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Palaeoproterozoic Tectonics and Metallogenesis:

Comparative analysis of parts of the Australian & Fennoscandian Shields

Edited by

R. W. R. Rutland & Barry J. Drummond

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FOREWORD

This workshop arose from an individual project, initiated in 1994, under the auspices of the Australian Geodynamics Cooperative Research Centre (AGCRC). Initially the focus was on comparison between the Pine Creek Domain in the North Australian Province and the North Sweden sub-province of the Svecofennian Province. It became evident, however that a wider comparison of the whole pre-1800 Ma provinces was desirable.

Fieldwork in Sweden and Finland was undertaken with the willing co-operation and assistance of the Geological Surveys of Sweden and Finland, and, in particular with Dr. Stefan Bergman, Dr. Torbjörn Skiöld and Prof. David Gee in Sweden, and Prof. Ahti Silvennoinen in Finland.

The support of those organisations and individuals in that work and also in supporting this workshop is gratefully acknowledged, as is the support of the Swedish Natural Science Research Council. In Australia the significant financial support of the Department of Industry Science and Tourism is also gratefully acknowledged.

The project, and this workshop in particular, have also been sponsored by the AGCRC, the Australian Geological Survey Organisation (AGSO), the Northern Territory Geological Survey (NTGS, who have also produced the guide-book for the field excursion) and the Research School of Earth Sciences and the Department of Geology at the Australian National University. Their support is greatly appreciated.

The workshop aims to analyse the remarkable similarities and significant differences between the Palaeoproterozoic provinces in North Australia and Fennoscandia; and allow a more critical evaluation of the diverse models of tectonic evolution and metallogenesis which have been proposed. This in turn will assist assessments of prospectivity and future exploration

The focus is on the pre-1800 Ma. provinces and their ore deposits. The intention is that discussion of various lines of geochemical and geophysical evidence will be based on the most recent geological mapping and geophysical data-sets, and constrained by modern geochronology. The workshop brings together Australian experts with those from Finland and Sweden. They have been invited from the national geological surveys, and from other organisations, to ensure coverage of all relevant fields. The workshop is designed to provide maximum interaction between these experts and members of the Australian exploration industry.

Session 1 of the workshop is designed to provide an up-to-date factual overview of each region as a basis for the comparative thematic analyses and interpretations planned for Session 3 (after the field excursion). This abstract volume consists mainly of contributions invited for Session 1, but with some additional abstracts prepared for Session 3. It will be determined at the workshop what further publication, if any, should arise from the discussions.

R.W.R. Rutland

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Palaeoproterozoic of Northern Australia and Fennoscandia: an Introduction

R.W.R. Rutland

Australian Geodynamics Cooperative Research Centre
Research school of Earth Sciences, Australian National University
Canberra, ACT 0200, Australia.

Major Proterozoic Provinces are the preserved remnants of the presumed much more extensive continental crust during the Proterozoic Era. Within any one province the record is incomplete due to limitations in lateral extent, and limitations resulting from the present level of erosion or from superposed younger events. It is also likely that there has been selective preservation of particular crustal environments but it is supposed that the combined global record from the preserved elements will allow a reasonable reconstruction of Proterozoic crustal evolution.

There are however more particular reasons for comparative study of the Palaeoproterozoic provinces of Fennoscandia and northern Australia, as outlined below. In general, Northern Australia can be of interest to those studying the Fennoscandian Shield because the Palaeoproterozoic provinces show a relatively clear and simple record, particularly of stratigraphic relationships, and they are largely free from the very strong younger overprinting which is common in Fennoscandia. On the other hand, Fennoscandia probably has the most comprehensive and best-studied geological record, and in particular it displays the relationships of the Palaeoproterozoic provinces to the Archaean probably better than anywhere else. This workshop is designed to focus on the pre- 1800 Ma provinces, and, in particular, on the Svecofennian and North Australian provinces formed in the relatively short Orosirian Period between 2050 and 1800 Ma.

The two provinces are broadly comparable in overall dimensions. The North Australian Province (NAP) is roughly 2000km (in the WNW-ESE direction) by 1000 km (in the SSW-NNE direction). But much of this area is overlain by post-1800 Ma Proterozoic, and /or by relatively thin Phanerozoic cover. The pre-1800 orogenic domains allocated to the NAP are therefore widely separated and of limited size. The Pine Creek domain, for example , has dimensions of roughly 300 by 200 km (similar to those of the Swedish sub-province north of Skellefte). The Svecofennian of the Fennoscandian Shield forms a triangular area with sides about 1000 km long, but a similar area lies beneath the East European Platform to the south (e.g. Gorbatschev and Bogdanova, 1993). The Karelian Province in the Shield, which has an Archaean basement but which also suffered Svecofennian orogeny, adds an area roughly 1000 by up to 300 km.

An important difference between the two regions appears to be in the level of erosion, even though both regions are characterised by areas of unusually thick high velocity lower crust (e.g. Goncharov and Drummond, this volume). In Fennoscandia, there is a general lack of the post-1800 Ma supracrustal sequences which are so important in Northern Australia. Inevitably therefore the world- class mineral deposits which occur in the post- 1800 Ma sequences in Australia are absent in Fennoscandia. Conversely, in the Svecofennian Province there are larger areas of high grade metamorphic rocks and more abundant early granitoid intrusions (1890 - 1860 Ma). There is also a wide range of economically significant mineral deposit types (e.g.Gaál, 1990; Ekdahl, 1993; Peltonen, 1995; Frietsch et al., 1997) which are much less well represented in the North Australian Province. One of the major purposes of this workshop is to examine the reasons for these differences in the context of the regional geology and tectonic setting, some aspects of which are introduced below.

Many Palaeoproterozoic orogenic provinces, including most of those in the North Atlantic region, are flanked on both sides by Archaean crust and are interpreted in terms of Collision Orogeny. The Lapland Granulite Belt is a Fennoscandian example, and the Gascoyne Province an Australian example. The Svecofennian and North Australian provinces, however , have different, non-linear characteristics and share a tectonic setting between Archaean provinces and younger linear collision complexes which were active through much of the Proterozoic after 1850 Ma (the Sveconorwegian belt in Fennoscandia and the complex of orogenic belts in the Amadeus Transverse Zone in Australia). In both cases these younger complexes may encompass the sites of Pacific type plate margins which were active prior to 1850Ma.

Relationships to Archaean provinces

Palaeoproterozoic provinces are of special importance in showing the transition from Archaean to Proterozoic tectonic styles. In the North Australian Province the exposed orogenic domains are of limited extent and the relationships to Archaean craton or to contemporary active continental margins are not available for study, although it is probable that the Kimberley Basin is underlain by Archaean rocks (the relations of the Palaeoproterozoic Gascoyne Province to the Archaean of the Pilbara Craton are well displayed but that province is separated from the North Australian Province by the younger orogenic belts of the Amadeus Transverse Zone [e.g. Rutland, 1976]). There is therefore particular interest in the evidence from Fennoscandia, where the relations between Archaean (Karelian) and Paleoproterozoic (Svecofennian) orogenic provinces are well displayed, especially in Finland (e.g. Gaál and Gorbatschev, 1987).

The Fennoscandian Shield lacks the clear distinction between a major late Archaean orogenic event (at 2.6 - 2.7 Ga) associated with extensive plutonism, and younger unconformable early Proterozoic cover sequences, such as characterises both the Canadian and Australian Shields. Instead, the Karelian Province displays early Proterozoic as well as Archaean greenstone belt development. It also displays substantial reworking and plutonism during the development of the Svecofennian Province to the south-west. Consequently, various definitions, both of the provinces and of the orogenies have been proposed.

Traditionally, rocks of the Palaeoproterozoic Svecofennian orogeny were separated from the Palaeoproterozoic and Archaean rocks of Karelia to the north-east on the basis of the near N-S orogenic trends in the Karelian province and the near E-W trends in the Svecofennian province.

It was soon recognised, however, that many of the structures in the Karelian Province were also of Svecofennian age and the term Svecokarelian has often been applied to the orogeny.

More recently, Gaál and Gorbatschev (1987) suggested that the *stratigraphic* boundary between the pre-orogenic Karelian and orogenic Svecofennian assemblages should be drawn at the boundary between the Jatulian and Kalevian Groups, to which they allocated dates of 2.3-2.1 and 2.0 Ga., respectively. However, these groups occur together in a number of areas, and such areas are now normally included in the Karelian Province in Finland even though they may be involved in strong thrusting and metamorphism of Svecofennian age. The 1996 map, produced by collaboration between Finland, Norway and Sweden in the Mid-Norden Project, adopts a boundary (usually tectonic) between different stratigraphic facies of Svecofennian rocks as the boundary between Karelian and Svecofennian provinces. Thus, rocks previously allocated to a North Svecofennian Sub-province in Finland are now included in the Karelian Province. Others have attempted to define a province boundary at the inferred southern limit of Archaean basement (e.g. Kärki et al. 1993)

These different approaches arise from the fact that the Svecofennian and Karelian provinces do not fit comfortably into Phanerozoic models of the relationship between orogenic belts and their forelands. There is no clear tectonic front to the Svecofennian orogenic structures and there is not a dominant fold trend parallel to the cratonic margin. The Svecofennian orogenic structures have varying trends, and strong N-S structures affect both provinces. More remarkably, the Svecofennian granitic plutons extend far into the Karelian Province, so that there is a strong contrast between the SE (Russian and Finnish) and NW (Swedish) parts of the latter province.

The structural and stratigraphic transition from the Karelian to the Svecofennian provinces is reviewed in this volume by Sorjonen-Ward and Pajunen for Finland and by Bergman for Sweden. It is especially important for the understanding of the tectonic setting of ore deposits. Ekdahl (1993) notes, for example, in relation to the Raahe-Ladoga Zone, that *"The effectiveness of exploration ..... has been continually impeded by an inadequate understanding of stratigraphic relationships. This results from a combination of intense deformation, high metamorphic grade and the paucity of distinctive or laterally persistent marker horizons. The area belongs to the so-called Savo Schist Belt..... which straddles the boundary zone between the Svecofennian and Karelian* *orogenic domains. The complexity of this zone has made correlation between the two domains difficult, so that an improved understanding of lithostratigraphy is of general geological as well as economic importance."*

The differences in the Karelian Province between Finland and Sweden are accompanied by strong differences in th Svecofennian Province across the Gulf of Bothnia. The thick high velocity lower crust, identified in Finland is absent in Sweden (e.g. Korja et al., 1993) and the gravity trends are also distinctly different (Ruotoistenmäki, T., and others, 1997). These differences are reflected in the geology and correlations across the Gulf of Bothnia have not been well established.

Relations to younger orogenic belts

In Australia a high grade metamorphic belt , dated at about 1770-1780 Ma, and associated granitic intrusions dated between 1770 and 1710 Ma mark the Strangways Orogeny, in the northern part of the Arunta Inlier (Collins & Shaw, 1995; Zhao & Bennett, 1995). Foreland folding, associated with the orogeny extends northwards though the Davenport Province, and into the Tennant Creek Province (Donnellan et al., this volume). It seems possible that the Strangways Orogeny is superimposed on an earlier active continental margin to the North Australian Province, some evidence for which has been adduced from the calc-alkaline character of the ca.1880 Ma Atnarpa Igneous Complex (Zhao and Cooper,1992), further south in the Arunta Inlier.

In Sweden, the N-S trending deformation front of the Sveconorwegian orogenic belt clearly cuts the E-W Svecofennian structures (Stephens et al., 1994). The Trans-Scandinavian Igneous Belt (TIB), dated between 1.85 - 1.69 Ga, and the similar Revsund-Sorsele Suite (1.8 - 1.78 Ga) are normally included within the Svecofennian orogenic domain. However, they also form a roughly meridional belt which cuts the main Svecofennian structures and which is partly incorporated in the eastern zone of the Sveconorwegian orogenic belt.

Fennoscandia is also characterised by important deformation and metamorphism superimposed on the pre-1850 Ma sequences, especially in southern Finland and southern Sweden at about 1840-1830 Ma (Korsman et al., 1988, Koistinen et al., 1997, Ehlers, this volume).

Most of the structures within the Svecofennian orogenic domain proper pre-date 1.83 Ga, but the structures do include discrete late shear zones with near E-W or SE-NW trends, and which are associated with granites dated at about 1.82 Ga and which cut intrusive rocks as young as 1.78 Ga. More generally, in extensive areas of both the Svecofennian and Karelian provinces there is pervasive younger S-type granite and pegmatite dated between 1.82 and 1.75 Ga which is clearly in part contemporaneous with the plutonism in the TIB.

North Australia has much less deformation and plutonism superimposed after intrusion of the main suite of granitoids associate with the Barramundi orogeny (ca.1870-1850 Ma.). Younger shear zones nevertheless are important in some domains and may exercise control on mineralisation.

The development of the Mt. Isa Province on the eastern margin of the North Australian Province was also superimposed on an earlier pre-1850 Ma domain. There, several episodes of rifting and granite plutonism are terminated by orogeny at about 1610 - 1550 Ma (Wyborn et al. 1988). It can be suggested that the South Finland province has been produced by much the same (extensional) tectonic processes as the Mt. Isa Province. The similarity is not evident because of the different levels of erosion between the two regions. In Mt. Isa the post-1850 Ma supracrustal sequences are preserved, while in South Finland the supracrustal sequences which must have existed during the main phase of metamorphism between 1830 - 1840 Ma have since been removed. It is notable that in both cases the orogenic trend is roughly at right angles to the post-1800 orogenic complexes, which as noted above, may conceal earlier plate boundaries.

Tectonic evolution

Both the Svecofennian and the North Australian provinces show very rapid tectonic evolution of similar character over large (non-linear) areas. In both regions the oldest sequences have been related to rifting which in Finland is evidenced by several phases of mafic magmatism dated between 2.1 and 1.95 Ga and in both regions the main orogenic development has occurred between 1920 and 1850 Ma, although, as noted above, later orogenic episodes have been superimposed in some areas.

Nevertheless, it is not to be expected that there will be precise synchroneity of events between the two provinces during their evolution. What is remarkable is that the qualitative changes during that evolution are broadly the same in both regions and qualitative comparisons can be made between similar episodes even if they are not of precisely the same age. For example, the main phase of granitoid emplacement is approximately 1890 - 1860 Ma in Finland and 1870 - 1850 Ma in Australia, but in both regions these granitoids have been interpreted as derived from an underplate after an important immediately preceding period of compressive deformation and high temperature metamorphism (Wyborn, 1988; Lahtinen, 1994) In both regions, the felsic rocks evolve geochemically through time (e.g. Wyborn et al, 1987) and there are many similarities in the qualitative changes involved.

Similarly, in the North Australian orogenic Province, there seem to be close qualitative similarities in the basin evolution in the various domains, so that a consistent sequence of rift , sag, and flysch phases can be identified (Etheridge et al., 1987), but there is not precise synchroneity between the different domains. A similar sequence appears to occur in the Svecofennian. Precision in the geochronology is therefore required to identify diachroneity of evolutionary events and to assist in relating them to geodynamic processes (see Skiöld and Page, this volume).

Tectonic Models

In view of the similarities in tectonic setting, age and evolution of the two regions it is remarkable that quite different tectonic models have been applied. There is a widespread view that the later part of the Paleoproterozoic was characterised by very rapid growth of juvenile crust (Patchett and Arndt, 1986). For the Svecofennian this view has generally been combined with subduction-related models of lateral accretion (e.g. Gaál and Gorbatschev, 1987). In contrast, in Australia, the major Paleoproterozoic provinces have been interpreted as largely ensialic (Etheridge et al, 1987), so that crustal growth, by the addition of juvenile material, is essentially by vertical accretion. The workshop should permit critical evaluation of the evidence for these different hypotheses, in the context of the probable differences in crustal responses to plate tectonic processes because of the secular change in the thermal regime.

Recently there has been an attempt to develop a general model for North Atlantic provinces which takes account of evidence from seismic reflection profiling, but which also satisfies interpretations based on geochemical evidence (Snyder et al 1996). This model is heavily dependent on the Svecofennian of Fennoscandia, notably the Skellefte district, and on the Babel lines in the Gulf of Bothnia. They conclude that *"The overall geometry of the Orogen suggests that the Svecofennian arc(s) were accreted ('obducted') as crustal flakes and were subsequently translated northward above the footwall Archaean lithosphere."* Tectonic relations across the Gulf of Bothnia are complex, however, and assumptions about the strike and age of various tectonic elements are suspect.

A general objective of this Workshop is to test these models of the tectonic setting of mineral deposits by assessing various lines of geophysical, petrological, geochemical and isotopic evidence against relatively detailed structural and stratigraphic frameworks constrained by modern geochronology. There is, of course, a major scale problem, and it can be very difficult to relate the cartoon models to the geology that can be seen in the field or observed on 1:50,000- 1:250,000 scale maps. However, modern aeromagnetic mapping has provided a powerful tool both in Fennoscandia and in Australia.

It seems possible that a re-evaluation of the models in the light of the constraints imposed by improved knowledge of regional relationships may have considerable implications for our understanding of the distribution of mineral deposits. For example, it now appears that some of the felsic volcanic sequences in northern Sweden and Finland, with their important VHMS deposits, post-date the main compressional orogenic event (e.g. Rutland et al. this volume), as do the felsic volcanics of the Cullen Event in Australia (Jagodzinski, this volume). Their tectonic setting may therefore have more in common than current published interpretations suggest.

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Pine Creek Orogenic Domain : Structural and Stratigraphic Framework

Masood Ahmad

Department of Mines and Energy
Northern Territory Geological Survey
PO Box 2901
Darwin, NT 0801, Australia

The Palaeoproterozoic Pine Creek Orogenic Domain forms part of the North Australian Craton and is the most significant metallogenic province of the Northern Territory. This Domain unconformably overlies late Archaean(~2500 Ma) granites and metasediments and contains an alternating sequence of psammitic and pelitic sedimentary strata, tuff, minor volcanics and dolerite sills which were deformed and metamorphosed during the Top End Orogeny at 1870-1855 Ma and intruded by syn to late orogenic granites. The Pine Creek Orogenic Domain hosts a variety of mineral commodities including uranium, gold, tin-tungsten-tantalum, lead-zinc-copper, iron and manganese. Many of these deposits are spatially and genetically associated with granite intrusives and are located within their contact aureole.

Discussion

The geology, stratigraphy, metamorphism and structural setting of the Pine Creek Orogenic Domain are detailed in several previous studies (Walpole et al. 1968; Needham et al. 1980; Stuart-Smith et al. 1993; Ahmad et al. 1993). The metamorphosed and deformed Palaeoproterozoic Pine Creek Orogenic Domain sequence is exposed over an area of ~66,000 km^2 and is unconformably overlain by the relatively undeformed Palaeo- to Mesoproterozoic McArthur Basin to the east and by the Victoria Basin to the west and southwest. Phanerozoic strata cover the sequence towards the south and north. Stratigraphic correlations based on previous studies are summarised in Figure 1 and simplified geology and major mineral deposits are shown in Figure 2.

From west to east, the Pine Creek Orogenic Domain can be divided into five sub- regions, ie. Litchfield Province, Rum Jungle Region, Central Region, South Alligator Valley and Alligator Rivers Region. The Litchfield Province represents an area of isoclinally folded amphibolite to greenschist facies metamorphics. The Central Region exhibits sub greenschist facies metamorphics and simple structures dominated by upright northwest and north trending folds. The Rum Jungle Region represents Archaean metamorphics and granitic basement and exhibits polyphase upright folds, domes and basins. The South Alligator Valley Region is characterised by lower greenschist facies metamorphism and north to northwest trending upright folds. The East Alligator Rivers Region is characterised by amphibolite facies metamorphism and upright to recumbent folds.

The sequence unconformably overlies late Archaean (~ 2500 to 2700 Ma) granitic basement which is exposed near Batchelor (Rum Jungle and Waterhouse Complexes) and Jabiru (Nanambu Complex) and intersected in drill holes near Woolner (Woolner Granite). The Pine Creek Orogenic Domain comprises an alternating sequence of psammitic, and pelitic rocks with minor carbonate sediments and volcanics. Dolerite sills (Zamu Dolerite and equivalents) intruded the sequence prior to deformation and metamorphism.

The age of the Pine Creek Orogenic Domain sequence is constrained between 2470 and 1870 Ma (Page et al. 1980). Age determinations (U-Pb zircon) on tuffs in the Mount Bonnie Formation gave a depositional age of 1885±2 Ma (Needham et al. 1988). Regional metamorphism and deformation, during which the strata were tightly folded and faulted, followed sedimentation. Regional metamorphic grades range from sub-greenschist facies in the Central Region to upper amphibolite facies along the western (Litchfield Province) and eastern (Alligator Rivers Region) margins. This period of deformation and metamorphism is known as the Top End Orogeny

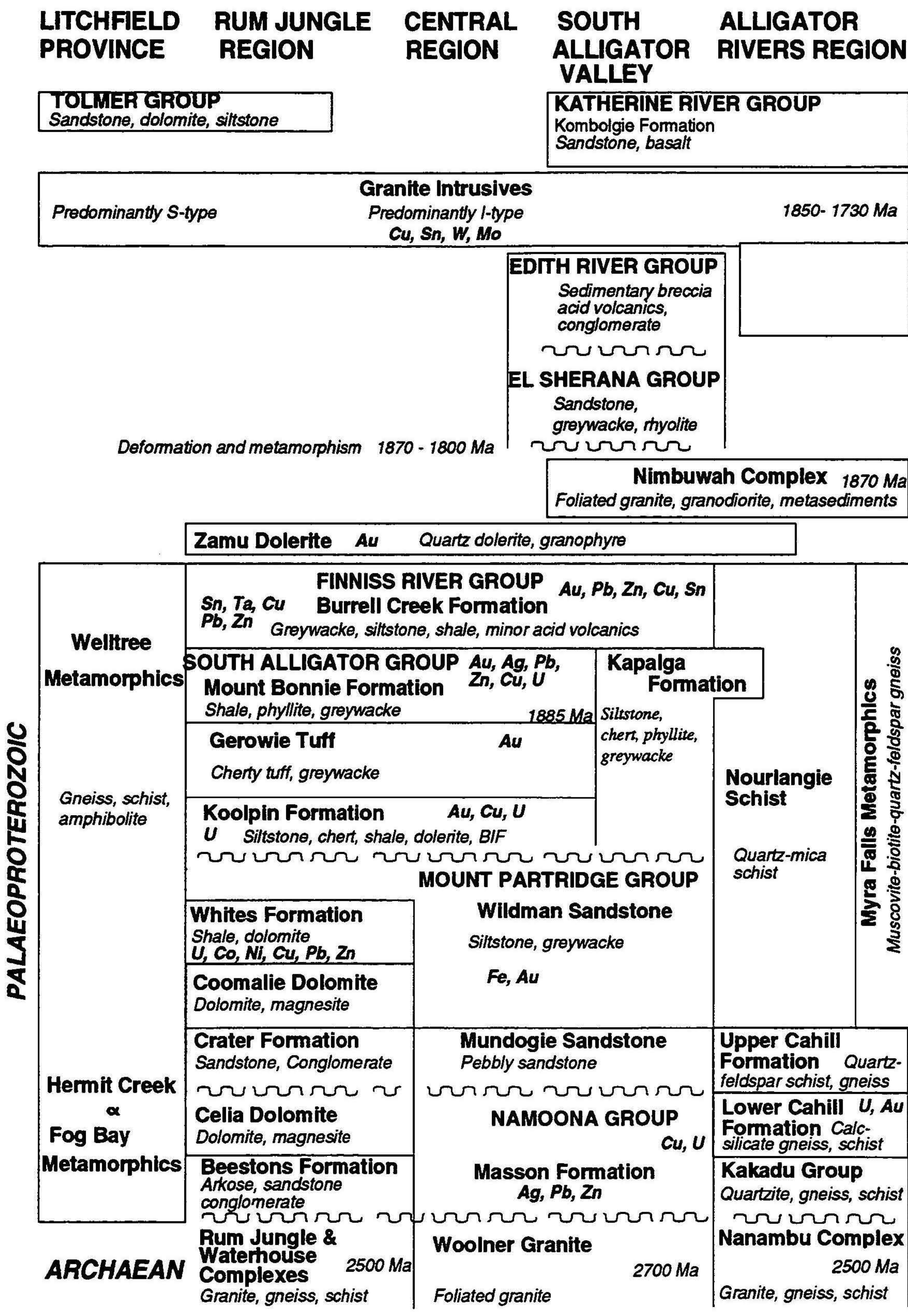

Figure 1 Stratigraphic correlations, lithology and mineralisation in the Pine Creek Orogenic Domain

