfjöll (Johansson 1983), a basement window about 300 km south of the Sjangeli area.

The limited amount of data from these supracrustal units (Table 1) does not allow for an equally thorough discussion of the data as for the Sjangeli area. Instead, these samples are compared with the sulfide samples from the Sjangeli area (Fig. 6).

(1) The lead composition of massive sulfide mineralizations from supracrustal belts within the Rombak-Sjangeli window show the same variation as the type (i) Cu-mineralizations of the Sjangeli area, and could therefore have been affected by similar processes. The Svanåreivo samples (Johansson 1983), however, have significantly higher 206Pb/204Pb ratios at a given 208Pb/204Pb ratio, and further the 207Pb/206Pb ratios are slightly higher than for the type (i) Cu-mineralizations.

Johansson (1983) interpreted the lead composition of the Svanåreivo deposit (20 km north of Sjangeli) as the result of the addition
of a preferentially leached, radiogenic, Karelian lead component. The lead isotopic composition of granites form the Sjangeli area shows the same thorogenic lead as the Svanqeraive deposits (cf. Figs. 5 & 6), and therefore it is possible that lead leached from granites and syenites with a similar lead composition as the Sjangeli granites could have been involved in the formation of the Svanqeraive galena mineralizations. However, if the lead composition of the leached lead and the old lead component were non-cogenetic, model ages such as c. 1970 My for Svanqeraive and c. 1660 My for Nasafjäll (Johansson 1983) do not give the age of the source.

(2) The lead addition during the Caledonian metamorphism seems to be common for basement mineralizations throughout the entire Rombak-Sjangeli basement window. The degree of observed lead radiogenicity depends not only on the amount of circulating fluid and the effectiveness of lead precipitation, but also on the lead concentrations and the total amount of lead present in the mineralizations. The low radiogenicity of lead from Kopparåsen and Sildvik may reflect that their original lead composition is less contaminated by a radiogenic lead component.

Conclusions
The trace lead of sulfides from Cu-, Fe-, and Zn-mineralizations from the Sjangeli area show at least two events of isotopic contamination and homogenization during which U and Pb were mobile.

(1) At about 1600 - 1800 My stratiform Cu-mineralizations became remobilized into vein deposits. Some of these discordant deposits also acquired high μ ratios and low k ratios. The sulfide lead of the stratiform Cu-mineralizations and their host-rock, the basic metatuffs, was probably re-equilibrated, and the tuffs obtained U.

(2) At about 430 My, externally derived highly radiogenic lead was added to the sulfides of the various Fe- and Cu-mineralizations, as seen by mixing lines. Further, U was added to massive pyrrhotite mineralizations.

(3) The two events of isotopic disturbance are related to the Svecokarelian orogeny and the Caledonian orogeny. Both events resulted in metamorphism with the redistribution of lead and uranium. The generated fluids must have been sufficiently oxidized to transport U.

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