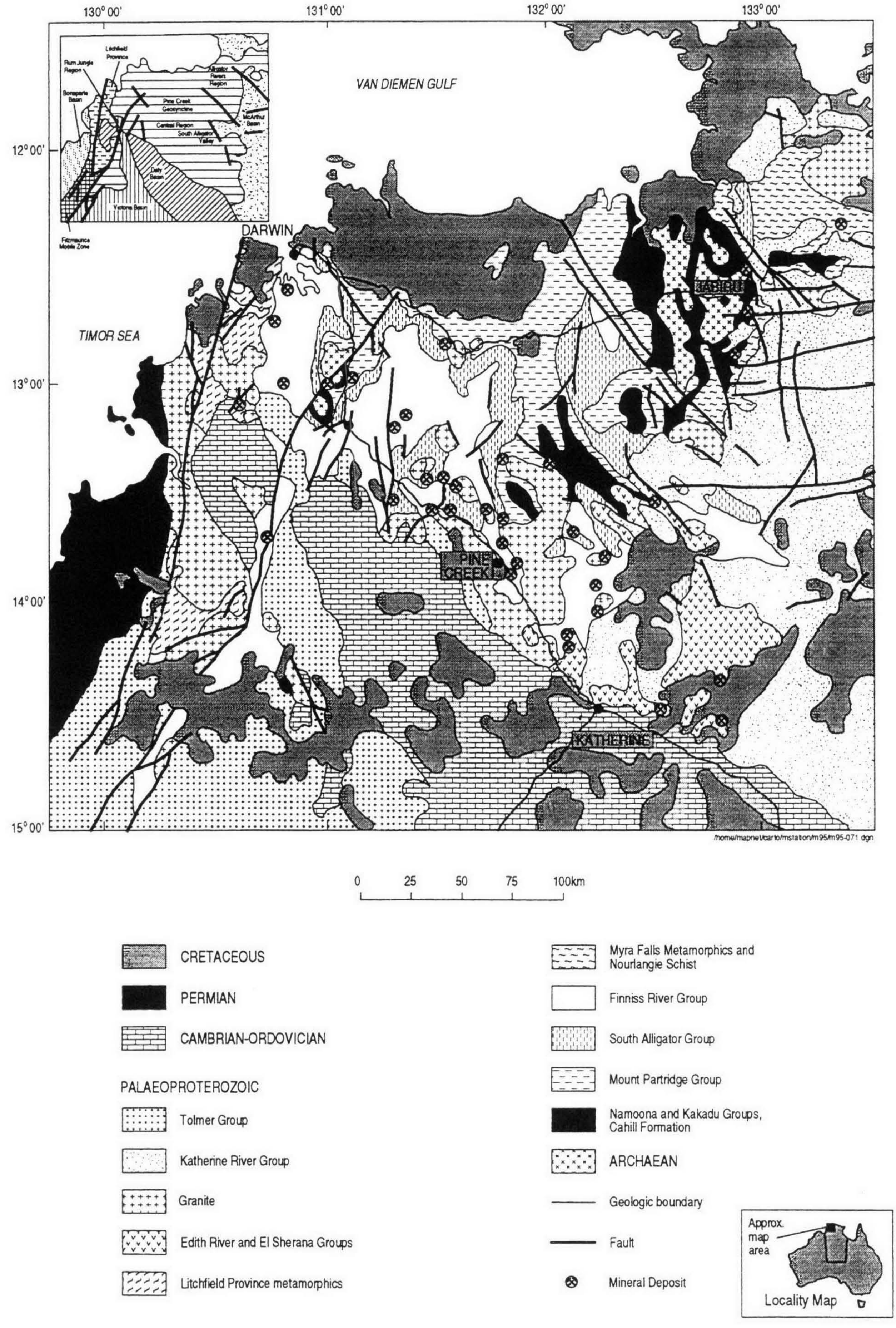

Figure 2 Regional geology and mineral deposits in the Pine Creek Orogenic Domain

and is constrained at 1870-1855 Ma (Needham et al. 1988).

The earliest deformation D_1 caused north trending monoclinal wrapping, recumbent north trending isoclinal folding and tectonic sliding. The D_2 deformation is represented by thrusting and recumbent folding which is more common in the Litchfield Province and the Alligator Rivers Region. The following D_3 deformation event is more widespread and is represented by north to northwest trending non-cylindrical, close to tight folds. These folds are an important structural control for the vein type mineralisation in the Central Region. The D_4 deformation phase is associated with the granitic intrusions and has produced open small amplitude large scale folds which are responsible for the formation of elongated basins and domes. The D_5 deformation has produced steeply plunging polyclinal kinks and drag folds associated with dextral movements along major faults.

A period of granite intrusion, contact metamorphism and minor volcanism dated at 1800-1850 Ma accompanied and followed the regional metamorphism and deformation. Based on age and composition, Stuart-Smith et al. (1993) recognised three broad groups of granites in the Central Region: an older group dominated by more mafic granite phases, followed by concentrically zoned granites and leucogranites, and a youngest felsic granite phase. Stuart-Smith et al. (1993) summarised geochronological data on these granites and concluded that U-Pb zircon ion probe ages range from 1800 to 1835 Ma and are more precise than either Rb-Sr whole rock (~ 1780 Ma) or conventional U-Pb zircon ages (1860 to 1750 Ma). Granites of the Central Region have predominantly I-type characteristics and have magnetite as a common accessory. Magnetic susceptibilities range from 10^{-3} to 10^{-4} emu/g (Bajwah 1994). Many of these granites appear to be minimum melts and indicate crystal fractionation (Ferguson et al. 1980). Mineralogical and petrological data on granites in the Litchfield Province suggest similarity to S-type granites (Ahmad 1995).

Albite-epidote hornfels is present in all contact aureoles, commonly with a narrower, inner continuous zone of hornblende-hornfels, and the effects of contact metamorphism are seen up to 10 km away from the exposed granite contact. Geophysical data (Lewis et al. 1995) suggest that granitic rocks are virtually everywhere present in the Pine Creek Orogenic Domain at a depth of 1-5 km. The geophysical data, however, cannot distinguish between the Archaean basement and the Palaeoproterozoic granites.

Tectonic evolution of the Pine Creek Orogenic Domain is described by Needham et al. (1988). An extension event at about 2000 Ma resulted in the formation of a basin in which about 10 km of clastic, organic and chemical sediments were deposited. Initial depositional environments ranged from neritic to intertidal to fluviatile followed by flysch-like sedimentation towards the end. During the early depositional stage (Namoona and Mount Partridge Groups) Archaean palaeohighs at Rum Jungle in the east, Woolner to the north and Nanambu to the east formed islands and were probably a major sediment source. Sedimentation during this stage, which has been considered as rift phase, includes fluvial to shallow marine conglomerate and arkose succeeded by supratidal to intertidal carbonate facies. The overlying sag phase sediments are represented by the South Alligator Group which was deposited under shallow marine, low-energy conditions and includes pyritic carbonaceous shale, chert, carbonate, banded-iron-formation, tuff, and siltstone. This is followed by the Finniss River Group represented by a monotonous flysch sequence of greywacke and siltstone deposited in high-energy, deeper marine environments. Sedimentation and lithification was followed by intrusion of dolerite sills.

A period of extension succeeded the Top End Orogeny, resulting in the formation of grabens in the southeast. Two unconformity bounded sequences (El Sherana and Edith River Groups) of sediments and volcanics were deposited in these grabens (Needham et al. 1988).

The Central Region, in which the I-type granites dominate, contains the majority of gold, base metal and tin-bearing veins, as well as stratabound gold and polymetallic deposits. The Rum Jungle and Alligator Rivers Regions are areas of partly exposed and shallow Archaean granitic basement and contain significant uranium deposits. The Litchfield Province is characterised by a prevalence of S-type granites which are related to the development of Sn-Ta bearing pegmatites (Ahmad et al. 1993).

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Structural Framework in Northernmost Sweden

Stefan Bergman

Geological Survey of Sweden
PO Box 670, S-751 28 Uppsala, Sweden
stefan.bergman@sgu.se

In the Fennoscandian Shield, successively younger stages in the early tectonic evolution of the continental crust in Northern Europe are recorded from Archean orogens (>3Ga) in the northeast through the Svecokarelian and Gothian orogens, to the Sveconorwegian orogen (c.1Ga) in the southwest. In northernmost Sweden, there is opportunity to study Precambrian events in the time span 2.8-1.8Ga.

The general NNW to NNE structural grain (Figs.1 & 2) in the region is Proterozoic in age. Regional Archean structures are not obvious due to this overprinting. On outcrop scale, foliated clasts of Archean rocks in an unconformably overlying conglomerate and cross-cutting mafic dykes (2.2 Ga) prove the existence of Archean deformation and metamorphism. The age of this event is ca 2.7 Ga, as shown by geochronology.

In Northeastern Norrbotten, some controversial high-grade gneisses are on many maps interpreted as Archean. New field data suggest that most orthogneisses are comparable to the 1.88 Ga old Haparanda Suite granitoids and that the deformation is Svecokarelian in age. The high-grade metamorphism was possibly caused by a mafic intrusion shown by a large positive gravity anomaly. Local banded gneisses with more complex structures than the surrounding Haparanda Suite may be the only Archean remnants in that area.

Rifting in the late Archean to early Proterozoic produced voluminous mafic volcanic rocks (e.g Kiruna Greenstone Group) that host economic ore deposits (e.g Viscaria). A new U-Pb zircon age determination shows that this mafic volcanism was still active at 2.09 Ga. Indirect evidence for deposition of evaporites is provided by a regional Cl-anomaly in glacial deposits and widespread polyphase scapolitization. Locally, a strong fabric in mafic tuffs not shared by overlying Svecofennian felsic volcanic rocks may indicate a deformation phase between 2.1 and 1.9 Ga.

Svecofennian mafic to felsic volcanic rocks formed in connection with tectonic processes along a convergent plate margin. They can be divided into the lower calc-alkaline Porphyrite Group and the overlying Kiruna Porphyry Group. These volcanic rocks contain the iron ores at Kiruna and Malmberget. Two separate intrusive suites (1.9-1.86 Ga) are broadly coeval with the volcanic rocks; the calc-alkaline Haparanda Suite granitoids and the slightly younger Granite-Syenite-Monzonite Suite (GSM). An early phase of the Svecokarelian orogeny affected the Haparda Suite whereas rocks of the GSM are generally isotropic. This deformation is earlier than the main phase of deformation and metamorphism in central Sweden, which occurred after c. 1.85 Ga. there is evidence for regional deformation younger than 1.8 Ga also in northernmost Sweden. The strain pattern is generally heterogeneous, which is shown by lenses of undeformed rocks between zones of variable strain intensity.

The highest grade of metamorphism was reached in the southern and eastern parts of the area; in the west some greenschist facies rocks have very well-preserved primary structures. A network of greenschist - to amphibolite facies shear zones transects the region. The NNE-SSW shear zone between Karesuando and Malmberget, which in part follows the boundary between Archean and Proterozoic rocks, is one of the most prominent shear zones in Sweden Shear sense indicators show Proterozoic rocks, is one of the most prominent shear zones in Sweden. Shear sense indicators show dominantly steep western-side-up movements. The same shear sense is recorded in several shear zones in the western part of region. Some shear zones in the southeast have moved with the eastern-side-up. At least some of these shear zones are younger than the 1.8 Ga old Lina granite.

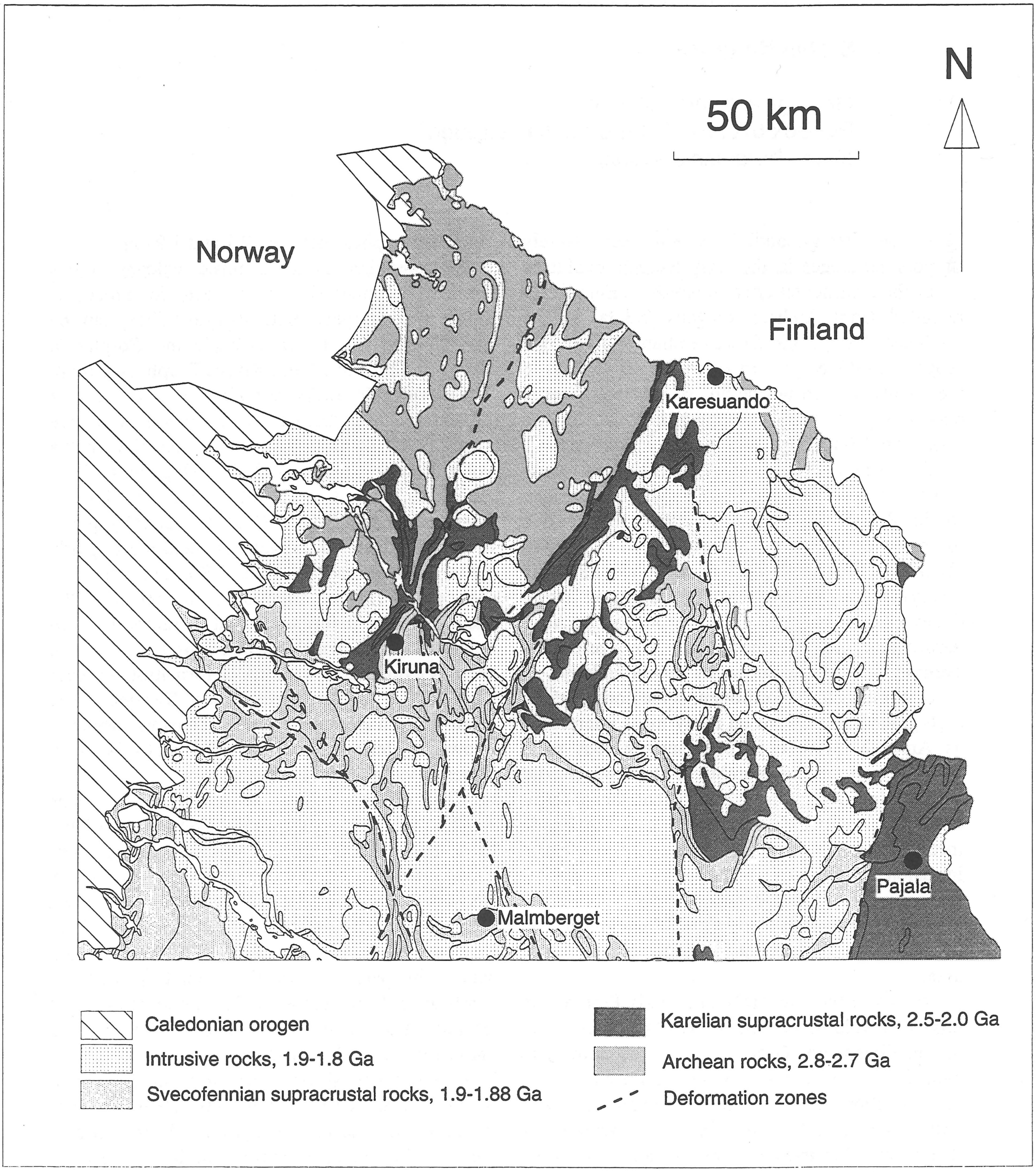

Figure 1. Geological map of northermost Sweden, modified from: Bedrock map of Sweden, Geology, National Atlas of Sweden, 1994.

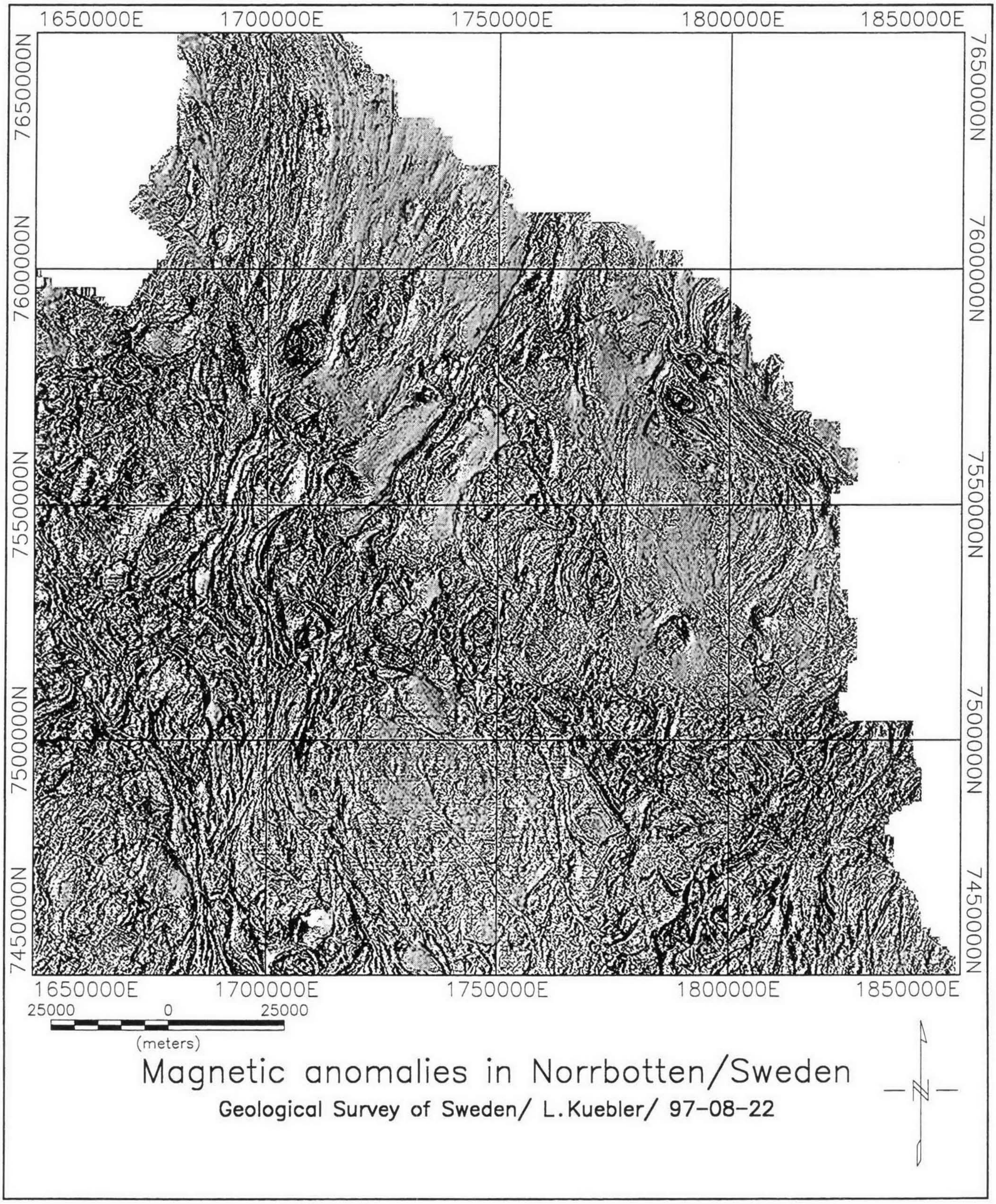

Figure 2. Shadowed total field magnetic anomaly map of northernmost Sweden. Orientation of illumination is 105°, inclination is 40°.

Stratigraphic and Tectonic Framework of the Palaeoproterozoic Skellefte District, Northern Sweden

Jeanette Bergman Weihed

Insititute of Earth Sciences
Uppsala University, Norbyv. 18B, S-752 36 Uppsala, Sweden
weihed@algonet.se

The Skellefte district forms part of the Svecofennian c. 1.90-1.85 Ga supracrustal sequence and associated intrusive rocks in the northern part of Sweden. The district itself is composed of subaqueous volcanic rocks (Skellefte Group) which are overlain by and partly intercalated with a sequence of younger shallow water to subaerial sedimentary and volcanic rocks (Vargfors Group). These rocks are bordered to the south and east by a vast area of highly metamorphosed greywackes (Bothnian Group) and to the north by subaerial volcanic rocks (Arvidsjaur Group) which are similar in age to the rocks of the Vargfors Group. The supracrustal sequence is intruded by 1.95–1.85 Ga calc-alkaline I-type granitoids (Jörn type), by S-type anatectic granites at c. 1.82–1.80 Ga (Skellefte granites), and by younger post-volcanic A- to I-type granitoids at c. 1.80–1.78 Ga (Revsund granitoids).

At least two major phases of folding and one

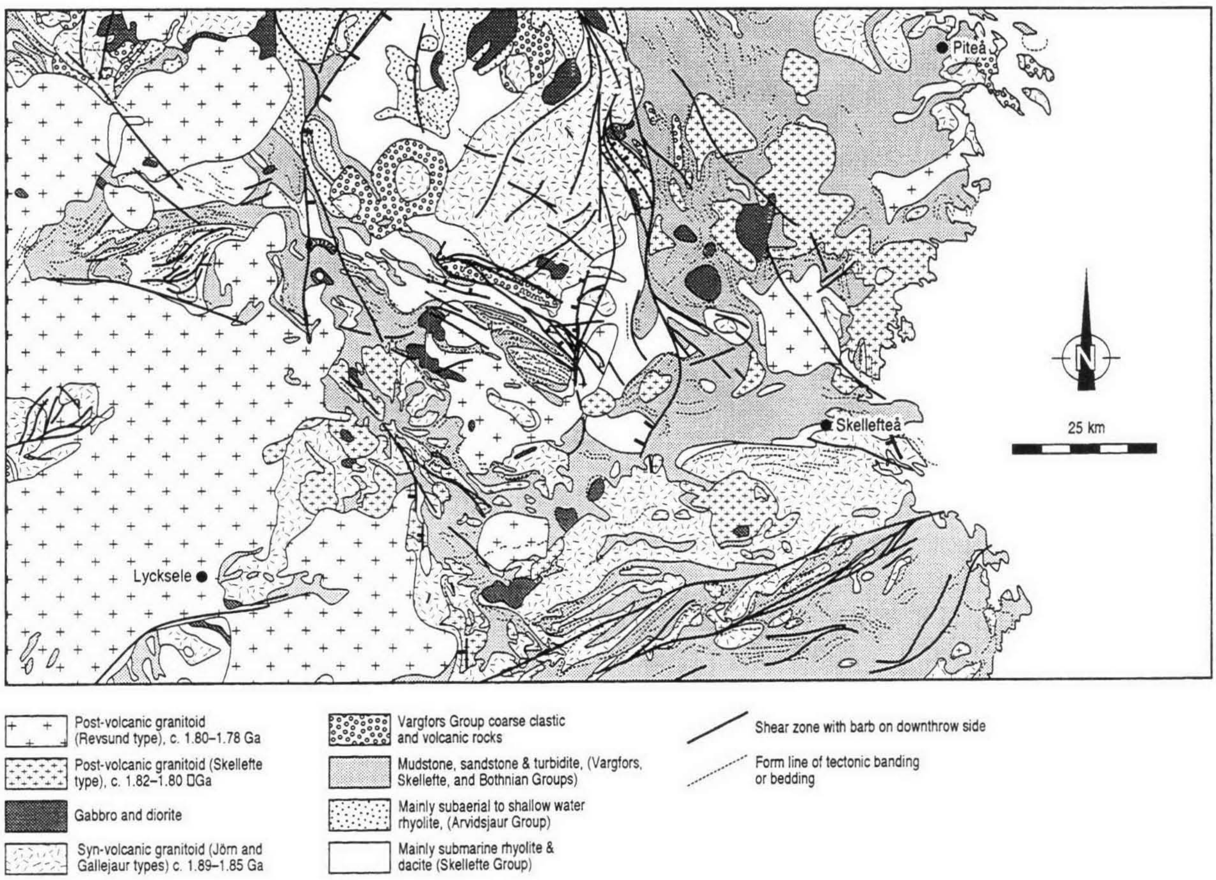

Figure 1. Generalised geological map of the Skellefte district and surrounding areas with form lines of bedding and/or tectonic banding and major shear zones indicated.

phase of metamorphism have affected the supracrustal rocks. First major folds are upright and tight to isoclinal. Axial surfaces strike NE in the western and eastern parts of the district and NW in the central part of the district where numerous shear zones parallel the axial surfaces. These structures are interpreted to represent the last stage of NW-SE convergence during subduction and formed between 1.85 and 1.80 Ga. A locally observed earlier cleavage may have formed during recumbent folding. A later deformation, which involved E-W shortening, produced open folds with steep N-NE striking axial surfaces and related shearing along N-striking reverse-slip zones sometime after 1.80 Ga.

Stratigraphy

The oldest and lowermost rocks in the Skellefte district (Fig. 1, the Skellefte Group) are dominated by syn-eruptive subaqueous mass flows, syn-eruptive sills, cryptodomes, and lavas with associated hyaloclastites (Allen et al. 1996) but facies associations vary between different areas of the Skellefte district. Most of the volcanic units are submarine, but shallow marine to subaerial areas have also been identified (op. cit.). The composition of the volcanic rocks is mainly felsic with more than 50% rhyolite and dacite but a continuous range from rhyolitic to basaltic rocks occurs and the variation in composition is large between different areas of the district. In the upper parts, the volcanic rocks are intercalated with sedimentary rocks which range in composition from mudstone to breccia conglomerate. Immediately below these sedimentary intercalations the majority of VHMS deposits are found. The total thickness of the Skellefte Group is more than 3 km. Age determinations on the volcanic rocks of the Skellefte Group give ages between 1889 and 1882 Ma (Welin 1987, Billström and Weihed 1996).

Rocks of the Skellefte Group are overlain by fine- to coarse-grained sedimentary units and minor volcanic rocks of the Vargfors Group which are mostly conformable to and interfingering with the underlying Skellefte Group. However, some disconformable contacts have been observed both to the Skellefte Group and within the Vargfors Group (Allen et al. 1996). The deposition of rocks of the Vargfors Group generally marks a period of extensive uplift in the area as shown by the shallow water sedimantary rocks containing erosional products of syn-volcanic granitoids as well as the welded ignimbrites. One such welded ignimbrite has been dated at 1875±4 Ma (Billström & Weihed 1996) which is c. 5 million years younger than the Skellefte Group. The total thickness of the Vargfors Group in the Nicknoret area exceeds 1 km.

Areas to the south and east of the Skellefte district (Fig. 1) are dominated by highly metamorphosed turbiditic mud- and sandstones, intruded by several generations of granitoids. This vast area is generally referred to as the Bothnian Basin. The sedimentary rocks are intercalated with mafic volcanic rocks with minor felsic eruptive centers and these rocks may be equivalents of the Vargfors Group rocks. The succession is probably more than 10 km thick (Lundqvist et al. 1990) and spans in age from >1.95 Ga to c. 1.85 Ga. The older, >1.95 Ga, Knaften area, which consists partly of mafic and felsic volcanic rocks (Wasström 1993), is also located in this general region. Its stratigraphic relationship to the surrounding supracrustal rocks is still not clear Volcanic rocks of the Arvidsjaur Group crop out to the north of the Skellefte district (Fig. 1). They are dominated by porphyritic volcanic lavas, subvolcanic intrusions, and tuff which are interpreted to be coeval with the Vargfors Group (cf. Billström and Weihed 1996). The common occurrence of welded ignimbrites and accretionary lapilli indicates that the depositional environment of many of these volcanic rocks was shallow water or subaerial.

No basement has been identified to the rocks in the Skellefte district. The very primitive εNd values of the rocks show that they do not contain any Archaean crustal material. However, a slightly older basement of c. 2.0-1.9 Ga may be possible. Allen et al. (1996) argued on the basis of the overwhelmingly felsic character of the volcanism that the Skellefte Group must have been emplaced on a rifted continental basement during arc extension. Slightly older volcanic rocks and associated intrusions have been described from the Knaften area, about 100 km south of the Skellefte district (Wasström 1993, 1996) and these may be possible remnants of a basement.

To the north of the central part of the Skellefte district a zoned batholith of granite to gabbro is present. The batholith (the Jörn Granitoid Complex, Wilson et al. 1987) is composed of four intrusive phases with ages from 1.89 to 1.87 Ga (Wilson et al. 1987). The batholith thus overlaps in age with the volcanic rocks of the Skellefte Group and both volcanic rocks and the batholith have a calc-alkaline character. It has therefore been proposed that the volcanic rocks of the Skellefte

Group are comagmatic with the Jörn Granitoid Complex. Age determinations on deformed Jörn-type granitoids to the south indicates that calc-alkaline magmatism continued to about 1.85 Ga (cf. Billström & Weihed 1996).

The Gallejaur intrusion situated immediately west of the Jörn Granitoid Complex ranges from gabbro to monzonite in composition and it is related to the mafic volcanic rocks of the Vargfors Group. Age determinations on both the Gallejaur intrusion and the Vargfors Group are identical within error limits c. 1875 Ma. (Weihed & Mäki 1997).

In the high-grade gneisses and migmatites south and east of the Skellefte district, minimum melt granitoids commonly referred to as Skellefte granites are present. These granites have ages around 1.80 Ga (cf. Billström and Weihed 1996). The coarse feldspar-porphyritic Revsund granitoids have been dated at 1.80-1.78 Ga (cf. Billström and Weihed 1996) and, thus, partly overlap in age with the Skellefte granites. The Revsund granitoids were probably derived from a deeper source and crystallised at a shallow level whereas the minimum melt Skellefte granites were more static in the crust (cf. Weihed & Mäki 1997).

Structural Evolution

Only a few structural studies exist from the Skellefte district and even less has been done on the metamorphic history. In general, the supracrustal rocks of the Skellefte district have been affected by at least two major phases of folding and were metamorphosed during the Svecokarelian orogeny at c. 1.85 to 1.80 Ga. The rocks within the Skellefte district have been subject to regional metamorphism in greenschist to lower amphibolite facies but the metamorphic grade increases towards the south and east. Plate tectonic interpretations of the area generally agree on the presence of a former subduction zone dipping north beneath the Skellefte district, and this is supported by a magnetotelluric survey (Rasmusen et al. 1987) which found a low-resistivity slab dipping north under the Skellefte district and by a seismic reflection profile in the Bothnian Bay (BABEL group 1990) which reports a north-dipping reflector east of the Skellefte district.

The early major folds are tight to isoclinal, with upright axial surfaces and variably plunging fold axes (cf. Weihed et al. 1992). Axial surfaces strike north-east in the eastern and western parts of the Skellefte district and west-northwest in the central part of the district (Fig. 1). In the strongly altered volcanic rocks and the sedimentary units in the western Skellefte district, fold axes plunge 45° towards SW (Edelman 1963) and a crenulation cleavage developed along the axial surfaces. In the central Skellefte district, fold axes are variable. Most fold axes plunge 45° SE, but a range of plunges between SE and SW has been observed. An axial planar foliation developed as a penetrative grain shape cleavage in coarser volcanic rocks and as a crenulation cleavage in laminated tuffs and schists. Penecontemporaneous slip along the cleavage is common in this area and shear zones occur in the same general orientation as the axial surfaces. Most of these shear zones have a reverse oblique-slip movement where the southern side has moved upwards over the northern side and they have clearly formed before the metamorphic peak (Bergman Weihed 1997). In the vicinity of Boliden in the eastern Skellefte district fold axes and lineations plunge 75° E with a related axial planar grain shape cleavage, and similar observations were made in the Boliden mine (cf. Bergman Weihed et al. 1996).

In the central Skellefte district, most early folds occur in lower-strain lenses between the northwest-striking shear zones whereas the areas to the east and west are less intensely deformed and have less shearing parallel to the northeast-striking axial surfaces of the folds. It is proposed that these first regional folds in the whole Skellefte district and the northwest-striking shear zones in the central district formed in response to oblique convergence from the south-east and that the northwest-striking structures in the central Skellefte district were constrained in orientation by the presence of the large Jörn batholith which acted as a large block resisting deformation. This deformation occurred between 1.85 Ga and 1.80 Ga before or during peak metamorphism and may represent the last stage of convergence during subduction.

The presence of a crenulation cleavage as an axial planar cleavage to the first major observed folds indicates that an earlier foliation is present. Such a foliation is reported by Allen et al. (1996) who interpret it as a bedding parallel foliation. Folds related to this early foliation have been observed in thin sections from the sedimentary units west of Kristineberg (Bergman Weihed 1997) and similar observations were also made in mudstones in the central part of the Skellefte district (Bergman Weihed unpublished data). The early foliation may have formed during an early recumbent phase of folding and this has been suggested for the Långdal area by Talbot (1988) and Assefa (1990).

The second major phase of folding produced open folds with steep north- to northeast-striking axial surfaces and fold axes coaxial with the early folds. A spaced axial planar S2 crenulation cleavage developed locally and, in the central part of the Skellefte district, the intensity of the second folding increases towards the south with larger amplitude folds and a more penetrative cleavage. A number of north-striking shear zones are present in the Skellefte district and surroundings (Fig. 1). These have a dominating reverse dip-slip movement and they are interpreted to have formed during or after the second major phase of folding and after peak metamorphism. The dominantly dip slip reverse movements observed on these shear zones indicate an east-west shortening during this period of deformation. Since these shear zones deform all rocks in the area, they must have formed after c. 1.80 Ga which is the age of the youngest intrusive phase.

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Palaeoproterozoic of the Kimberley to Granites-Tanami Region, Northwestern Australia

D.H.Blake

Australian Geological Survey Organisation,
GPO Box 378, Canberra, ACT 2601, Australia
(dblake@agso.gov.au)

The Kimberley to Granites-Tanami region (Fig. 1) has a Palaeoproterozoic history of mafic, felsic, and alkaline volcanism; deep water turbiditic and shallow water clastic, calcareous, and carbonaceous sedimentation; felsic and mafic to ultramafic plutonism; several phases of deformation and greenschist to granulite facies metamorphism (Table 1); and significant Au, Cu, Pb, Zn, platinum group element (PGE) and rare-earth element (REE) and associated rare metals mineralisation. The main Palaeoproterozoic tectonic units in the region are the King Leopold and Halls Creek orogens, Speewah and Kimberley basins, and The Granites-Tanami block. These are overlapped by Mesoproterozoic to Palaeozoic sedimentary rocks of the Birrindudu, Bonaparte, Canning, Ord, Victoria River, Wiso, and Wolfe basins. The oldest rocks known in the region are late Archaean felsic gneisses in the northwest and far east of The Granites-Tanami block, which is separated by younger sedimentary basins from the Halls Creek Orogen to the northwest and the Palaeoproterozoic Tennant Creek and Pine Creek provinces to the east and north, respectively.

Regional Geology

The geology of the Kimberley to The Granites-Tanami region was first mapped systematically during the 1960s and early 1970s by the Bureau of Mineral Resources (now Australian Geological Survey Organisation, AGSO) and the Geological Survey of Western Australia (GSWA). Parts have been remapped in greater detail by AGSO (since 1990) and GSWA (since 1986). The more recent work has included interpretations of airborne geophysics, gravity, Landsat imagery, U-Pb zircon dating (by R.W.Page, AGSO), rock and mineral chemical analyses, and isotope studies.

The east-southeast trending King Leopold Orogen and north-northeast trending Halls Creek Orogen flank the Kimberley Craton, which consists of concealed Archaean and/or early Palaeoproterozoic rocks covered by gently dipping cover rocks of the 1835 Ma Speewah Group and 1800 Ma Kimberley Group, laid down in the Speewah and Kimberley basins, and sills of 1790 Ma dolerite. The King Leopold Orogen is formed of folded and metamorphosed turbiditic sedimentary rocks containing detrital zircon as young as 1870 Ma (Marboo Formation), metadolerite, and less deformed 1865-1850 Ma felsic volcanics and granitic plutons. Contacts with unconformably overlying sandstone of the Speewah Group to the north are commonly sheared.

The Halls Creek Orogen to the east has been subdivided by Tyler & others (1995) into western, central and eastern zones, separated by major sinistral strike-slipe faults trending north-northeast. The western zone correlates directly with the King Leopold Orogen, being dominated by metamorphic rocks mapped as Marboo Formation and 1860-1850 Ma felsic volcanics and granites, but it also contains layered mafic-ultramafic intrusions emplaced at around 1855 Ma and some younger granite. The felsic volcanics are overlain disconformably in the west by the Speewah and Kimberley groups. The central zone includes high-grade 1865-1843 Ma Tickalara Metamorphics, most of which are probably metaturbidites, 1855 Ma Panton and 1845 Ma Sally Malay mafic-ultramafic intrusions, low to high-grade metamorphics of the 1843 Ma Koongie Park Formation, voluminous 1830-1810 Ma granites and mafic intrusions, and unconformably overlying Kimberley Group. No correlatives of the pre-1850 Ma metamorphic and igneous rocks exposed in the western zone have been identified in this zone. The Koongie Park Formation hosts subeconomic VMS deposits, and the mafic-ultramafic intrusions in both western and central zones are prospective for PGE, Cr, Ni, Cu, Co and Au.

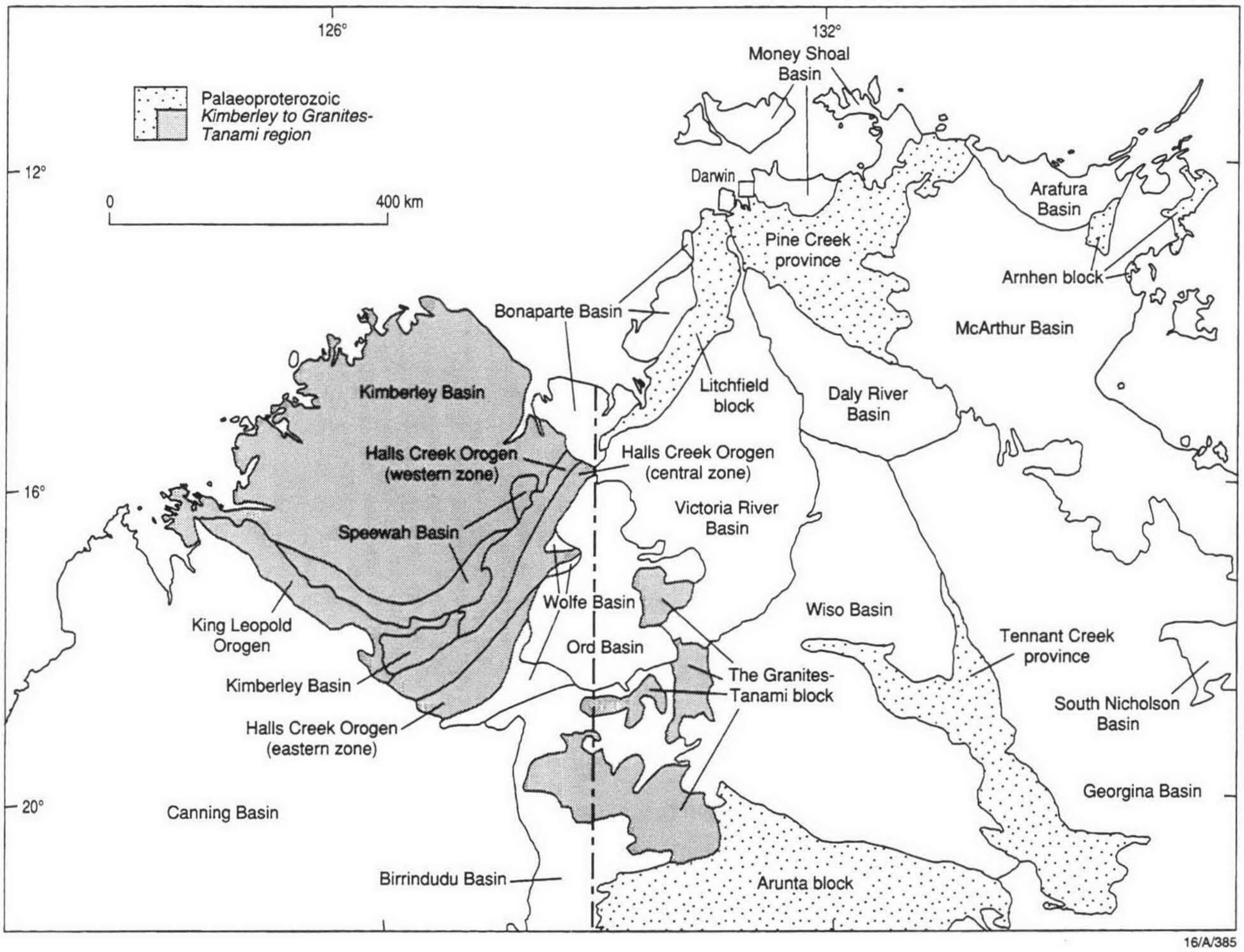

Figure 1. Palaeoproterozoic provinces of the western and central parts of northern Australia

The eastern zone of the Halls Creek Orogen is dominated by the 1880-1845 Ma Halls Creek Group, comprising the Saunders Creek, Biscay and Olympio formations, but it also contains the 1910 Ma bimodal Ding Dong Downs Volcanics, the oldest unit known in the orogen, and, in the far south, 1820-1790 Ma granite. The turbiditic Olympio Formation, which also contains alkaline volcanics dated at 1857 Ma and 1848 Ma, may be 20 Ma younger than the underlying Biscay Formation, a unit of mainly basaltic and calcareous rocks, but including 1880 Ma felsic volcanics. Gold-bearing quartz veins occur in both formations, sub-economic rare earth and associated rare metals mineralisation is hosted by alkaline volcanics in the Olympio Formation, and minor Cu-Pb-Zn deposits occur in the Biscay Formation. The turbidites of the Olympio Formation are probably more than 10 Ma younger than those of the Marboo Formation of the western zone, but may be similar in age to some in the Tickalara Metamorphics of the central zone. The Biscay Formation appears to be older than any rocks exposed in the other two zones. There are no correlatives in the eastern zone of the Koongie Park Formation.

Major folding and regional metamorphism in the central and eastern zones of the Halls Creek Orogen affected the Koongie Park Formation and Halls Creek Group at around 1835 Ma, before the emplacement of 1830-1810 Ma granites and mafic intrusions. This was the main deformation event of the Halls Creek Orogeny. No comparable deformation of this age has been recognised in either the western zone of the orogen or the King Leopold Orogen.

Coeval granitic to tonalitic and mafic to ultramafic plutonism took place repeatedly between 1860 and 1810 Ma in the Halls Creek Orogen. This led to the formation of numerous net-veined complexes, formed by commingling of felsic and mafic magmas, and possibly also, at least according to Blake (in Blake & Hoatson, 1993), to most or all of the high-grade metamorphism in the orogen. Emplacement depths for the mafic-ultramafic intrusions (Trudu & Hoatson, 1996), and hence also for adjacent coeval granites, ranged from about 8 km to 23 km.

Table 1. Palaeoproterozoic events in the Kimberley to Granites-Tanami region

Age (Ma)	*King Leopold Orogen*	*Kimberley Craton*	*Halls Creek Orogen*			*Granites-Tanami block*
			western zone	*central zone*	*eastern zone*	
~1790		Dolerite sills	Dolerite sills		Granite	
~1795						Granite
~1800	Clastic sediments, basalt	Clastic sediments, basalt	Clastic sediments, basalt	Clastic sediments, basalt	Clastic sediments, basalt	
1800-1830			Granite	Granite, gabbro	Granite	Clastic sediments, felsic volcanics
~1835	Clastic sediments	Clastic sediments	Clastic sediments	*Regional deformation;* ?clastic sediments	*Regional deformation*	*?Regional deformation*
~1845				Bimodal volcanics, calcareous sediments, turbidites, mafic-ultramafic intrusions	?Turbidites	Turbidites
~1850	*Regional deformation*		*?Regional deformation*	?Bimodal volcanics, turbidites	Turbidites, alkaline volcanics	?Turbidites
1855-1865	Felsic volcanics, granites		Felsic volcanics, granite, mafic intrusions	Mafic-ultramafic intrusions	Turbidites, alkaline volcanics	?Turbidites
~1865	*Regional deformation*		*?Regional deformation*	*?Regional deformation*	?Erosion	?Erosion
1865-1880	Turbidites		Turbidites	?Turbidites	Bimodal volcanics, clastic and calcareous sediments	?Clastic and calcareous sediments
~1910					Bimodal volcanics	

Two main models have been proposed for the development of the Halls Creek Orogen prior to 1800 Ma. In one model an ensialic setting is envisaged, with extension and crustal thinning, but no generation of oceanic crust, being followed by convergence with limited subduction (e.g., Hancock & Rutland, 1984). In an alternative model, invoking Phanerozoic-style plate tectonics, Tyler & others (1995) suggest that the orogen represents convergence and collision of three separate terranes, previously widely separated, and subduction of oceanic crust between them, in the period 1850-1800 Ma: one terrane includes the western zone, and also the King Leopold Orogen and Kimberley Craton, one is represented by the central zone, and the other includes the eastern zone and probably both the Pine Creek province and The Granites-Tanami block. In this model the faults separating the western, central and eastern zones of the Halls Creek Orogen are Palaeoproterozoic terrane boundaries which have been reactivated as major strike-slip faults.

The Granites-Tanami block to the southeast (Fig 1; Table 1) is a poorly exposed area of mainly Palaeoproterozoic rocks overlain by Mesoproterozoic and younger sedimentary rocks. The older Palaeoproterozoic rocks, mainly metasediments, belong to the Tanami Complex of Blake & others (1979). They include folded greenschist facies turbidites (Killi Killi beds) similar in age to those of the Olympio Formation and an apparently underlying sequence of partly calcareous metasediments (Mount Charles beds) which may correlate with the Biscay Formation. The Mount Charles beds host major stratabound gold deposits. The main folding and metamorphic event affecting the Killi Killi and Mount Charles beds predates sandstone deposition and felsic volcanism that took place between 1825 Ma and 1810 Ma. This deformation may correlate with either the 1835 Ma deformation of the Halls Creek Orogeny to the west or the main deformation, at about 1855 Ma, of the turbiditic Warramunga Group in the Tennant Creek province to the east. Younger granite plutons include The Granites Granite, dated at 1795 Ma. The 1825-1810 Ma sandstone and felsic volcanics in The Granites-Tanami block are similar in age to sandstone and bimodal volcanics in the Hatches Creek Group of the Tennant Creek province, and apparently slightly older than sandstone and basalt of the Kimberley Group to the northwest.

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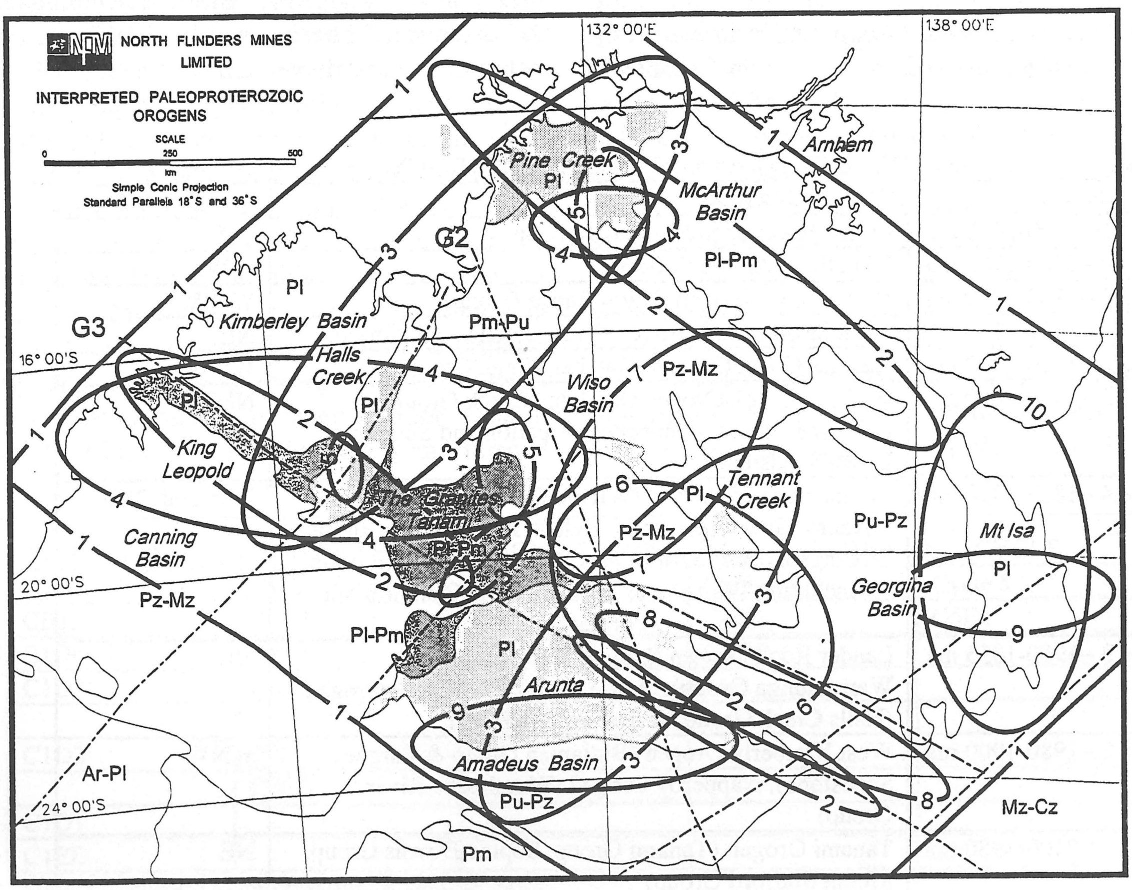
Palaeoproterozoic Geological Events and Gold Mineralisation in the Halls Creek - Granites - Tanami Orogenic Domain, Northern Australia

Puquan Ding

North Flinders Mines Limited
28 Greenhill Road, Wayville, SA 5034
Australia

Intense exploration and geological research has been carried out by NFM geologists in The Granites - Tanami region during the last decade. The general geological research and regional correlation have been further extended to other areas in northern Australia, including the Tennant Creek, Pine Creek, Halls Creek, Mount Isa and Arunta regions at various levels. A total of ten sequential Palaeoproterozoic orogenic events represented by angular unconformities have been identified in northern Australia. Each unconformity separates a new cycle of stratigraphic sequence from the previously deformed basement (Table 1). The deformations associated with each of these orogenies have been temporarily labeled as D1 to D10 in a younging order. Different orogenies are of variable intensities, and even the intensity of one orogeny varies from place to place. The palaeobasins associated with these orogenic cycles can be different in size, location and orientation (Figure 1).



The Halls Creek - Granites - Tanami Orogenic Domain

The Halls Creek - Granites - Tanami Orogenic Domain (H-G-T Domain), recorded only the first five of the ten orogenies, i.e. D1 to D5, and the post-D5 strata are platform cover sequences which are not deformed or very gently deformed. The D6 to D10 orogenies, which are not included in this paper, occurred in the Tennant Creek, Arunta and Mount Isa Domains further to the east. Each orogeny was followed by uplifting and erosion, which often exposed the metamorphosed roots of the orogen to the surface, forming the base of the subsequent basin. The rocks below each unconformity would have subsided again to a deep position during a later cycle while the sediments were accumulated within the basin above the unconformity. During the next orogeny, the rocks below the unconformity would have recorded one more deformational event than that above the unconformity. The Tanami Group recorded ductile deformations of D1 to D5, while the cycle 5 strata recorded only one ductile deformation. The correlation of deformations recorded in each tectonic cycle is shown in Table 2.

The Frankenia Orogeny (D1)

The cycle 1 Tanami Group, which is named by NFM to replace the old term "Tanami Complex", is a typical metamorphosed sedimentary sequence which was deposited unconformably on Late Archean gneissic basement (~ 2500 ma). The Tanami Group was intensely deformed by D1 in the region and initially formed a NE trending fold belt. A zircon age of ~1980 Ma given by a syn-orogenic granitoid is used to represent this orogeny, which is informally named the Frankenia Orogeny. The Sophie Downs Group in the Halls Creek is newly defined and recognised as cycle 1 by the author. It is an amphibolite facies supracrustal sequence which had been intensely folded by two deformational events before the intrusion of 1910 Ma Sophie Downs Granite. The Sophie Downs Group extended from the type area around the Sophie Downs Granite into the Black Rock Anticline area to the north, where this Group was mapped as Tickalara Metamorphics. A correlation between the Halls Creek and The Granites - Tanami regions are shown in Table 3. The Tanami Group may have accumulated 12-15 km thick sediments in a palaeobasin during a period of about 450 ma. A complex metamorphic and dynamic process occurred before the orogeny. This pre-orogenic tectonism was considered unusual when it was recognised for the first time in 1990, but later similar features have been found in other orogenic cycles and I believe that any thick pile (> 12 km) of sediments accumulated in an extensional basin would have a similar metamorphic and dynamic history.

Table 1. TYPE OROGENS AND STRATA OF EACH TECTONIC CYCLE

CYCLE	TYPE OROGEN	ORIENTATION
cycle 10 1700-1600 ma	Mount Isa Orogen (Mount Isa Group)	N-S
cycle 9 ~1740-1700 ma	? Southern Arunta Orogen	E-W
cycle 8 ~1760-1740 ma	Harts Range Orogen (Harts Range Group)	WNW
cycle 7 ~1780-1760 ma	Tomkinson Creek Orogen (Tomkinson Creek Group)	NE
cycle 6 ~1810-1780 ma	Hatches Creek Orogen (Hatches Creek Group)	WNW
cycle 5 ~1830-1810 ma	Birthday Creek Orogen (Birthday Creek Group: including Mount Winnecke Formation and Supplejack Down Sandstone)	NNW
cycle 4 ~1855-1830 ma	Wilson Creek Orogen (Wilson Creek Group, including Nanny Goat Creek beds, Helena Beds and part of the Nongala beds); Ord River Orogen (Ord River Group including Whitewater Volcanics and Panton River Formation)	E-W and ENE
cycle 3 ~1900-1855 ma	Lander Rock Orogen (Lander Rock beds and Warramunga Group); Halls Creek - Pine Creek Orogen (Halls Creek Group) & Tanami Mine Rift.	NE
cycle 2 ~1980-1900 ma	West Kimberly Orogen (Bullaman Group & Pargee Sandstone); Napperby Orogen (Woodforde River Group)	WNW
cycle 1 >2100-1980 ma	Tanami Orogen (Tanami Group, Sophie Downs Group, Mount Stafford Group)	NE

The Pre-D1 History Of The Tanami Group

The lower part of the Tanami Group had been metamorphosed to lower amphibolite or upper greenschist facies with the initiation of bedding parallel schistosity (S0). A simultaneous subhorizontal shearing occurred within the lower part of the Tanami Group. Bedding parallel or subhorizontal pervasive low-strain bulk stretching was enhanced by bedding parallel discrete high-strain shear zones. A pervasive bedding parallel or extremely low angle schistosity (ES1) and mineral lineation (EL1) were thus formed. In some of the discrete shear zones, the bedding parallel compositional bands have been further deformed by intermittent shearing and formed recumbent folds (EF1a) with the development of axial plane schistosity and shear bands (ES1a). The earliest hydrothermal fluid entered the shear zones and caused metasomatism, adding Si and Fe into the system, forming Fe-rich schists and the earliest cherty bands and quartz veins. The ES1a and its associated banded structures have been further deformed and formed the second phase of recumbent folds (EF1b). Some of the earlier formed cherty bands and quartz veins were pulled apart, boudinaged and often formed elongated chert or quartz augens in schists. New shear bands were further developed along the axial plane schistosity (ES1b) and transposed limbs of EF1b folds. Sulphides and/or chlorite-rich alteration bands and chlorite - quartz veins occurred along ES1b through a second phase of metasomatism. Later the earliest acid magma intruded into the bedding parallel shear zones and formed a thick sill of granitoid which gives a crystallization age of ~2450 Ma. Pervasive, although less intensive bedding parallel foliation (ES1c) and elongation lineation (EL1c) were developed within the granite sill by subhorizontal shearing. Finally a large quantity of basic magma intruded into the Tanami Group at different stratigraphic levels and formed several thick and numerous thin concordant dolerite sills. Further shearing caused bedding parallel discrete shear zones within thick dolerite sills, particularly along their margins, but caused pervasive schistosity (bedding parallel ES1d and EL1d) within the thin sills which were further boudinaged in places. The bedding parallel fabrics, recumbent folds, foliated granite and dolerite sills were all folded during the Frankenia Orogeny and formed regional upright folds (F1) with axial plane schistosity (S1). Similar features of pre-orogenic deformation have also been observed in the Halls Creek region. The orientation of the stretching lineation has been used to restore the extensional direction of lower crust during the basin development, and it is found that the compressional direction of each orogeny parallels the early extensional direction which is always perpendicular or at high angle to the axis of fold belt. In other words the compression was the reverse movement of the extension.

Ding Dong Downs Orogeny (D2)

The Tanami Group is overlain unconformably by the cycle 2 Pargee Sandstone which has been interpreted as the basal sandstone of cycle 2 stratigraphic group. The F1 folds in the Tanami Group have been refolded during D2 and formed NW trending F2 folds which are, in fact, the time equivalent of the F1 folds affected the Pargee Sandstone. The cycle 2 Bullaman Group in the Halls Creek region has been well dated and shows strong evidence of unconformity, although indirect, between the Bullaman Group and the Sophie Downs Group.

Halls Creek Orogeny (D3)

The Ding Dong Downs Volcanics of the Bullaman Group is overlain unconformably by the Saunders Creek Formation of the cycle 3 Halls

Table 2. CORRELATION OF DEFORMATIONS RECORDED IN EACH TECTONIC CYCLE

Cycle 1	Cycle 2	Cycle 3	Cycle 4	Cycle 5
CID5	C2D4	C3D3	C4D2	C5D1
C1D4	C2D3	C3D2	C4D1	
C1D3	C2D2	C3D1		
		C3ED		
C1D2	C2D1			
	C2ED			
C1D1				
C1ED				

Note: C - cycle; D - deformation; E -pre-orogenic

Creek Group. The Halls Creek Orogeny, produced NNE trending F1 folds in the Halls Creek Group, F2 folds in the Bullaman Group and F3 in the Sophie Downs Group. The Tanami Mine Succession has been correlated to the Halls Creek Group on the basis of structural evolution. The F1 folds in the Tanami Mine Succession trend NE and parallel the F3 folds within the Tanami Group, but parallel the F2 folds within the Pargee Sandstone.

Ord River Orogeny (D4)

The angular unconformity between the Whitewater Volcanics and the Halls Creek Group has long been asserted. However the angular unconformity between the newly named Panton River Formation and the Halls Creek Group is recently recognised by the author. So a new name, the Ord River Group, is created to include both the Whitewater Volcanics and the Panton River Formation into the cycle 4 stratigraphy. The Ord River Orogeny produced ENE trending F1 folds in Ord River Group and in the Wilson Creek Group (in The Granites - Tanami region), but F4 folds in the Tanami Group, or F3 folds in the Pargee Sandstone and F2 folds in the Halls Creek Group.

Winnecke Range Orogeny (D5)

Angular unconformity has also been recognised between the cycle 5 Supplejack Downs Sandstone and the cycle 4 Nannygoat Creek beds of the Wilson Creek Group in The Granites - Tanami region, and between the cycle 5 Caroline Pool Group and its basement in the Halls Creek region. A new name, the Birthday Creek Group, is created to include both the basal Supplejack Downs Sandstone and the upper Mount Winnecke Formation. Cycle 5 orogeny caused northerly trending F1 folds in the Birthday Creek Group and the Caroline Pool Group, but F5 in the Tanami Group.

Gold Mineralisation

Gold mineralisation has been found in the cycle 1 Tanami Group, the cycle 3 Halls Creek Group and the Tanami Mine Succession, and the cycle 4 Ord River Group. We believe that gold mineralisation occurred as early as cycle 1, and probably every orogenic cycle made its own contribution, although the later ones are more apparent. Pre-orogenic mineralisation caused strata bound deposits (e.g. The Granites Gold deposit) while syn- or post-orogenic mineralisation was mainly controlled by fracture systems (e.g. Callie and Tanami Gold deposits)

Table 3. . CORRELATION BETWEEN THE HALLS CREEK AND THE GRANITES TANAMI REGIONS

	Halls Creek region	The Granites - Tanami region
	?	Birrindudu Group
Cycle 6 stratigraphic unit	Platform cover Kimberley Group (intruded by ~1800 Ma Hart Dolerite and associated granophyre).	?
	------------------------- Angular unconformity -------------------------	
Orogenic deformation in Tanami region only, caused northerly trending very tight folds	Weak deformation	D5 (~1810 Ma) in the Tanami Group (C1D5) but D1 in the Mount Winnecke Formation (C5D1)
Cycle 5 stratigraphic unit	Platform cover Speewah Group (1834 +/- 3 Ma felsic volcanics in the Valentine Siltstone)	Birthday Creek Group (1815-1830 Ma) (including Mount Winnecke Formation and Supplejack Downs Sandstone)
	------------------------- Angular unconformity -------------------------	
Orogenic deformation caused E-W to ENE trending folds	D4 in the Sophie Downs Group (C1D4) but D1 in the Ord River Group)C4D1) at ~1835 Ma; deformed the 1845+/- 2 Ma, 1843 +/- 5 Ma and 1840 +/- 4 Ma felsic rocks	D4 in the Tanami Group (C1D4) but D1 in the Wilson Creek Group (C4D1); deformed ~1845 Ma tonalite intrusive body
Pre-orogenic bedding parallel ductile shearing deformation	C4ED weak shear on bedding surfaces	
Cycle 4 stratigraphic unit	Ord River Group: including Whitewater Volcanics (~1855 Ma), Panton River Formation (1848 -1835 Ma)	Wilson Creek Group (including parts of the previous Nongala Beds, Helena Beds and Nannygoat Creek beds).
-------------	---- Angular unconformity -----	------------ ?
Orogenic deformation caused NNE trending fold belt	D3 in the Sophie Downs Group (C1D3) but D1 in the Halls Creek Group (C3D1); deformed 1857 Ma, 1870 Ma and 1880 Ma volcanic and subvolcanic rocks within the Halls Creek Group	D3 in the Tanami Group (C1D3) but D1 in the Tanami Mine Succession (C3D1); deformed 1870 Ma and ~1860 Ma tonalite dykes which intruded into the S3 in the Tanami Group
Pre-orogenic bedding parallel ductile shearing deformation	C3ED caused bedding parallel schistosity and some stretching lineation on bedding surface within the Halls Creek Group	No C3 ED
Cycle 3 stratigraphic unit	Halls Creek Group (including Saunders Creek Formation, Biscay Formation and Olympio Formation)	?Tanami Mine succession (Lander Rock beds)
---------------	----- Angular unconformity ------	---------- ?
Orogenic deformation caused NW trending fold belt	D2 at ~1900-1910 Ma in the Sophie Downs Group (C1D2) but D1 in the Bullaman Group (C2D1)	D2 in the Tanami Group (C1D2) but D1 in the Pargee Sandstone (C2D1)
Pre-orogenic bedding parallel ductile shearing deformation	C2ED caused bedding parallel schistosity and some stretching lineation on bedding surface within the Ding Dong Downs Volcanics	No C2ED
Cycle 2 stratigraphic unit	Bullaman Group (including ~ 1910 Ma Ding Dong Downs Volcanics at the top)	Pargee Sandstone (the basal unit of the cycle 2 strata);
------	------------ ? ---------------	------ Angular unconformity
Orogenic deformation caused NE trending fold belt	?C1D1	C1D1 at about ?1980 Ma
Pre-orogenic bedding parallel ductile shearing deformation	C1ED	C1ED
Cycle 1 stratigraphic units	Sophie Downs Group (intruded by the ~1910 Ma Sophie Downs Granite)	Tanami Group (intruded by a pre-orogenic granite sills at ~2450 Ma)
Basement		Archaean granitic gneiss (~2500 Ma)

Tennant Creek Province: Stratigraphic and Structural Framework

Nigel Donnellan[1], Robert S. Morrison[2] and Kelvin J. Hussey[1]

[1]NTGS
PO Box 2655
Alice Springs, NT 0871, Australia

[2]WMC Resources Ltd
St. Ives Gold, Kambalda PO
Kambalda, WA 6442, Australia

Palaeoproterozoic rocks of the Tennant Creek Inlier crop out over an area of >45,000 km2 in central Australia. The Inlier includes the Ashburton and Davenport provinces, respectively to the north and south of the Tennant Creek province. The latter is interpreted as the upper, exposed portion of a Palaeoproterozoic greenstone/continental arc. At the present level of exposure, it comprises a turbiditic flysch succession, the Warramunga Formation, which together with granite and granodiorite were respectively deposited and intruded penecontemporaneously with the Barramundi Orogeny (c. 1870±20 Ma; Green 1992). Extrusive, predominantly subaerial volcanic rocks; rhyolitic and rhyodacitic ignimbrite, lava and tuff; and associated volcaniclastic and clastic sedimentary rocks, are also included in the Tennant Creek province although they were not deformed during the Barramundi Orogeny. These volcano-sedimentary rocks are called the Flynn Subgroup (Donnellan et al. 1995). The Flynn Subgroup has a transitional contact with the Hayward Creek Formation, the lowermost unit of the Tomkinson Creek Subgroup cropping out to the north in the Ashburton province. In the south the lowermost subgroup of the Hatches Creek Group, namely the Ooradidgee Subgroup, in the Davenport province has been variously correlated with rocks of the Tennant Creek province and particularly with those of the Flynn Subgroup (Blake 1984; Donnellan et al., 1995). S1 was reactivated c.1810-1790 Ma, and the entire Inlier was deformed contemporaneously with the early Strangways Orogeny (c.1770 Ma) of Collins and Shaw (1995) in the Arunta Inlier.

Stratigraphy and Structure

Warramunga Formation

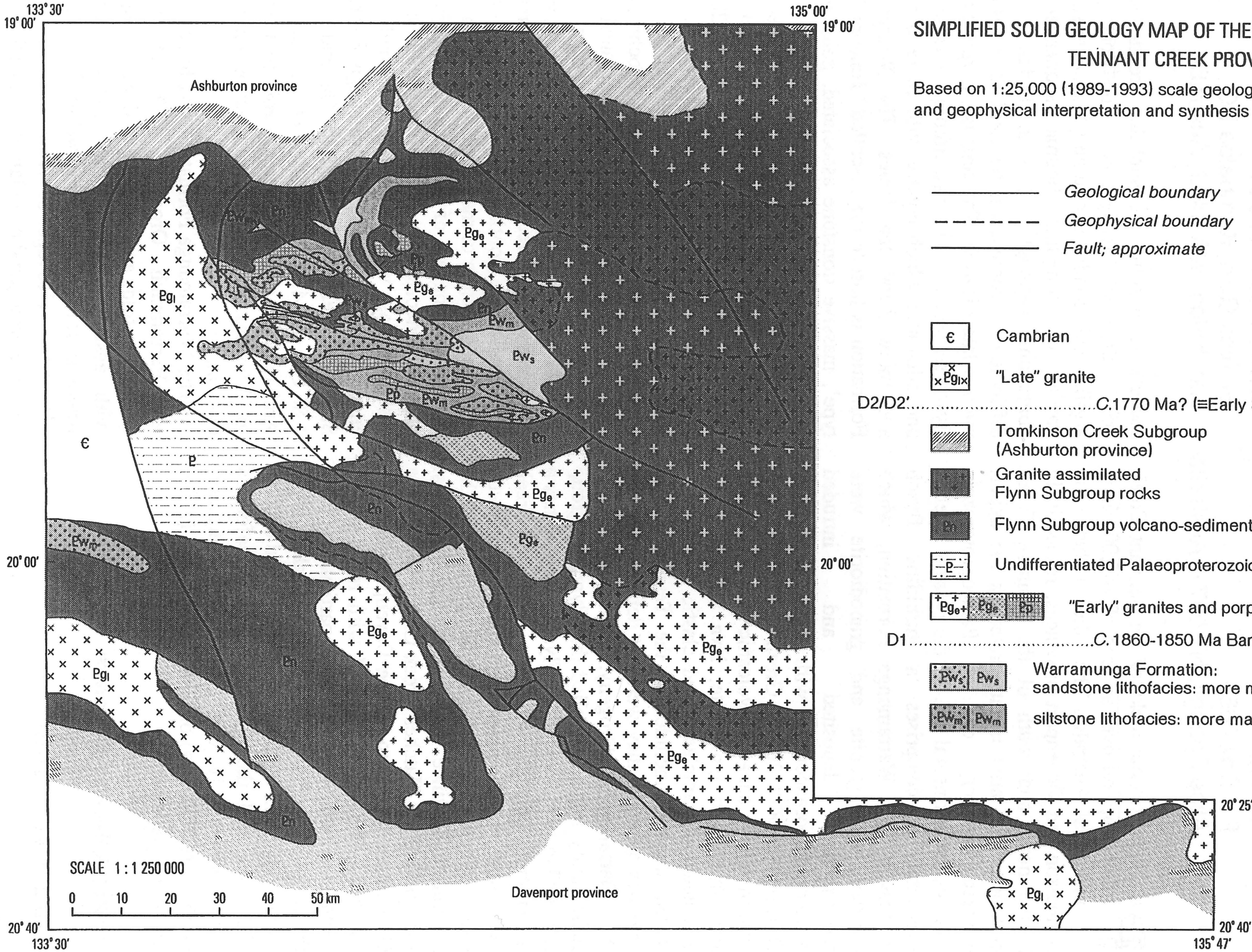
The Warramunga Formation is a polydeformed sub- to lowermost-greenschist facies, turbiditic succession of lithic and sublithic arenite, wacke and siltstone; terrigenous mudstone and argillaceous, banded ironstone ('haematite shale'). Partial or complete bouma sequences are ubiquitous although, as with bedding, often difficult to recognise in the field as a result of intense lateritisation. The unit is interpreted as a flysch succession, with two lithofacies (sandstone and siltstone) which equate with more proximal and more distal fan facies. The Warramunga Formation is host to the so-called 'Tennant Creek type', massive ironstone associated, gold-copper-bismuth mineralisation of the Tennant Creek gold field. Until the end of 1996 the gold field had produced about 140t of gold, together with 280,000t of copper, 14,000t of bismuth, 220t of selenium and 53t of silver (Ferenczi, 1996 pers.comm.).

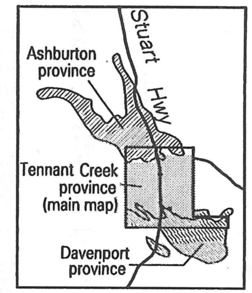
Single crystal zircon U-Pb geochronology (Compston 1995), and geochemical and petrographic considerations (Donnellan 1994) indicate the 1.86 Ga Warramunga Formation was predominantly derived from penecontemporaneous felsic volcanic (continental arc) rocks, together with an admixture of mafic volcanic detritus and a minor component of acid metamorphic and plutonic igneous rocks. Compston (1995) reported detrital zircons with ages between 3.0 and 2.0 Ga, and c.1.93 Ga from the Warramunga Formation. Sm/Nd model ages are ≤2.45 Ga and εNd at 1.86 Ga ≤ -4.9 (Donnellan et al. in prep.). Archaean and early Paleoproterozoic contributions to Warramunga Formation sedimentation are evident.

The Warramunga Formation was folded during the Barramundi Orogeny (D1). These first generation folds are moderate to tight, and upright, with east/west to east-southeast/west-northwest trending fold axes and a well-developed axial planar slaty cleavage. S2 (NW-SE) and S2' (NE-SW) are orthogonal crenulation, or locally fracture or slaty, cleavages. S2 generally antedates S2' although locally the converse is true indicating this

SIMPLIFIED SOLID GEOLOGY MAP OF THE PALAEOPROTEROZOIC TENNANT CREEK PROVINCE

Based on 1:25,000 (1989-1993) scale geological mapping by the NTGS: and geophysical interpretation and synthesis by L.J.D. Farrar (1991-1994).





Geological boundary

Geophysical boundary

Fault; approximate

€ Cambrian

Ꝑg$_l$ "Late" granite

D2/D2'...C.1770 Ma? (≡Early Strangways Orogeny).

Tomkinson Creek Subgroup (Ashburton province)

Hatches Creek Group (Davenport province)

Granite assimilated Flynn Subgroup rocks

Flynn assimilated granite

Ꝑn Flynn Subgroup volcano-sedimentary rocks

Ꝑ Undifferentiated Palaeoproterozoic

Ꝑg$_e$ Ꝑg$_e$ Ꝑp "Early" granites and porphyry

D1...C. 1860-1850 Ma Barramundi Orogeny

Ꝑw$_s$ Ꝑw$_s$ Warramunga Formation: sandstone lithofacies: more magnetic; less magnetic

Ꝑw$_m$ Ꝑw$_m$ siltstone lithofacies: more magnetic; less magnetic

figure: 1

is a genuine, conjugate deformation as previously suspected by Dunnet and Harding (1967). The style of folding produced in D_2/D_2' in the Warramunga Formation is well represented on the mesoscale. Excellent examples of these folds are to be found in the Mary Lane shear zone. Mesoscale anticlinal folds are symmetric and asymmetric chevron folds; asymmetric, box folds and doubly peaking anticlines; symmetric doubly peaking anticlines ('M' folds). Mesoscopic synclinal folds are predominantly concentric. However, there are examples of concentric anticlinal folds; symmetric and asymmetric synclinal chevron folds; and asymmetric, box and doubly peaking synclines. The entire outcrop area of the Warramunga Formation is interpreted to define asymmetric anticlinal box, and concentric synclinal folds, with large-scale chevron folds developed in zones of high-strain (kink bands).

Barramundi Igneous Association

Representatives of the BIA in the Tennant Creek province are loosely referred to as the 'early granites'. These are biotite-bearing, epizonal granites and mesozonal granodiorite with seriate to porphyritic and 'incipient' rapakivi textures, with interstitial mafic minerals, and blue, opalescent ovoid quartz (restite?). A reddish colour, rapakivi texture and an association with quartz-porphyry, rhyolitic ignimbrite and dolerite are all A-type granite characteristics (cf. Windley 1993, and Anderson and Bender 1989). The early granites are, however, syntectonic with respect to the Barramundi Orogeny in the Tennant Creek province; this is evidenced by, for example, an east-west foliation locally in both the Tennant Creek Granite and the Mumbilla Granodiorite.

A characteristic feature of both the Tennant Creek Granite and the Mumbilla Granodiorite is a varied and abundant enclave population. Igneous enclaves include felsic porphyry, cognate, cumulate-rich granitic and gabbroic enclaves and late-stage lamprophyric and doleritic enclaves; mafic, microgranular enclaves are minor (at least at the present level of exposure). Xenoliths of country rock are minor indicating that stoping and assimilation were not significant modes of intrusion. Rafts or roof pendants of relict Flynn Subgroup rocks (100-200m in diameter) in the Mumbilla Granodiorite and the Cabbage Gum Granite do however indicate passive emplacement at shallow depths through stoping. The Tennant Creek Granite has a 100m thick contact metamorphic aureole composed of spotty (quartz and chlorite porphyroblastic) hornfels in distal turbidites of the Warramunga Formation.

'Granite assimilated Flynn Subgroup' and 'Flynn Subgroup assimilated granite' (Figure 1) are major components of the regional, largely non-outcropping, geology which have been interpreted on geophysical evidence by L.Farrar (1994).

Geochronological data of Compston (1995) indicate that the felsic intrusive rocks (c.1860-1833 Ma) in part antedate the felsic volcanic, and sedimentary rocks of the Flynn Subgroup (c.1845-1830).

Geochemically intrusive rocks of the BIA in Tennant Creek province are weakly metaluminous to predominantly peraluminous; high-K calc-alkaline to shoshonitic. They have Sm/Nd DM model ages $\leq$2.6 Ga and εNd at 1.85 Ga $\leq$ - 4.6. Inherited zircon ages are (3.0 Ga (Compston 1995). They are interpreted by Donnellan et al., (in prep.) to be derived from melting of biotite- and hornblende-bearing tonalitic Archaean or earliest Palaeoproterozoic grey gneissic source rocks (cf. Skjerlie and Johnson 1993). Relative to such a source they show marked HREE, Th, Rb and K-enrichment, and marked Sr-depletion but only slight or no Ba depletion.

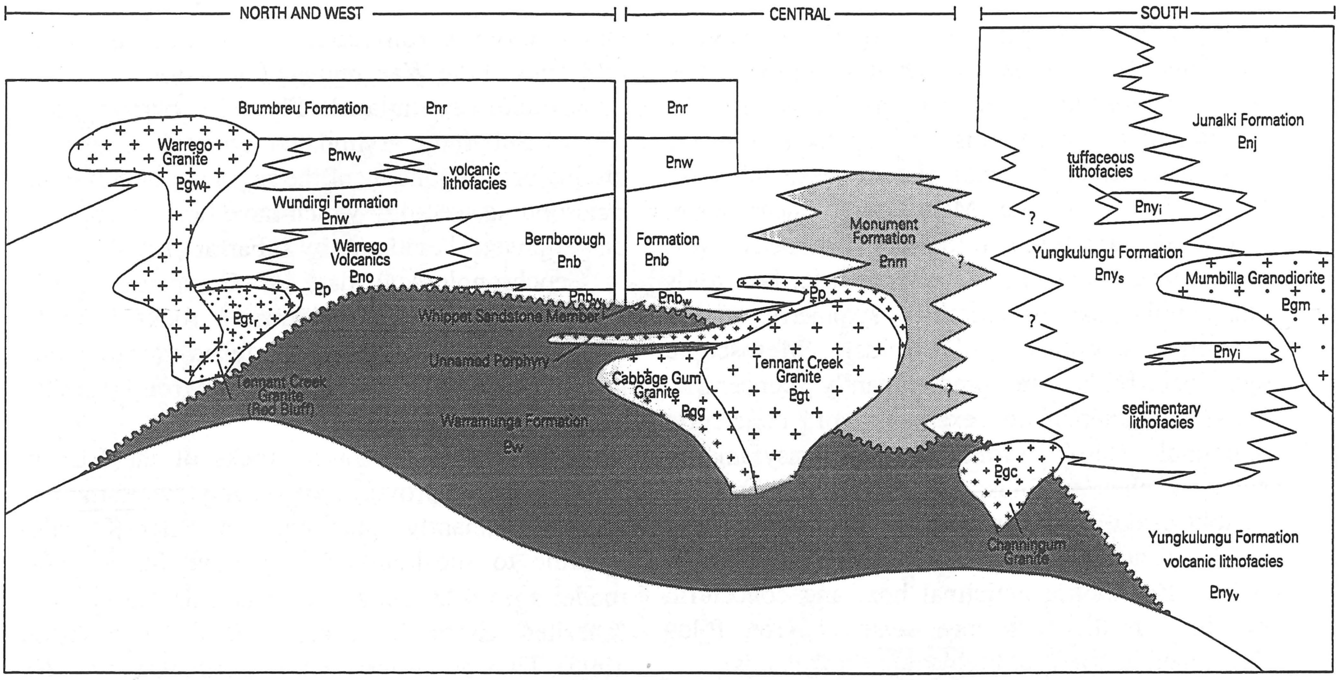
There are at least two episodes of mafic/intermediate magmatism in the Tennant Creek province. One episode is essentially contemporaneous with the syn-tectonic granites of the BIA and comprises dolerite and gabbro. There is a second episode of dolerite, quartz-diorite and diorite, with a late episode of lamprophyre intrusion. Geochronological data of Compston and McDougall (1995) indicate the later episode is coeval with deposition of the Flynn Subgroup (c. 1840-1820 Ma), whilst the lamprophyres intruded at ~1685 Ma. Compston (1995) reports a preferred age of ~1700-1650 for the only representative of the 'late' granites which crops out in the Tennant Creek province, namely the two-mica, peraluminous, S-type Warrego Granite, and 1712 Ma for the Gosse River East 'Granite' which was intersected in drill core.

Flynn Subgroup

The Churchills Head Group comprises the volcano-sedimentary Flynn Subgroup of the Tennant Creek province and the conformably overlying sandstone and subordinate carbonate rocks which, together with minor volcanic rocks, constitute the Tomkinson Creek Subgroup. The latter Subgroup is confined to the Ashburton province, a Palaeoproterozoic stable platform to

Figure: 2a

Schematic summary of the Palaeoproterozoic stratigraphy of the Tennant Creek province.



FLYNN SUBGROUP

Ᵽnr *Lithic and volcanic arenite; heavy mineral-bearing quartz arenite; granule and pebble beds; felsic tuff.*

Ᵽnw *Sublithic and lithic arenite, siltstone, shale, tuff; chert (silicified tuff?).*

Ᵽnw$_v$ *Rhyolitic lava and ignimbrite, tuff and silicified tuff (chert).*

Ᵽno *Chert (silicified tuff?); tuff; white weathering siltstone and shale; fine grained sublithic arenite.*

Ᵽnb *Felsic crystal-lithic tuff, ignimbrite, lapilli tuff, and lava (?); chert. Siltstone, shale and fine grained lithic arenite and wacke.*

Ᵽnb$_w$ *Lithic arenite; siltstone and shale; pebble beds and granule conglomerate; heavy mineral-bearing laminae; cross-laminated or cross-bedded, trough cross-bedded, rippled.*

Ᵽnm *Rhyolitic and rhyodacitic tephra, tuffaceous sandstone and siltstone; chert; shale.*

Ᵽnj *Lithic arenite and volcanic arenite with interbedded laminated siltstone and mudstone. Rhyodacitic lava, crystal-lithic tuff and lapilli tuff, and ignimbrite.*

Ᵽny$_s$ *Quartz-magnetite (heavy mineral-bearing) lithic to sublithic arenite and volcanic arenite, siltstone and shale. Interbedded felsic tuff; siltstone and fine grained volcaniclastic rocks.*

Ᵽny$_i$ *Welded and non-welded felsic crystal-lithic tuff and lapilli tuff; minor felsic volcanic lava and ignimbrite.*

Ᵽny$_v$ *Felsic crystal-lithic tuff and lapilli tuff. Rhyolitic and rhyodacitic lava; foliated to massive and porphyritic. Felsic ignimbrite. Volcanic arenite with interbedded siltstone.*

EARLY GRANITE

Ᵽp *Felsic porphyry; quartz and / or feldspar phenocrysts in a felsic aphanitic groundmass; massive to foliated.*

Ᵽgt *Granite: Biotite-bearing; seriate porphyritic to equigranular, medium to coarse grained and massive. Rapakivi texture, and blue ovoid quartz. Foliated to sheared in part, locally gneissic. Locally abundant porphyritic granite tourmalinisation (minor luxullianite) and alkali metasomatism. (Ᵽgt$_r$, Tennant Creek Granite at Red Bluff).*

Ᵽgg *Granite: Biotite-bearing; seriate porphyritic to equigranular, medium to coarse grained and massive. Rapakivi texture. Foliated to sheared in part, locally gneissic. Minor metasediment xenoliths.*

Ᵽgc *Granite: Biotite-bearing; seriate porphyritic to equigranular, medium to coarse grained and massive. Minor rapakivi texture. Minor metasediment and dolerite xenoliths, rare porphyritic enclaves. Alteration including silicification, albitisation and tourmalinisation (minor luxullianite).*

Ᵽgm *Granodiorite; ranges from tonalite to granite: Biotite-bearing; megacrystic, seriate porphyritic to equigranular, coarse to very coarse grained and massive. Strongly foliated to sheared in part. Locally abundant porphyritic enclaves with rapakivi texture and blue ovoid quartz. Abundant metasediment xenoliths. Localised alteration including silicification, albitisation, and tourmalinisation.*

WARRAMUNGA FORMATION

Ᵽw *Lithic arenite including volcanic arenite ('metagreywacke') siltstone, shale, slate and terrigenous mudstone; 'haematite shale'; minor phyllite; chert including jasper. Partial or complete bouma sequences.*

LATE GRANITE

Ᵽgw *Two-mica, corundum-bearing, granite and granodiorite; quartz-biotite metamorphic aureole.*

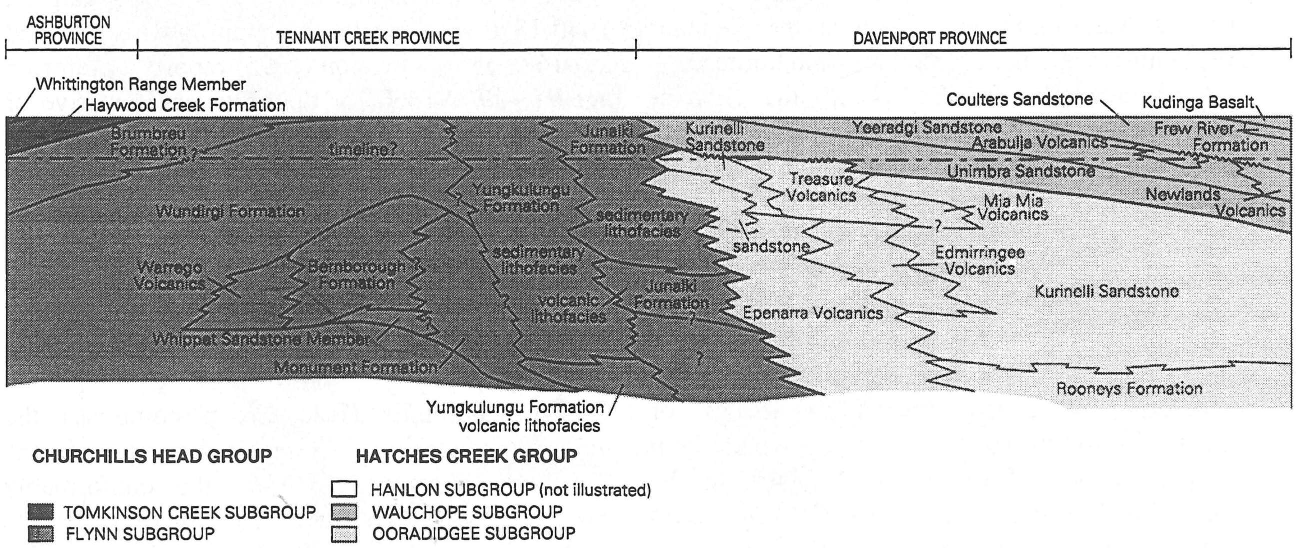


Figure: 2b

Schematic interpretation of the relationships between the Tennant Creek province, Flynn Subgroup and lowermost Tomkinson Creek Subgroup (Haywood Creek Formation), and the Davenport province, Ooradidgee and Wauchope Subgroups. The diagram is based on the stratigraphies presented by : Blake and Horsfall (1986), Blake *et al.* (1987), Stewart and Blake (1986), Wyche *et al.* (1986) (Davenport province), and Donnellan *et al.* (1995) Tennant Creek and Ashburton provinces. The oldest rock unit outcropping in the area is the turbiditic Warramunga Formation (*c* 1860 Ma, Compston 1995).

the north of the Tennant Creek province. The Flynn Subgroup has been divided into three local successions. (1) A predominantly sedimentary succession with subordinate volcanic rocks comprising the Wundirgi and Brumbreu Formations and the Warrego Volcanics which crop out in the north and northwest of the Tennant Creek province. (2) A predominantly volcanic, central, succession comprises the Monument and Bernbourough Formations. The northern and northwestern succession is separated from the southern succession by the locus of 'early' granitic intrusion (Tennant Creek Granite) which is coincident with the predominantly volcanic, central succession. (3) In the southeast of the Tennant Creek province the Yungkulungu Formation comprises lower and upper intervals dominated by volcanic and sedimentary rocks respectively. In the south the Junalki Formation is a probable correlative of the upper, sedimentary part of the Yungkulungu Formation although spatially separated by the Mumbilla Granodiorite. Interpreted relationships between these local stratigraphies are summarised in Figure 2a. The probable lithostratigraphic relationship between the Flynn Subgroup and the lower part of the Tomkinson Creek Subgroup in the Ashburton province, and between the Flynn Subgroup and the Ooradidgee and Wauchope Subgroups of the Hatches Creek Group in the Davenport province are illustrated in Figure 2b.

Sedimentary structures indicate that all local stratigraphies show an upward progression from relatively deep-water, via shallow-marine to fluviatile sedimentation. The volcanic rocks are generally subaerial.

F2/F2' (with respect to the Warramunga Formation) folding is not readily mappable on a regional scale in the Flynn Subgroup, however, the map (Figure 1) shows that the Tennant Creek Province describes a large concentric, NW-SE orientated dome. This corresponds with the well-exposed upright, concentric dome and basin folds (cf. Stewart 1987) arranged en echelon throughout the Davenport province.

Tectonic Setting

The Tennant Creek province is interpreted as a Palaeoproterozoic greenstone (back arc basin)/continental arc and is an example of Mitchell and Reading's (1978) strike-slip geosynclinal/orogenic cycle. The Barramundi geosyncline at Tennant Creek was initiated in response to movement about a dextral shear with a left-hand (releasing bend), contemporaneously with subduction further to the south in the Arunta Inlier. The latter is manifest, for example, in the 1879 Ma gabbro-diorite-tonalite-trondhjemite suite, of the Atnarpa Igneous Complex, which is interpreted as a cordilleran margin (Zhao and McCulloch 1995). The Warramunga Formation was derived from a penecontemporaneous volcanic edifice, probably associated with just such a subduction related cordilleran margin. Subduction at a similar, prior margin was stalled by thickened oceanic plateau crust (cf. Tarney 1992), probably c.1.9 Ga. This plateau crust subsequently cooled and sank. A compensatory upwelling of undepleted asthenospheric mantle introduced heat into the deep crust to produce the BIA at Tennant Creek and initiate extension (cf. Stewart 1987) and bimodal magmatism in the Davenport province. Extension in the Davenport province increased the spatial separation of the Tennant Creek province and the Arunta Inlier. An integrated model for the evolution of the Tennant Creek and eastern Arunta Inliers is presented in detail by Donnellan et.al. (in prep.).

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Crustal Structure in the Archaean and Proterozoic Provinces in Australia

Barry J Drummond, Alexey G. Goncharov & Clive D.N. Collins

Australian Geodynamics Cooperative Research Centre, AGSO
GPO Box 378 Canberra ACT 2601 Australia
bdrummon@agso.gov.au agonchar@agso.gov.au ccollins@agso.gov.au

The Australian continent (including New Guinea to the N) contains geological provinces which range in age from Archaean to Recent. The western two-thirds (approximately) is underlain by Precambrian basement ('the Shield'); basement in the eastern third consists of Phanerozoic fold belts. Except for the far northeast, where the continent is colliding with the Pacific Plate, the topography is generally subdued, and apart for the southeastern mainland, rarely exceeds 1,000 m. Large areas of the Shield have elevations between 500 and 1,000 m. Wellman (1976) found that within the Shield, isostatic compensation was nearly complete near the base of the crust for 3°x3°areas, although density differences in the upper mantle may extend to considerable depths (Dooley, 1991; Drummond, 1997). The near achievement of isostasy is generally reflected in estimates of crustal thickness from seismic studies, with the thickest crust under the areas with the highest elevation. This study looks first at crustal thickness in Australia, and then examines how seismic velocity, which generally reflects density, varies with depth to achieve the compensation. Mechanisms by which the velocity structure might be emplaced are then examined.

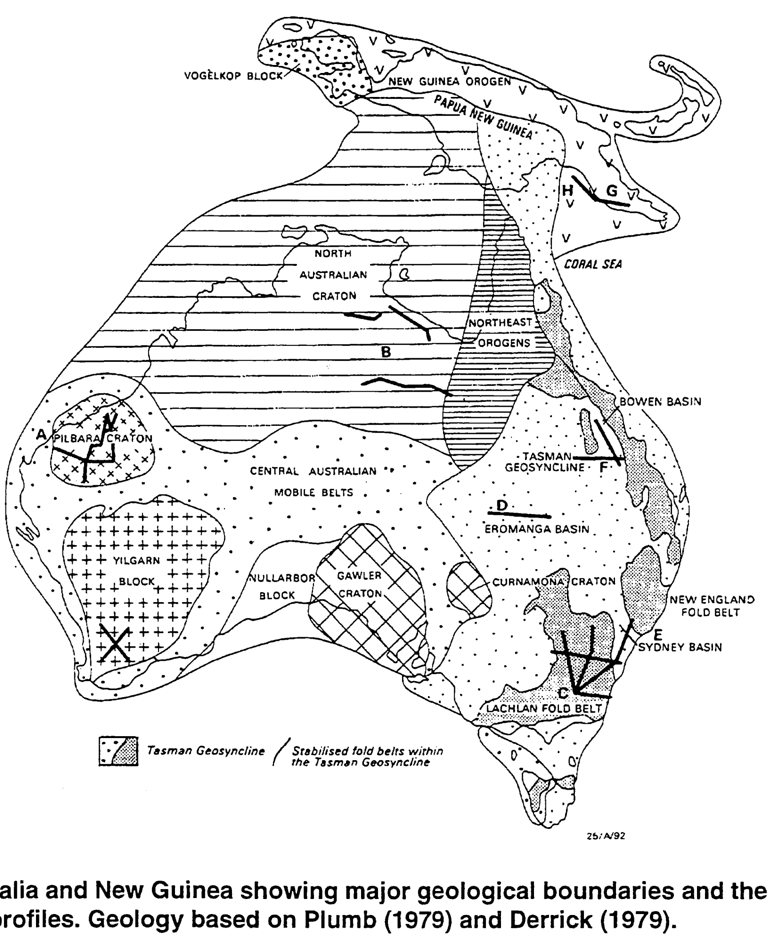

Fig. 1. Map of Australia and New Guinea showing major geological boundaries and the sites of pre-1986 seismic refraction profiles. Geology based on Plumb (1979) and Derrick (1979).

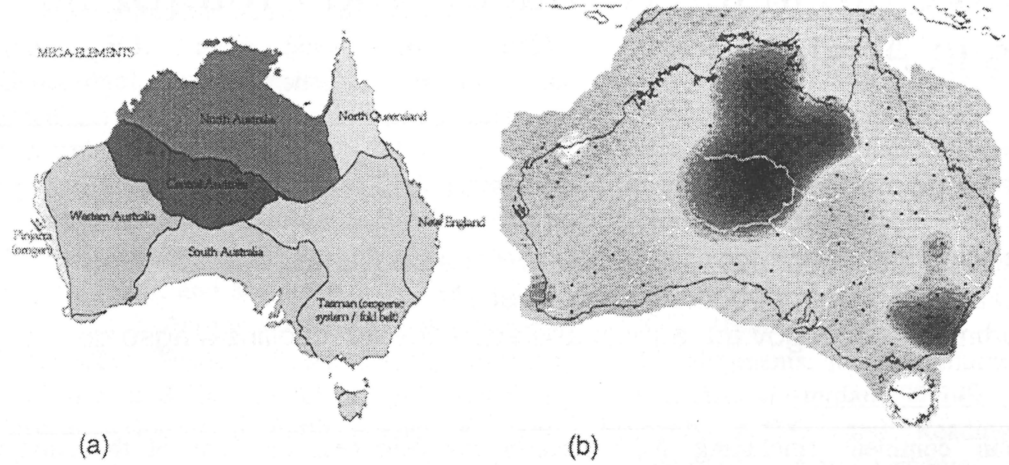

Fig. 2. (a) Mega-elements of Shaw et al. (1996); (b) Moho topography. Black dots show data points. Dark grey in central northern Australia and SE Australia represents crust up to 50 km thick. Light grey represents crust which is mostly between 35 and 45 km thick. The thinnest crust may be under Tasmania (white) and the Archaean Pilbara Craton in the northeast. Images courtesy Simon Cox of the AGCRC; Moho topography image reworked by Ed Chudyk of AGSO.

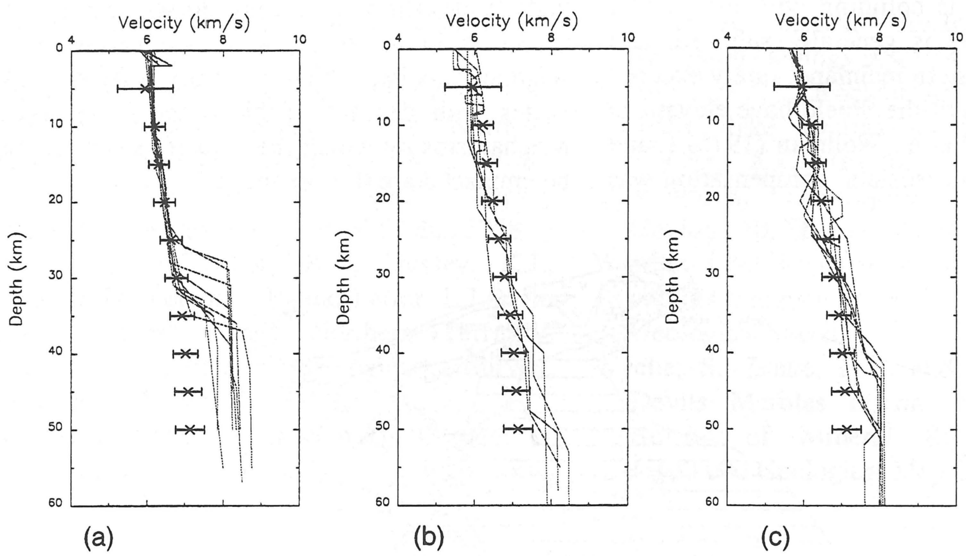

Fig. 3. Velocity versus depth for (a) Archaean, (b) Proterozoic and (c) Early Palaeozoic provinces. Crosses and "error" bars indicate the global average values at 5 km depth intervals plus scatter, from Christensen & Mooney (1995)

Continental fabric

Fig. 2(a) shows the mega-elements of Shaw et al. (1996). These are the basic building blocks of the continent, and are reflected in the regional gravity and magnetic maps of the continent from which they were largely interpreted. Each represents a group of crustal elements with similar geophysical, geological and age characteristics lying within a common set of crustal boundaries. Drummond (1997) suggested that these blocks also have mantle roots extending at least to the base of the lithosphere.

Fig. 2b is a simplified image of Moho depth based on figure 5 of Collins (1991) plus additional depth estimates from crustal transfer functions in the eastern third of Australia (van der Hilst & Kennett, 1997). Although the depths are constrained only by sparse data in the Shield regions, Moho depth shows little if any correlation with the mega-element boundaries. The crust in the Archaean shield is thin in the NE (white) and of average thickness elsewhere (grey). Most of the Proterozoic Shield has average thickness in the south but very thick crust (>45 km, dark grey) in central Australia and northern Australia.

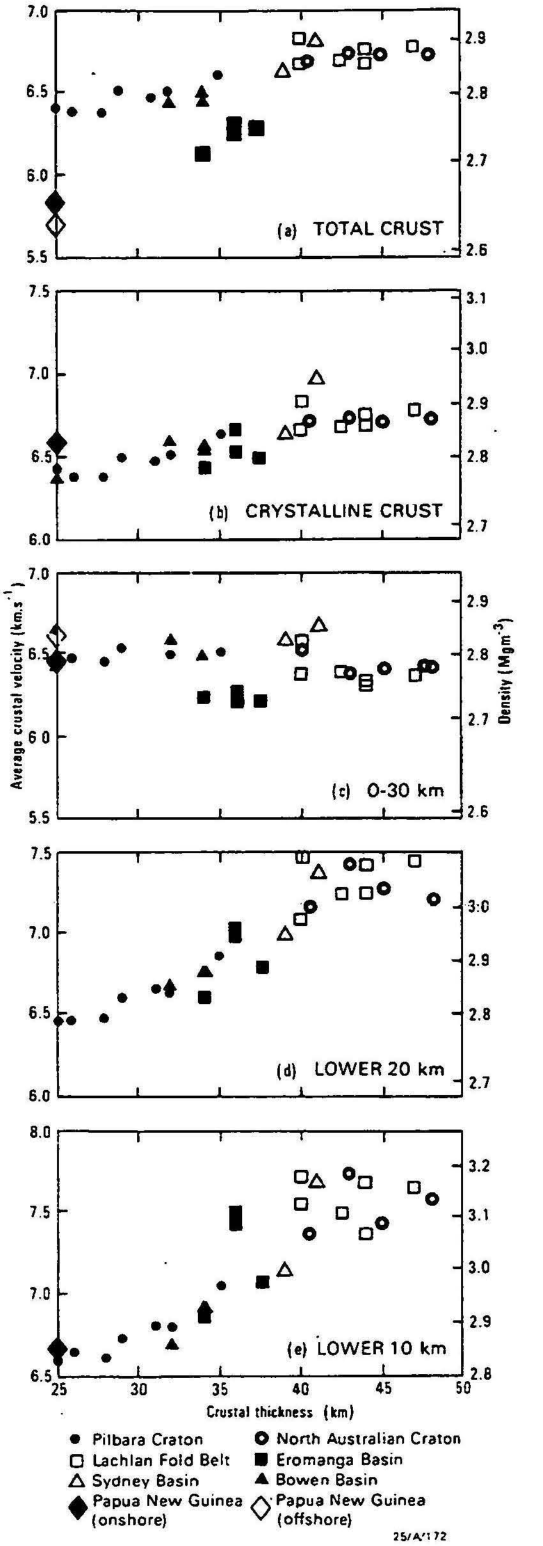

Fig. 4. Average seismic velocities at 25°C and 1Gpa (left axis) and density (right axis) as a function of total crustal thickness. (a) average total crustal velocity (density); (b) average velocity in the crystalline crust; (c) average velocity in the upper 30 km, excluding sediments (or total crust where crustal thickness <30 km; (d) average velocity in the lower 20 km of the crust; and (e) average velocity in the lower 10 km of the crust. (from Drummond & Collins, 1986)

Variation of velocity with depth

Fig. 3 shows the variation of seismic velocity with depth for Australian Archaean, Proterozoic and Early Palaeozoic provinces. Global average values at 5 km depth intervals (Christensen & Mooney, 1995) are also shown. Archaean crust (Fig. 3a) follows global average values until the top of the Moho transition zone, normally between 25 and 35 km depth. The Moho is quite pronounced, usually with a velocity increase of approximately 1.0 kms^{-1} over a depth range of 3-5 km between velocities typical of the crust and those typical of the uppermost mantle. Griffin & O'Reilly (19987) questioned whether the Moho in Archaean terranes was the crust-mantle boundary, or instead the boundary between crustal rocks with a maximum metamorphic grade of garnet granulite (above) and mafic rocks in the eclogitic stability field (below).

Proterozoic crust (Fig. 3a) is thicker than Archaean crust, especially in northern Australia. The velocity/depth curves imply a more heterogeneous upper crust, and a thick lower crust with velocities typically between 7.0 and 8.0 kms^{-1}. Below 35 km, velocities are higher than global averages for comparable depths. The velocity increase across the Moho transition zone is typically <1.0 kms^{-1}. Proterozoic velocity/ depth curves are more similar to Early Palaeozoic curves (Fig. 3c) than to Archaean curves.

The variation of seismic velocities with crustal thickness (depth to the top of the Moho transition zone) is shown in a different way in Figure 4. In Fig. 4a, average crustal velocity for the seismic profiles shown in Fig. 1 is plotted against crustal thickness. Velocities have been adjusted to those expected at 25°C and 1 GPa so that curves from provinces with a wide range of geothermal gradients can be compared. Most points fall in a trend of increasing average crustal velocity with crustal thickness. Several points fall below the trend; these profiles have thick sedimentary basins. If the effects of sedimentary basins is removed (Fig. 4b), all points fall on the trend. Fig. 4c shows average crustal velocity in the top 30 km of the crust (or total crust where the crust is thinner than 30 km). The average velocity in the uppermost crust shows little if any variation with crustal thickness, consistent with Fig. 3 where velocities in the upper 20-30 km generally fall close to global average values. The major variations in average crustal velocity come from variation in the lower crust. Figs. 4d and 4e show average velocity in the lower 20 km and 10 km, respectively. These values

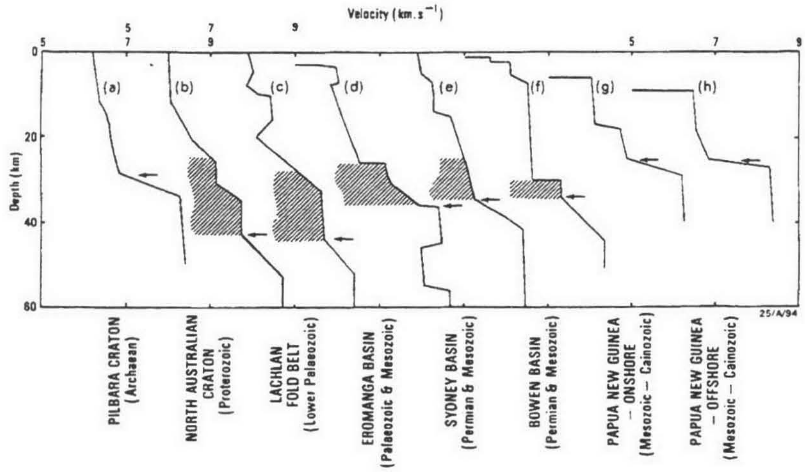

Fig. 5. Representative velocity/depth models for eight geological provinces in the order of province age. Velocities adjusted to 25°C and 1 Gpa. Arrows indicate top of Moho transition zone. Shaded regions have velocities >7.0kms^{-1} (from Drummond & Collins, 1986).

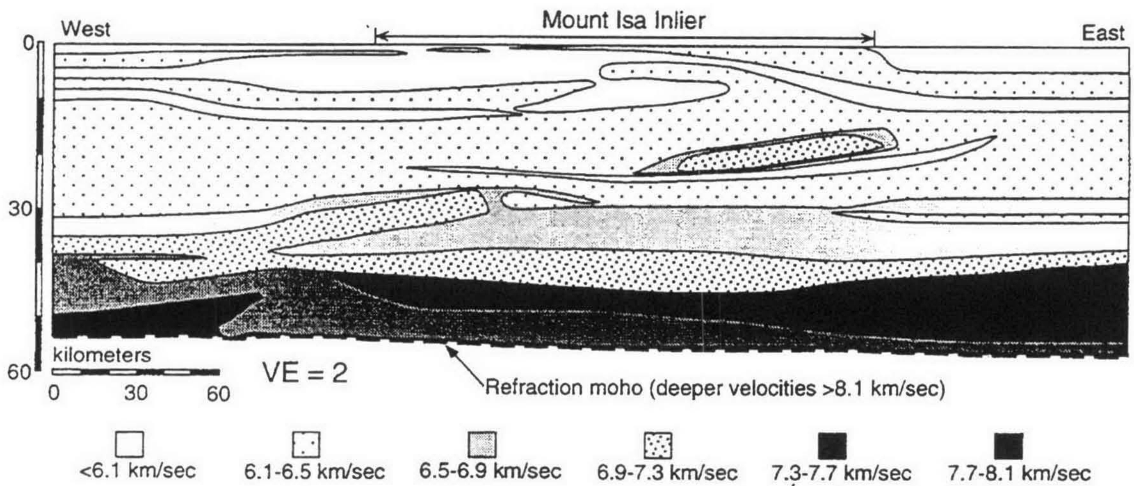

Fig. 6. Crustal cross section from the Mount Isa Inlier. (after Goncharov et al., 1997)

show a marked increase in lower crustal velocity with crustal thickness.

The evolving continent

The right hand axes in Figure 4 show expected density, derived using the empirical velocity-density function of Dooley (1991), and readjusted for ambient temperature and pressure. They indicate that in thick crust, densities of 3.1-3.2 tm^{-3} are common in the lowermost crust. This means that the density difference between the lowermost part of thick crust and the mantle is probably of the order of 0.15-0.25 tm^{-3}. If the uppermost crust has an average density of 2.8 tm^{-3}, and if isostatic compensation is nearly complete at the base of the crust (Wellman, 1976), a kilometre of surface elevation could be compensated by up to 18 km of crustal root. This explains the great variations in crustal thickness which correlate with only modest changes in surface elevation in Australia.

Although their data set was not large, Drummond & Collins (1986) thought that their data showed a general increase in crustal thickness with increasing age of the geological province. An increase in the thickness of high velocity (dense) lower crust was the main cause. This is summarised in Fig. 5. They appealed to underplating as the most likely cause for this increase, with the amount of underplated rock increasing as the provinces evolved. Other models, eg., those which invoke tectonic stacking of the crust, with subsequent reworking into a density-stratified crust, have mass balance problems when applied to large areas of Shield, especially if no high density rock is added in the process. Subduction without mafic addition would be included in this style of model.

Rifting models can emplace mafic rocks into the crust, and increase its average density. Figure 6 shows a cross section of the Mount Isa Inlier (Goncharov et al., 1997). This region has undergone a series of rift and orogenic episodes. The co-linear lenses of high velocity rocks (6.9-7.3 kms^{-1}) dipping from east to west through the middle to upper crust in the models are interpreted as mafic rocks. The seismic models give no indication of the age of these rocks relative to the rest of the crust. Whether they were underplated to thin extended crust which was subsequently shortened, thrusting the lower crust to higher levels, or magmatically emplaced into a post-orogenic thickened crust is uncertain at this time (Goncharov et al., 1997). This cross section does show, however, that high crustal velocities can be explained by models which include mafic rocks at a range of crustal levels in a former rift setting.

The Mount Isa region has higher than regionally average gravity suggesting that it is structurally different from neighbouring regions. Drummond & Collins (1986) felt that models which invoke underplating in a rift setting might be difficult to invoke for large areas of the Proterozoic Shield. Much of the shield has a simpler gravity signature and therefore may have been more stable tectonically.

Why Archaean crust is apparently not underplated remains uncertain. The hypothesis of Griffin & O'Reilly (1987) may apply, and any underplated mafic rocks are now cold and in the eclogite stability field. If so, they would either be seismically indistinguishable from the mantle, or may have delaminated and sunk into the mantle. However, the thermal constant of the crust is such that if the lower crust of Archaean provinces is cold enough to fall in the eclogite stability field, Proterozoic lower crust should probably also be cold and in the eclogite stability field. The available seismic data suggest otherwise. Other explanations may apply, but perhaps the simplest is that Archaean crust was never underplated

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Stratigraphy and structural evolution of the early Proterozoic Svecofennian rocks of south Finland.

Carl Ehlers* and Alf Lindroos

Department of Geology and Mineralogy
Abo Akademi University
FIN-20500 Turku, Finland

The Palaeoproterozoic Svecofennian crust of southern Finland varies in thickness between 42 and 65 kilometres (Luosto 1991) and has a history dating back to some 2.1 Ga which is the age of the oldest detrital zircons in the greywackes of southern Finland (Claesson et al. 1993). The main magmatic epoch took place around 1.89 - 1.88 when most of the syntectonic granodiorites and tonalites intruded in connection with intracrustal collisions and thickening of the crust (Lahtinen 1994). In southern Finland the older crust is metamorphosed and intruded by a second generation of granites and migmatites. These late Svecofennian S-type granites have ages of 1.84-1.83 Ga (Suominen 1991) and they form a 500 kilometres long and 100 kilometres wide belt of intrusions and are associated with high grade metamorphism and formation of migmatites (Fig. 1).

The supracrustal rocks consist of a volcano - sedimentary sequence which exhibits a change in lithology that could correspond to a change in depositional and tectonic environment along a profile from the south coast northwards some 200 kilometres towards the Tampere area. The southernmost part of this profile, off the southern coast line, is characterized by quartz banded, skarn-rich iron formations. North of the iron formations there is a volcanic sequence with frequent thin layers of pillowed basaltic rocks capped by thin layers of marbles and small iron formations. Thin layers of picritic lavas occur in

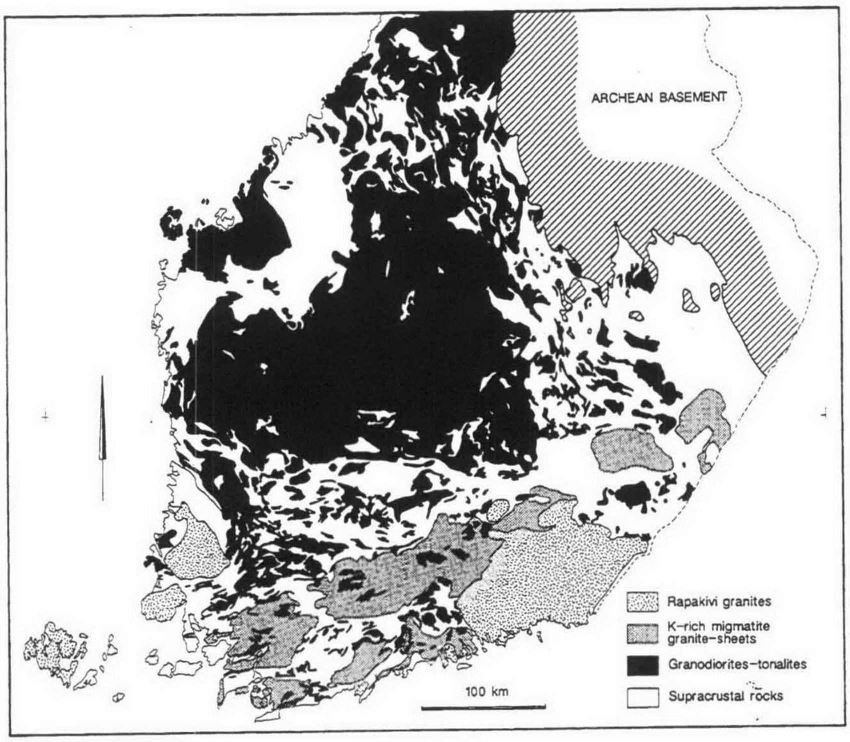


Figure 1

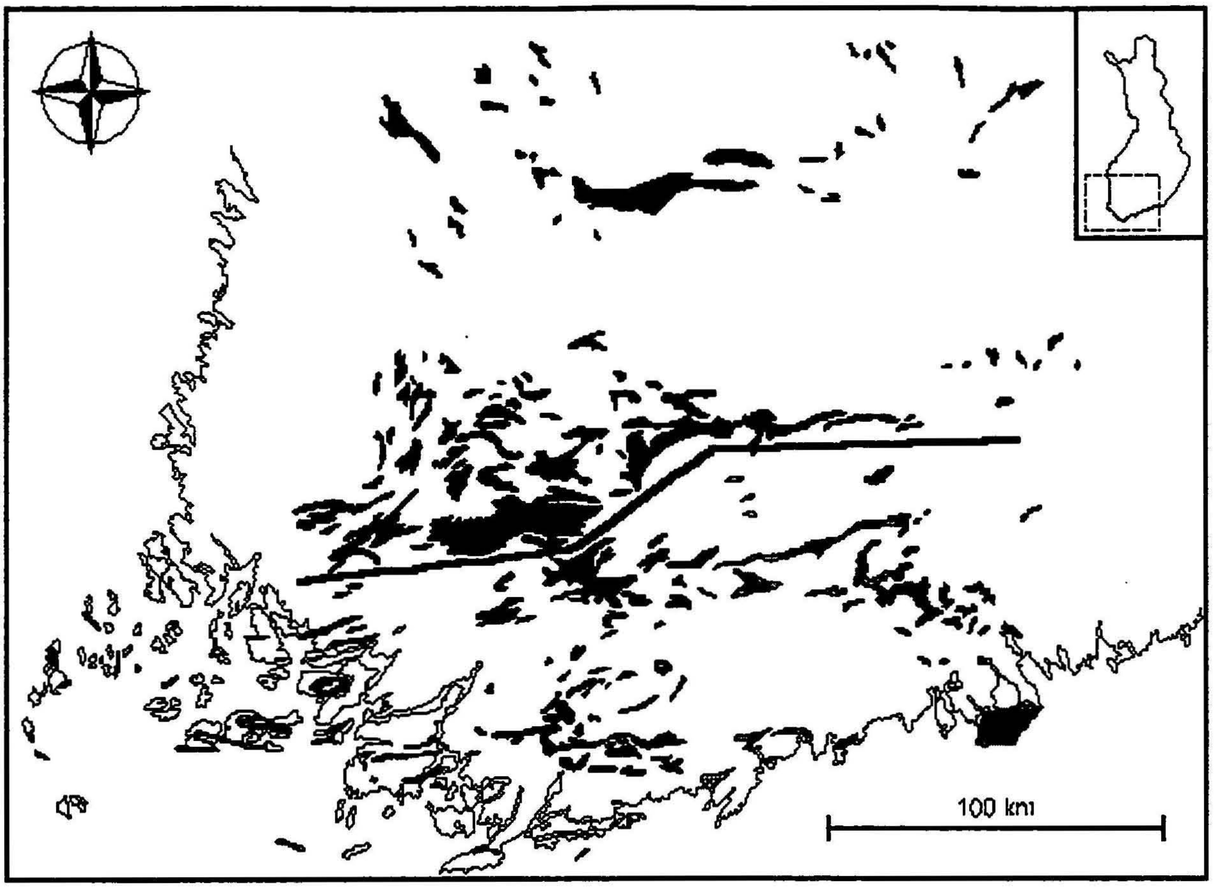
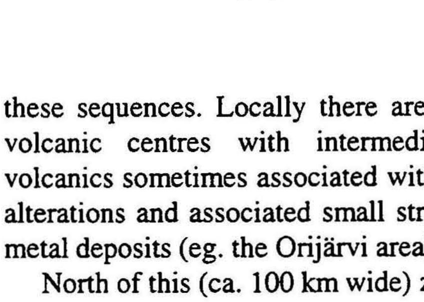


Figure 2A Mafic volcanics in S Finland

Figure 2B. Basaltic rocks from the southern zone (Asterisks) and northern zone (Crosses)



these sequences. Locally there are more evolved volcanic centres with intermediate and acid volcanics sometimes associated with hydrothermal alterations and associated small strata-bound base metal deposits (eg. the Orijärvi area).

North of this (ca. 100 km wide) zone of shallow marine successions overlying pillowed lavas, much thicker mafic to intermediate volcanics occur, consisting of pyroclastics, lavas and epiclastic successions. The thin pillowed lavas and marbles have all but disappeared and greywackes are abundant over a large area . Further north in the Tampere area thick turbiditic sequences underlie the dominantly mafic and intermediate volcanic rocks. Coarse conglomerates are common in the sequences.

Fig. 2A shows the distribution of mafic volcanic rocks in the Svecofennian crust of southern Finland. The domain of thin pillowed lavas with marbles and iron formations along the south coast is shown by grey colored volcanics and the thicker sequences with abundant pyroclastics are black.

The geochemical signatures of the two domains are different as indicated in the diagram in Fig. 2B (only rocks of basaltic composition). This plot has been used to separate basalts with a subcontinental lithospheric source from basalts with a more asthenospheric source, perhaps indicating an extensional environment (Moore et al. 1994). The pillow lavas along the south coast have a less LILE enriched composition (and lie along the x-axis) compared to that of the lavas north of the tentative border between the two volcanic and lithologic domains indicated on Fig 2A. Generally we can conclude that the basaltic lavas within the southern volcanic domain have EMORB characteristics and

A tentative history of the Proterozoic crust in the migmatite zone of SW Finland.

Oldest detrital zircons in graywackes	2.1 - 2.0 Ga
Volcanics and sediments	> 1.9 (not well dated)
Intrusion of early granodiorites/tonalites - strong folding an stacking of sequences, thickening of crust - traces of early high grade metamorphism	1.89 - 1.88 Ga
Sheet intrusions of less deformed granodiorites	1.87 - 1.86 Ga
Intrusion of early steep metabasaltic dykes indicating local extension of thickened crust	1.86 Ga
Continued suborizontal shearing and stacking - thickening of crust	
Main episode of high grade metamorphism, formation of migmatites and intrusion of sheets of S-type microcline granites, last traces of subhorizontal def.	1. 84-1.8 Ga
Intrusion of small bimodal postorogenic intrusions and wide spread intrusion of pegmatites in a brittle regime - wide spread crustal extension and exhumation - only vertical shearing.	1.80 Ga
Exhumation and peneplanation (200 Ma of quiesence)	
Intrusion of anorogenic (?) bimodal rapakivi/anorthosite intrusions	1.6 Ga

shallow marine lithologies, suggesting extension and rifting, while the basaltic lavas northwards chemically and lithologically indicate an arc environment.

The Zone of migmatites and granites along the south coast of Finland is bordered by vertical regional shears and coincides precisely with the southern volcanic belt. Structurally the lithological units within this zone are compressed and stacked/overturned towards W and NW forming a subhorizontal schistosity subparallel to the present peneplane. Zones of intense subhorizontal thrusting and shearing are often concentrated in the thin marble layers which are thinned and sheared out. Subsequently both the early granitoids of the 1.89 - 1.88 Ga group and the late 1.84 - 1.83 Ga old microcline granites intruded along these early formed and reactivated subhorizontal schistosities and forms a layered sequence of strongly sheared early supracrustals and more or less sheared sheets of granites.

The higher metamorphism and late intrusion of granite sheets confined to this lithological zone along the south coast zone could be the result of a transpressional inversion of an extensional rift basin which has acted as an inherited zone of weakness in the Proterozoic crust.

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Palaeoproterozoic Crust of the Australian and Fennoscandian Shields: Geological Implications of Seismic Velocity Models

Alexey G. Goncharov & Barry J. Drummond

Australian Geological Survey Organisation, GPO Box 378, Canberra, ACT, 2601, Australia, e-mail: agonchar@agso.gov.au

Seismic velocity models of the Australian and Fennoscandian shields generally support the concept of thick Proterozoic and thin Archaean crust. Underplating by mafic melts near the base of the crust is likely to be a primary geological process responsible for this difference. Several important exceptions from this general rule revealed in the Fennoscandian Shield suggest that in some cases development of the crust at different depth levels was asynchronous. Seismic low velocity layers are quite common in the Paleoproterozoic crust and are due to variations in both composition and fabric of the rock. A balancing of high and low velocities along any vertical profile through the crust, if observed, may constrain geological interpretations of the seismic data. Petrological models derived from the seismic data delineate a source region for Proterozoic felsic rocks in the dioritic to gabbroic lower crust and also pose a question about the differences in upper mantle petrology of the two shields.

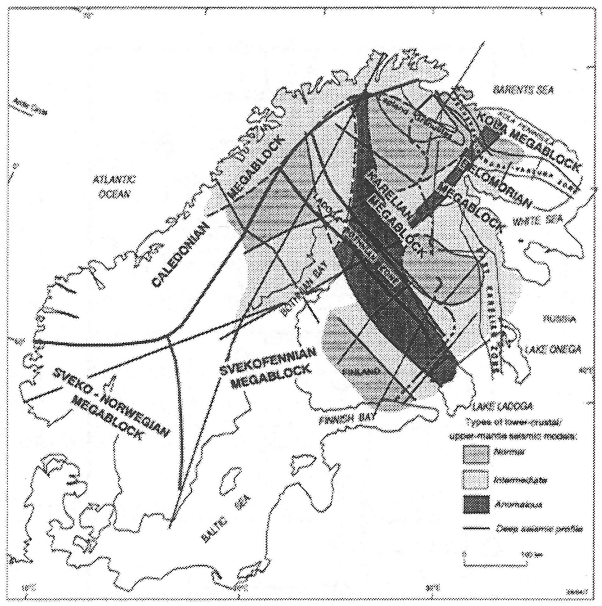

Fig. 1. Location of seismic profiles and regions of the Fennoscandian Shield defined by different types of lower-crustal and upper-mantle seismic models. Normal and anomalous models shown in Fig. 4, intermediate model is a transitional between these two.

Introduction

The crust and upper mantle transition zone in the Fennoscandian Shield has been studied by numerous seismic profiles including, more than 3000 km of reflection and deep seismic sounding (DSS) profiles in the eastern part of the Shield (Fig. 1). The profiles crossed all major structural elements of the region: Archaean basement outcrops, Proterozoic mobile zones, greenstone belts, and zones of Phanerozoic reactivation. The DSS data in the eastern part of the shield were recorded by reversed and overlapping profiles with 100-200 m geophone intervals. The recording geometry is very important, as it shows that the coverage of the DSS profiles in the eastern part of the Fennoscandian Shield is more dense compared to the western part of the shield and to many other parts of the world. This enabled the development of very detailed seismic models of the crust and the crust-mantle transition zone in the eastern part of the Baltic Shield (Goncharov et al.,1991).

Significant parts of the North Australian Precambrian Craton and Archaean Yilgarn block in Western Australia have been covered by refraction and wide-angle reflection seismic profiles. Seismic velocity models of the Australian Precambrian were summarised by Collins (1988) and interpreted by Drummond and Collins (1986).

Velocity Models

Fennoscandian Shield

Low velocity layers are quite common in an overall complicated velocity distribution in the crust and upper mantle. A typical example of the upper-crustal velocity models comes from the region of the Kola Super Deep Bore Hole (KSDBH). Seismic data from the KSDBH have radically altered the conventional idea of a monotonous velocity increase with depth in crystalline crust: the velocity profile has many low-velocity intervals, velocity decreases reach 1 kms^{-1} and were found even at a depth of more than 10 km (Fig. 2).

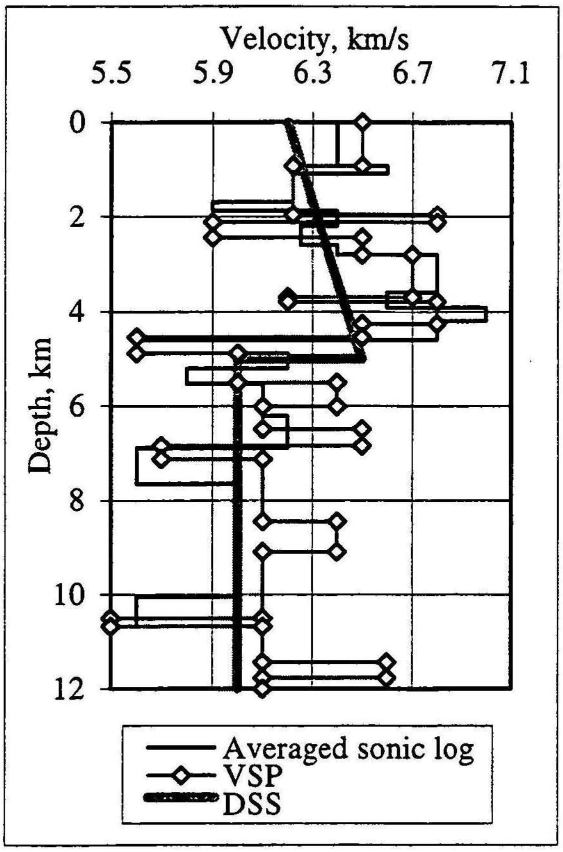

Fig. 2. Typical velocity models of the upper crust in the Fennoscandian Shield derived from the sonic log, vertical seismic profiling (VSP) and DSS data in the KSDBH and its region.

In most models the boundary between the crust and mantle is a thick (up to 10 km) transitional zone rather than a sharp discontinuity. Alternation of high- and low-velocity layers is also typical for this zone in some cases. The evidence for this is the form of the super-critical reflections from the Moho discontinuity (PMP, Fig. 3).

Lower crustal and upper mantle velocity models vary significantly from one area of the Fennoscandian Shield to another and can be subdivided into 3 groups: normal, intermediate and anomalous (see Figs.1 & 4). "Normal" models (the name is historic, and comes from early deep seismic research in the eastern part of the shield,

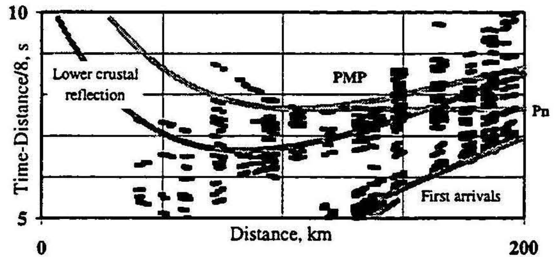

Fig. 3. Typical travel times of seismic waves in the eastern part of the Fennoscandian Shield. The first arrivals are discontinuous and time delayed and experimental later arrivals (dots) have systematically higher apparent velocities than the PMP and lower crustal reflection (continuous lines) computed for the model with simple sharp velocity boundaries. Pn - refraction in the upper mantle.

when relatively thin crust seemed to be the most common) defines the thinnest crust with little or no high-velocity (more than 7 kms^{-1}) rocks at its base, while "anomalous" crust corresponds to the thickest crust with a large volume of high-velocity rock in its lower part. Crustal velocities in the normal model are slightly lower, and in the anomalous model slightly higher, than in the global average model for shields and platforms (Fig. 4). Total thickness of the crust in the normal model is very close to the global average value (41.5 km), while in the anomalous model where the shallowest Moho was defined at 45 km depth, it is noticeably higher.

Australian Shield

Low velocity layers were also recognised as important seismic features of some Australian Precambrian terranes (Drummond et al., 1995). Velocities from the top of basement to around 30 km depth are similar in all Australian Precambrian areas. Where thickening of the crust occurs in Australia it is totally due to the thickening of the lower high- velocity (more than 7 kms^{-1}) crust. Velocities in both Australian Archaean and Proterozoic crust correlate well with the global average model for shields and platforms (Fig. 4). The Australian Archaean crust is on average thinner (about 35 km) than in the global average model (41.5 km).

A remarkable difference between the Australian Archaean and the Fennoscandian normal models, apart from the noted difference in crustal thicknesses, is the presence in the latter model of multi-layered Moho (or a thick zone of

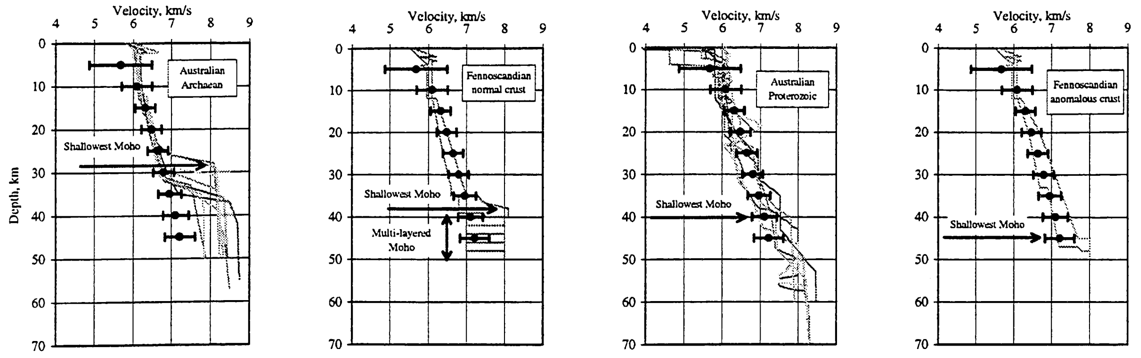

Fig. 4. Seismic velocity models of the Australian Archaean and Proterozoic crust compared to the models of normal and anomalous crust of the Fennoscandian Shield and to the global average model for shields and platforms (shown by dots with error bars, after Christensen & Mooney, 1995).

alternating high (8 kms^{-1}) and low (modelled as 7 kms^{-1}) seismic velocities, Fig. 4). For any low velocity layer in this model only the product of the velocity within the layer and the layer thickness can be estimated from the travel time data. This means that the existence of low-velocity layers in the model is certain but the actual velocity distribution within the low velocity layer is non-unique. Hence velocities as low as 7.0 kms^{-1} below the Moho represent only one solution amongst numerous possible. The difference in the sub-Moho velocity structure between Australian and Fennoscandian models may reflect different sub-Moho velocity structures in these regions. Alternatively it could be more apparent than real due to the resolution of the techniques used: observation geometries in the Australian Archaean terranes were less dense than those in the eastern part of the Fennoscandian Shield. There is evidence for inter-layering of high and low velocities in Australian shield, particularly in Proterozoic northern Australia and the Archaean Pilbara Craton. Drummond et al. (1995) and Drummond (1988) suggested that the Australian crust was far more heterogeneous than the simple velocity/depth models would suggest.

Geological Implications

Low velocity layers as indicators of geological processes

Low-velocity layers are due to variations in both composition and fabric of the rock. Sometimes they have clearly tectonic origin, for example in the KSDBH region in the depth range 4.5 - 7.0 km. They can also be explained by alternation of rocks with higher and lower SiO_2 contents. In many cases a balancing of high and low velocities can be seen along any vertical profile through the crust. This balancing under certain P-T conditions may translate into petrological models in which the proportional distribution of felsic, intermediate, and mafic rock in the crust is also balanced. In such cases evolutionary models of the crust are more consistent with magmatic rather than tectonic development: fractionation during melting of the crust will result in the depletion of its lower levels underneath high-velocity anomalies in the mafic component (Goncharov et al., 1997).

Thin Archaean and thick Proterozoic crust

Results of the deep seismic studies of the Australian Precambrian terranes support the concept of thickened Proterozoic crust compared to Archaean crust, and underplating by mafic melts near its base was suggested as the main process responsible for this thickening (Drummond & Collins, 1986). Generally the results from the Fennoscandian Shield are consistent with this. The evolution of its crust from the one characterised by the normal seismic model, through intermediate, to the other characterised by the anomalous seismic model (see Fig. 4) may be suggested. However in some cases there is no direct correlation between the type of the lower crustal seismic model and the age of rocks at the surface. For example, the anomalous thick crust is typical for the Ladoga-Bothnian zone in Finland and its possible extension in Russia which is not manifested at the surface (see Fig. 1). The age of rocks at the surface varies from Archaean to Paleoproterozoic along this deep crustal structure. Thin crust characterises only the central part of the Archaean Karelian Craton while towards its edges the crust is thickened, obviously due to the Proterozoic processes of crustal reworking similar to those documented in the Paleoproterozoic East-Karelian mobile zone (see Fig. 1). In contrast, large parts of the Proterozoic Svekofennian megablock have thin crust (normal seismic model, see Fig. 1) which may have been preserved there since Archaean time. These examples imply that in some cases the development of the crust at different depth levels was asynchronous and that some geological processes in the deep crust are not necessarily manifested at the surface immediately above corresponding geophysical anomalies.

Generalised petrological models of the crust and upper mantle

Three average seismic velocity models were interpreted in terms of the petrology of the deep crust: one typical for the Australian Archaean crust, another for the Fennoscandian normal crust and a third one representing both Australian Proterozoic crust and Fennoscandian anomalous crust. Separate layers in the resulting petrological models are characterised by a likely proportion of rocks of different bulk geochemistry. Bulk geochemistry of the rocks in all three models varies from almost 100% granite composition of the upper 20-25 km of the crust to ultramafic composition of the upper mantle. The velocity distribution of the upper mantle in the Australian Archaean terranes immediately underneath the Moho may be explained by 100% spinel lherzolite. Alternating ultramafic and more felsic rock layers will explain the observed seismic low velocity layers beneath the Moho in the Fennoscandian normal model. The exact proportion of different rock types within this multi-layered upper mantle cannot be defined due to the ambiguity of the seismic model.

Velocities in the Australian Proterozoic and Fennoscandian anomalous crust require a >15 km-thick layer of dioritic to gabbroic bulk composition in the lowermost crust. In contrast, the Fennoscandian normal crust and Australian Archaean crust require no more than 5 km of such rock in the lower most crust.

Geochemical features of Australian Proterozoic felsic rocks imply that they were derived from pre-existing plagioclase-bearing lower crustal sources, which must have been voluminous to have sourced the huge volume of felsic igneous rocks (Wyborn et al., 1992). The dioritic to gabbroic petrological composition determined for the Australian Proterozoic lower crust is consistent with the inferred origin of the granites: such a source would have contained abundant plagioclase and been in the appropriate compositional range to generate I-type granites (granodiorites). Some of the layers of predominantly gabbroic bulk geochemistry in the lower crust inferred from the seismic velocities would represent the more mafic, denser residues left after the extraction of felsic partial melts from an original dioritic source (Goncharov et al., 1997).

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Gold Mineralisation in the Pine Creek Geosyncline Northern Territory, Australia

John Goulevitch

Exploremin Pty Ltd
6 Porter Street Ludmilla NT 0820 AUSTRALIA

Over 2.88 million ounces (90 tonnes) of gold have been produced from the Pine Creek Geosyncline over the past 127 years. Of this, 86% has been produced since 1980. At the end of 1996, in excess of 9.3 million ounces (291 tonnes) were contained in Identified Mineral Reserves and Resources and sub-economic resources.

Scientific investigation

The earliest scientific investigations of gold in the region were undertaken nearly 100 years ago. More relevant in the present context are regional reviews conducted in the past 20-30 years by:

Ahmad et. al. (1993)
Crohn (1968)
Needham and Roarty (1980)
Needham and Stuart-Smith (1984)
Nicholson and Eupene (1984)
Nicholson and Eupene (1990)
Ormsby et. al. (1994)
Partington and McNaughton (1997)
Stuart-Smith et. al. (1993)

and studies on specific deposits during the same period by:

Alexander et. al. (1990) - *Cosmo Howley*
Cannard and Pease (1990) - *Enterprise*
Carvill et. al. (1990) - *Coronation Hill*
Dann and Delaney (1984) - *Enterprise*
Eupene and Nicholson (1990) - *Iron Blow/ Mt Bonnie*
Goulevitch (1980) - *Iron Blow/Mt Bonnie*
Hancock et. al. (1990) - *Jabiluka*
Hellsten et. al. (1994) - *Union Reefs*
Kavanagh and Vooys (1990) - *Woolwonga*
Matthai et. al. (1996) - *Cosmo Howley*
Miller (1990) - *Moline Dam/Northern Hercules*
Newton et. al. (in prep) - *Union Reefs*
Nicholson (1980) - *Golden Dyke*
O'Keefe (unpubl.) - *Faded Lily*
Ormsby et. al. (in prep) - *Mount Todd*
Partington et. al. (1994) - *Western Arm, Bridge Creek*
Poxon and Hein (1994) - *Mount Todd*
Prichard (1965) - *South Alligator Valley*
Quick (1994) - *Goodall*
Rabone (unpubl.) - *Rustlers Roost*
Sheppard (1996) - *Tom's Gully*
Simpson (1990) - *Tom's Gully*
Simpson (1994) - *Sundance*
Tsuda et. al. (1994) - *Tom's Gully*
Turner (1990) - *Union Reefs*
Valenta, (1991) - *Coronation Hill*
Wyborn et. al. (1994) - *Coronation Hill*

Distribution of gold deposits

Significant gold deposits in the Pine Creek Geosyncline are located on Figure 1. The regional geological framework is described by Ahmad earlier in this volume.

Styles of gold mineralisation

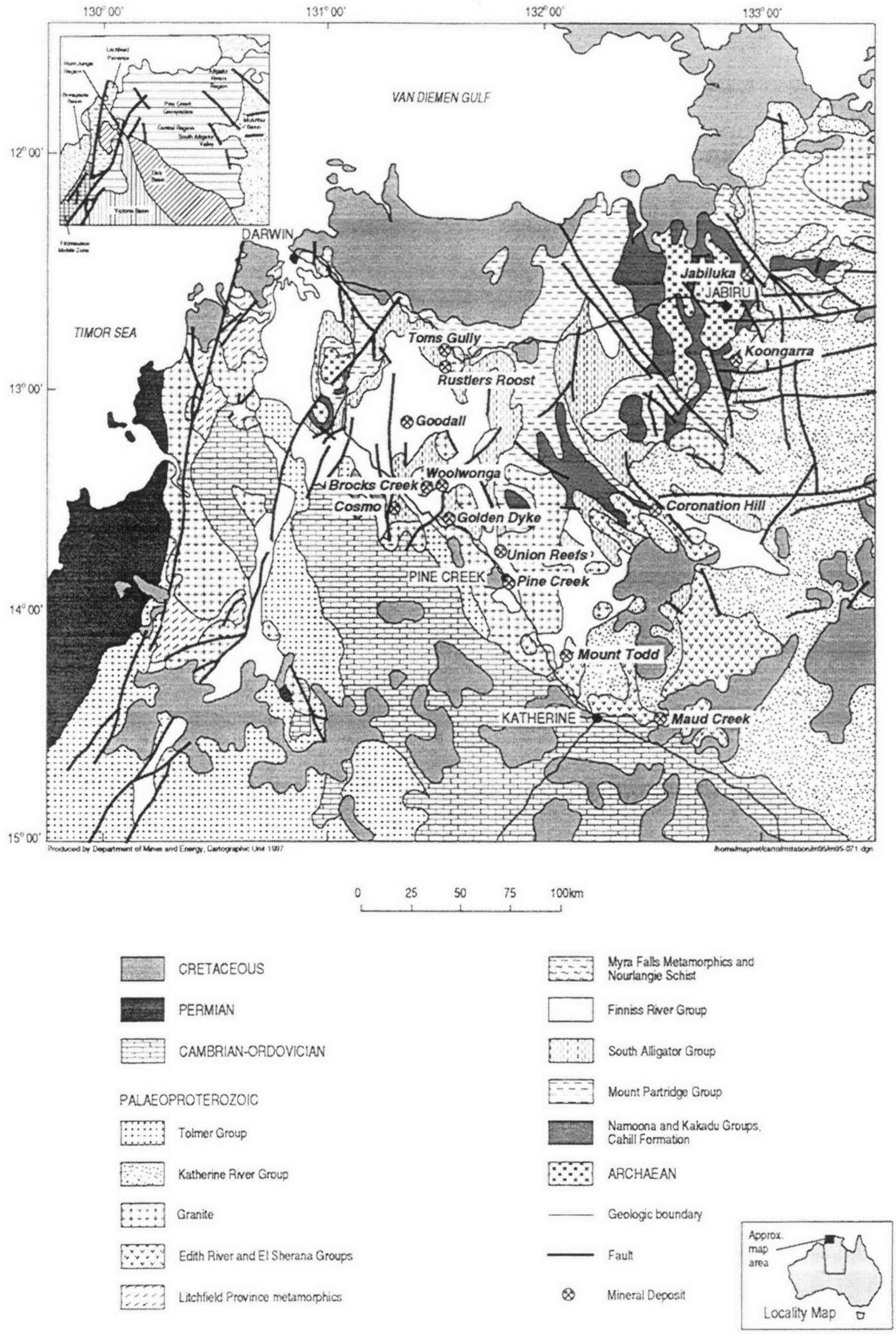
Cherty, sulphidic (pyrrhotite, pyrite, ±arsenopyrite) iron formation-hosted (± quartz-sulphide veins)
Cosmo Howley, Golden Dyke, Mount Porter, Fishers Lode, Black Rock, Davies No 2, Fenton, Maureen, Quest 29

Bedded chert-sulphide (± dolomitic siltstone)-hosted
Zapopan, Rustler's Roost, Northern Hercules

Stratiform/stratified (massive) silicate-carbonate-sulphide lodes (poly-metallic)
Mount Bonnie, Iron Blow, ?Moline Dam, ?Daly River

Poly-metallic carbonate replacement
Woodcutters

Quartz-sulphide veins ± disseminated sulphides (pyrite, arsenopyrite)
<u>Sediment/tuff-hosted</u>
Multiple vein (±breccia zones) - *Enterprise-International-Gandys Hill, Mount Todd, Cosmo-Howley, Woolwonga, Chinese Howley, Big Howley, Union Reefs, Union*



REGIONAL GEOLOGY WITH MAJOR GOLD DEPOSITS

Figure 1

Extended, Faded Lily (Brocks Creek), Alligator (Brocks Creek), Goodall, Spring Hill, Fountain Head, Yam Creek, Western Arm, Glencoe, Backhoe (Rustlers Roost), Temperance, Bridge Creek, Watts Creek, Fletchers Gully, Zapopan, Quigleys, Golf, Tollis

Single vein (-shear zone) - *Tom's Gully, Virginia, Frances Creek, Bandicoot, Marrakai, Great Northern, Great Western, Star of The North*

Breccia/Shear Zone - Maud Creek

Dolerite-hosted

Maud Creek, Chinese Howley South, Margaret Diggings, Quest 29, Maureen

Uranium-associated

Sediment-hosted - *Jabiluka, South Alligator Valley mines, Coronation Hill*

Intrusive-hosted - *Coronation Hill*

Breccia-hosted - *Coronation Hill, Jabiluka*

Unconformity-associated

Sediment-hosted - *Sundance*

Stratigraphic controls on gold mineralisation

Palaeoproterozoic

Mount Partridge Group

Mundogie Sandstone - *Frances Creek*

Wildman Siltstone/Whites Formation - *Tom's Gully, Watts Creek, Woodcutters*

South Alligator Group

Koolpin Formation - *Cosmo Howley, Golden Dyke, Mount Porter, Fishers Lode, Black Rock, Davies No 2, Fenton, Maureen, Quest 29, South Alligator Valley deposits, Bridge Creek*

Gerowie Tuff - *Chinese Howley, Big Howley, ?Alligator, Zapopan, Coronation Hill, Mount Porter (North)*

Mount Bonnie Formation - *Enterprise-International-Gandys Hill, Faded Lily, ?Alligator, Spring Hill, Woolwonga, Glencoe, Yam Creek, Princess Louise, ?Fountain Head, Mount Porter (North), Mount Shoobridge, Rustlers Roost, Western Arm*

Cahill Formation (?Meta-South Alligator Group)- *Jabiluka*

Finniss River Group

Burrell Creek Formation - *Mount Todd, Golf, Tollis, Quigleys, Union Reefs, Union Extended, Goodall, ?Fountain Head, Virginia, Bandicoot, Marrakai*

Noltenius Formation - *Fletchers Gully*

Tollis Formation - *Maud Creek*

Zamu Dolerite - *Chinese Howley South, Quest 29, Margaret Diggings, Maureen*

Maud Creek Dolerite - *Maud Creek*

Quartz-Feldspar Porphyry, Quartz Diorite - *Coronation Hill*

Cullen Batholith - *Bonrook*

Post-Palaeoproterozoic

?Coronation Hill Breccias - *Coronation Hill*

Tertiary-Recent - *alluvial deposits including Bridge Creek, Fountain Head, Pine Creek, Wandie, Yam Creek, Port Darwin Camp, Union Reefs*

Structural controls on gold mineralisation

Vein-deficient iron formation/chert/dolomitic siltstone-hosted mineralisation

- Occurs both isolated on fold limbs and also in axial zones of regional D2 (Partington and McNaughton, 1997) anticlines

Quartz-sulphide vein-type mineralisation

- Displays a strong spatial association with regional NNE-NNW trending D2 anticlines though concentrations of mineralisation restric-ted to synclinal closures are known.
- The D2 anticlines are commonly overturned to the east.
- Mineralisation is often more extensively developed in axial zones and on the west limbs than on the east limbs of the anticlines.
- Easterly verging ?D2 thrust fault±breccia systems are associated with the anticlines and the mineralised zones - but similar unmineralised thrust systems appear to be extensively developed throughout the Pine Creek Geosyncline.
- Early thrust faults in the Pine Creek Geosyncline are folded by the D2 anticlines - late thrust faults cut/accentuate D2 anticlines.
- Mineralised veins are often localised along bedding planes and axial plane cleavage - irreg-ular stockwork veins and regular and irregular sheeted veins are extensively developed.
- Early veins are folded at Union Reefs.

Age of mineralisation

Sediment-hosted vein- deficient mineralisation

- ?Syn-sedimentary, ?Syn- to post-regional metamorphism/D2 deformation, ?Syn-granite event

Vein-type mineralisation

- Syn/post-regional metamorphism/D2 deformation
- ?Pre/syn-granite event (Mt Shoobridge pegmatites cut gold mineralisation)

Mineralisation in granite

- Rare - Bonrook shear-greisen zone

Fluid inclusion & isotope studies

Vein-type mineralisation

- Mixed magmatic and metamorphic source for fluids

Genetic models

Syngenetic/Syn-sedimentary

- Cherty iron formations of the middle Koolpin Formation are regionally anomalous in gold and arsenic - mineralised cherty iron formation at Golden Dyke passes laterally into a bedded tourmalinite lens (Nicholson, 1980)
- Massive gold-enriched poly-metallic deposits 3 kilometres apart at Iron Blow and Mount Bonnie are stratifom, stratified and precisely stratabound (Goulevitch, 1980) - 10 kilometres to the west bedded tourmalinite occurs in the same stratigraphic position (Stephens, 1978)

Epigenetic

- Primary metamorphism-derived fluids
- Primary granite/granitisation-derived fluids
- Remobilised syngenetic/epigenetic mineralisation in Koolpin Formation
  - Diagenesis driven
  - Granite/granitisation driven
  - Metamorphism driven
  - Tectonism driven

The close spatial relationship between iron formation-hosted deposits in the Koolpin Formation and sediment/tuff-hosted quartz-sulphide vein-type deposits in overlying units (eg. Cosmo-Howley/Goodall line, Golden Dyke/Yam Creek line, Mount Porter/Mount Porter North line) implies a possible genetic relationship in either of the following ways.

1. Either both forms of mineralisation are expressions of the same hydrothermal event; or
2. The epigenetic deposits higher in the sequence/structure have been derived from syn-sedimentary or earlier epigenetic mineralisation lower in the sequence during tectonism, granitisation/granite intrusion, contact metamorphism or regional metamorphism (+/- tectonism).

While the combination of structural, stratigraphic and lithological controls outlined by Nicholson and Eupene (1984, 1990), Ormsby et. al. (1994) and Partington and McNaughton (1997) are universally accepted in a general sense for the quartz-sulphide vein-type mineralisation, opinions differ with regard to the origin of the stratabound, stratiform, vein-deficient gold mineralisation.

The emphasis placed by Matthai et. al. (1996) on stratigraphically and structurally controlled, vein-type gold mineralisation in the crest of the anticline at the Cosmo Howley deposit, ignores the abundance of high-grade gold mineralisation in both arsenopyrite-rich and arsenopyrite-poor cherty iron formations which are free of quartz veins. Vein-free, iron formation-hosted, gold mineralisation, sometimes with much higher grade than that in vein-type deposits, is the dominant style of mineralisation encountered at depth on the limbs of the anticline at Cosmo Howley and it was also extensive in the mined area. This is also the case at the Golden Dyke Mine and at other prospects where cherty iron formation-hosted gold mineralisation has been identified and extensively explored. As a result, a syn-sedimentary origin for the sediment-hosted mineralisation, particularly that in the cherty iron formations of the middle Koolpin Formation, is still widely accepted despite recent epigenetic interpretations by Matthai et. al. (1996) and Partington and McNaughton (1997).

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Uranium Deposits of the South and East Alligator Rivers Region

C P Hallenstein

AFmeco Mining and EXploration Pty Ltd.
GPO Box 2142, Darwin NT 0801
Australia

The Alligator Rivers Uranium Field (ARUF) lies about 250 km East of Darwin (Figure 1) and contains some of the world's major uranium deposits. All were discovered between 1969 and 1973 but to date only two deposits have been mined. The South Alligator Valley Uranium Field (SAVUF) is about 230 km SE of Darwin (Figure 1), where thirteen small uranium mines with a combined production of less than 1000 tonnes of U_3O_8 operated in the period 1955 to 1964. The deposits of both Fields occur in Palaeoproterozoic metasediments of the Pine Creek orogenic domain and are unconformity related. Recent geochemical studies (eg, Mernagh et al, 1994; Komninou and Sverjensky, 1996) indicate that uranium for both Fields was transported in low temperature, saline, oxidised fluids above the unconformity and deposited in structural and reducing traps in the Palaeoproterozoic basement.

Regional Geology

Jabiru Area (West ARUF)

A comprehensive description of the geology of the ARUF is given in Needham (1988). The oldest rocks of the ARUF comprise the Archaean granitic gneisses of the Nanambu Complex (2470 Ma) (Page et al., 1980), which forms the basement in the west (Jabiru area) of the Field. The Nanambu basement is overlain by an upward fining sequence, with the arkosic and arenitic Kakadu Group metasediments at the base, in turn overlain by the Cahill Formation and Nourlangie Schist (Figure 2). The lower part of the Cahill Formation, which includes calcareous rocks, carbonaceous schist, amphibolite and schist, hosts the three major deposits of the Jabiru region - Ranger, Jabiluka and Koongarra.

West Arnhem Land (East ARUF)

The east of the ARUF, in West Arnhem Land, is underlain by granitic and tonalitic migmatitic gneisses of the Nimbuwah Complex, which have an "I-type" pedigree and crystallisation ages of 1866 Ma (Page et al.,1989). Nevertheless, field relationships suggest that the Nimbuwah Complex could contain older elements as it forms the basement for the overlying Kakadu Group and Cahill Formation equivalents, which were metamorphosed in the 1870 to 1800 Ma Top End Orogeny. Past workers (eg, Needham, 1988) have placed the meta-equivalents of the Cahill Formation and Nourlangie Schists in the Myra Falls Metamorphics. Recent work by exploration companies in West Arnhem Land, however, has enabled an informal sub-division of the metasediments into four units: the calcsilicate unit (base), the lower arkosic unit, the amphibolitic unit (Zamu Dolerite) and the upper arkosic unit (top). The calcsilicate unit is considered equivalent to the calcareous unit of the Lower Cahill Formation in the west ARUF (Figure 2). The Nabarlek uranium deposit occurs in the amphibolitic unit, and the upper arkosic unit is interpreted as equivalent to the Upper Cahill Formation.

The metamorphic sequences of West Arnhem Land have been intruded by several post-orogenic granites, for which Page et al. (1980) obtained minimum intrusion ages of 1750 and 1780 Ma. The very extensive, generally flat-lying gabbroic intrusion, the Oenpelli Dolerite, intruded the Palaeoproterozoic rocks of the ARUF at around 1688 Ma (Page et al., 1980). Recent field observations have revealed that the Oenpelli Dolerite has also intruded the Kombolgie sandstone cover sequence.

The Kombolgie Formation is a widespread, several thousand metre thick sequence of late Palaeoproterozoic sandstones which unconformably overly the older metasediments. A thin basic

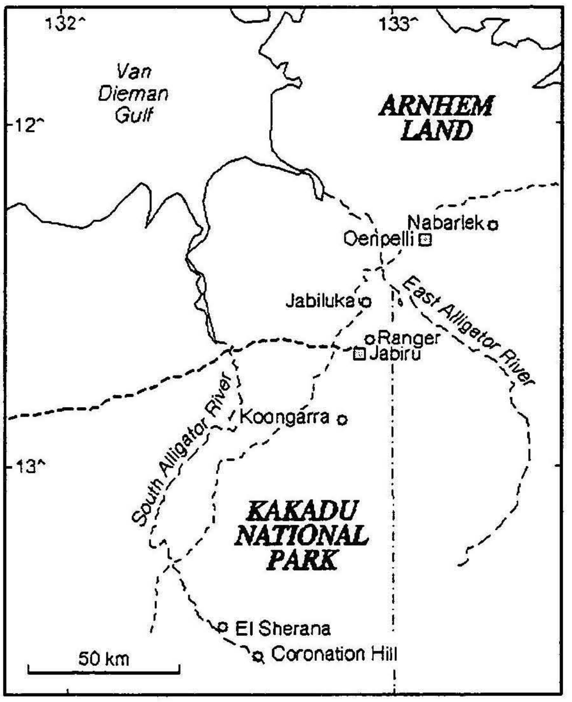

Figure 1. Main uranium deposits of the South and East Alligator Rivers region.

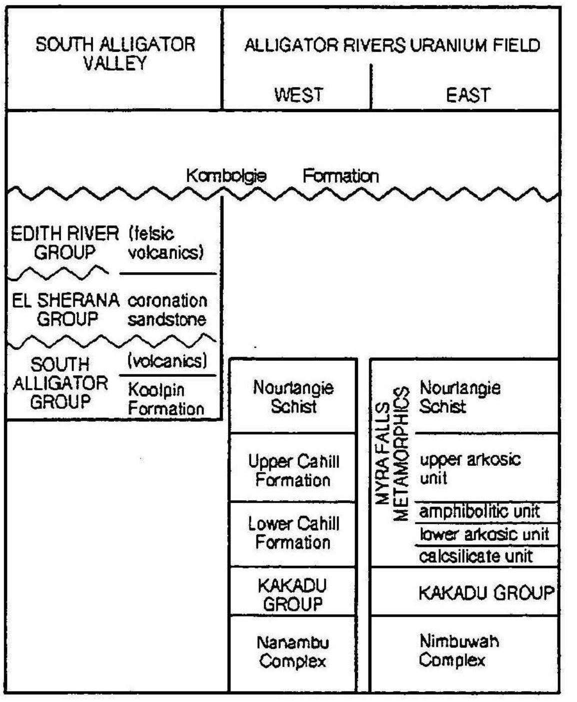

Figure 2. Schematic Precambrian stratigraphy of the South and East Alligator Rivers region.

volcanic member divides the sandstones into two units in West Arnhem Land.

South Alligator Valley

In the SAVUF, the Palaeoproterozoic basement sequence is part of the South Alligator Group, with the Koolpin Formation of that Group, consisting of pyritic, carbonaceous, pelitic sediments with bands of chert, being the main host of most of the uranium mineralisation (Carville et al., 1990). This unit is higher in the sequence than the Cahill Formation, perhaps at an equivalent stratigraphic position to the Nourlangie Schists (Needham and De Ross, 1990) (Figure 2). The Koolpin Formation is unconformably overlain by sub-aerial felsic volcanics and clastic sediments of the El Sherana Group, a late orogenic rift sequence formed about 1860 to 1850 Ma ago (Needham and De Ross, 1990). The El Sherana Group in turn is unconformably overlain by the Kombolgie Formation. Most of the gold and PGE mineralisation in the South Alligator Valley occurs within the El Sherana Group although some is associated with uranium mineralisation in the Koolpin Formation.

Uranium Deposits

Ranger

The Ranger 1 orebody has been exhausted and mining has commenced in the Ranger 3 orebody. The original estimated reserve for Ranger 1 was 52,000 tonne U_3O_8 (Kendall, 1990): Ranger 3 is reported to contain 19.9 Mt ore at 0.28% U_3O_8 - some 55,000 tonne of U_3O_8 (Register of Australian Mining, 1997/98). For both deposits, the mineralisation occurs as uraninite in a quartz-feldspar-biotite schist and microgneiss unit with some graphitic schist horizons. The unit directly overlies a thick unit of dolomite/magnesite carbonates at the base of the Lower Cahill Formation. Intense brecciation, low angle thrusting and chloritisation are reported from both orebodies (Kendall, 1990). Age dating by Maas (1989) suggested mineralisation ages from 1600 to c.1750 Ma, with a preference for 1737 Ma.

Jabiluka

There are two adjacent orebodies at Jabiluka, about 500 m apart. Jabiluka 1 contains 1.3 Mt of ore at 0.25% U_3O_8. Jabiluka II contains 52 Mt of ore at 0.39% U_3O_8, over 200,000 tonnes of contained U_3O_8 (Hancock et al., 1990). In more recent reports, reserves are given as 19.532 Mt at 0.46% U_3O_8, about 85,000 tonne of contained U_3O_8 (Register of Australian Mining, 1997/98). The uraninite mineralisation is hosted in quartz - sericite - chlorite - graphite schists of the Lower Cahill Formation, and is localised in E-W trending breccia zones (Hancock et al., 1990). Chlorite - sericite alteration is extensive around the deposits, and is recognisable several hundred metres above the unconformity (Gustafson and Curtis, 1983). A significant gold orebody is associated with Jabiluka II. Maas (1989) has reported a Sm-Nd age for primary mineralisation of 1614 Ma for Jabiluka II.

Koongarra

The Koongarra mineralisation occurs as uraninite in garnet - quartz - mica schist of the Lower Cahill Formation, with graphitic schist present in the hanging wall (Snelling, 1990). There are two orebodies, both being strongly controlled by a major NE trending reverse fault zone, which forms their footwall. The fault has a vertical displacement of at least 200 m with the Cahill Formation being thrown over the Kombolgie Formation (Snelling, 1990). Alteration at Koongarra extends up to 1.5 km from the orebody, perpendicular to the fault. Alteration in the outer zone is represented by chlorite replacement of biotite: in the inner zone (50 m from the ore), quartz and garnet are also replaced by magnesian chlorite and phengitic mica (Snelling, 1990). The Koongarra 1 orebody is reported to contain 14,500 tonne of U_3O_8 in ore of average grade 0.79% U_3O_8, as well as traces of gold. The estimated resource of Koongarra 2, which is some 100m along strike from Koongarra 1, is 2000 tonne U_3O_8 in ore grading about 0.3% U_3O_8 (Register of Australian Mining, 1997/98). Maas (1989) suggests a Sm-Nd age of between 1550 and 1650 Ma for the primary mineralisation at Koongarra.

Nabarlek

The Nabarlek orebody occurred as uraninite mineralisation in the amphibolitic unit of the Myra Falls Metamorphics. It had the form of a flattened pipe in the hanging wall of a NW-striking fault zone. Hydrothermal alteration extends over 1 km from the orebody and consists of an outer zone, which has an assemblage of iron-rich chlorite and white mica together with some silicification, and an inner zone dominated by magnesian chlorite, phengitic white mica and locally hematite (Wilde and Noakes, 1990). Production from Nabarlek totaled 10,860 tonnes U_3O_8 from ore which averaged 1.95% U_3O_8 (Queensland Mines Pty Ltd internal report). Production was completed in 1989 and the mine site has been rehabilitated. The Sm-Nd age obtained by Maas (1989) for primary mineralisation at Nabarlek is 1616 Ma.

South Alligator Valley Deposits

The uranium deposits of the SAVUF were all very small: past production ranged from 226 tonne U_3O_8 (El Sherana) to a few tonne (eg Scinto 6) (Crick et al., 1980). Most of the mineralisation is within the major NW striking fault system of the Valley. The host rocks are mainly ferruginous, carbonaceous and cherty shale and siltstone of the Koolpin Formation. The mineralisation typically occurs as pitchblende, associated with faults and shearing, and close to the unconformity with overlying coarse clastic sediments of the El Sherana Group (Crick et al., 1980). There are several phases of chlorite and hematite alteration in the South Alligator Valley: chlorite alteration and some desilicification are particularly associated with uranium mineralisation (Mernagh et al., 1994). Dating by Greenhalgh and Jeffery (1959) of mineralisation from the SAVUF suggests ages ranging from 600 to 900 Ma.

Genesis of uranium mineralisation

Early theories on the genesis of the ARUF deposits postulated the uranium to be sourced from the Archaean basement or the Palaeoproterozoic metasediments. More recent publications founded on fluid inclusion and isotopic studies suggest, however, that the source lay in the overlying Kombolgie Formation sandstones and interbedded volcanics, that the uranium was transported in saline brines and thereafter precipitated by reductants or other changes in the physicochemical environment (eg, Wilde et al., 1989; Mass, 1989).

Geochemical modeling of uranium deposit formation by Komninou and Sverjensky (1996) supports transport of uranium as uranyl complexes in Na-Ca-Cl basinal brines with high oxygen fugacities, low pH and a temperature of about 200°C. Komninou and Sverjensky also concluded that reduced Fe in aluminosilicates (garnet, biotite, amphibole) of the host rocks was the principal reductant of the uranium in the brines and that graphite was not a prerequisite for uranium precipitation.

The genesis of South Alligator Valley ore deposits (Au, PGE and U) was investigated by means of fluid inclusion studies by Mernagh et al. (1994). They concluded that SAVUF mineralisation (both precious metal and uranium) was of a similar origin to that of the ARUF. Metals were transported in highly oxidised, acid, CaCl dominated basin brines originating from the cover sequence and uranium was precipitated at temperatures of about 140°C by ferrous ion and carbonaceous reductants in the metasedimentary host rocks.

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Geology and Mineralisation of the Palaeoproterozoic Layered Mafic-Ultramafic Intrusions in the Halls Creek Orogen, Western Australia

Dean Hoatson

Australian Geological Survey Organisation
PO Box 378, Canberra, ACT 2601, Australia
dhoatson@agso.gov.au

Palaeoproterozoic layered mafic-ultramafic intrusions in the Halls Creek Orogen (HCO) of the East Kimberley, Western Australia, have generated considerable exploration interest since the mid 1960s when Dow & Gemuts (1967) reported the first occurrence of platinum in the Panton mafic-ultramafic intrusion. The HCO is an attractive province to explore, since it contains one of the most extensive mafic-ultramafic igneous associations in Australia; only the Giles Complex in the Musgrave Block of central Australia contains a larger total area of layered mafic-ultramafic intrusions (Glikson & others, 1996). The East Kimberley layered intrusions have been divided into seven major groups on the basis of field relationships, mineralisation, and U-Pb zircon-baddeleyite geochronology. The intrusions were emplaced into the crust at depths of ~8 to 24 km during at least three major periods, namely 1855 Ma, 1845 Ma, and 1830 Ma. The emplacement of the 1855 Ma intrusions (groups I to III) is contemporaneous with granitic plutons of the Bow River Batholith and felsic volcanics of the Whitewater Volcanics (both ~1850-1860 Ma), all of which represent a major magmatic event and a huge flux of heat into the crust. The identification of 'metallogenic corridors' has helped focus exploration for different styles of magmatic and hydrothermal platinum-group elements (PGE), chromium, nickel, copper, cobalt, gold, titanium, and vanadium associated with the intrusions.

Regional setting

The East Kimberley layered mafic-ultramafic intrusions (Fig. 1) are concentrated in the central and western zones of the HCO (Tyler & others, 1995); a well-exposed north-northeasterly-trending orogenic belt ~120 km long and at least 400 km wide along the southeastern margin of the Kimberley Basin. The orogen formed initially during the Palaeoproterozoic between a postulated Kimberley Craton, underlying the Kimberley Basin to the northwest, and a composite craton involving Archaean rocks in the Pine Creek and The Granites-Tanami provinces to the east. Sedimentary and volcanic sequences in Mesoproterozoic, Neoproterozoic, and Palaeozoic basins cover much of the composite craton (Tyler & others, 1995). The Palaeoproterozoic of the HCO is characterised by sedimentation, volcanism, and widespread intrusive magmatism involving the emplacement of largely post-tectonic granitic and gabbroic igneous complexes (Page & Hancock, 1988).

Geology & mineralisation

In spite of considerable exploration activity since the late 1960s, no economic deposits associated with the layered intrusions have been found in the HCO, and there has been little effort to identify the different phases of mafic-ultramafic magmatism and mineralisation on a regional scale. Recent investigations by the Australian Geological Survey Organisation (AGSO) and the Geological Survey of Western Australia (GSWA), as part of the National Geoscience Mapping Accord (NGMA) Kimberley-Arunta project, have re-evaluated the geological setting and economic potential of the intrusions (Hoatson, 1993, 1995; Hoatson & Tyler, 1993; Hoatson & others, 1995; Page & others, 1995; Trudu & Hoatson, 1996). On the basis of their age of emplacement, contact relationships with country rocks, degree of fractionation, style and intensity of deformation, and types of mineralisation, the layered intrusions have been assigned to seven major groups (designated I to VII). Intrusions which show weak or no layering and form small irregular bodies

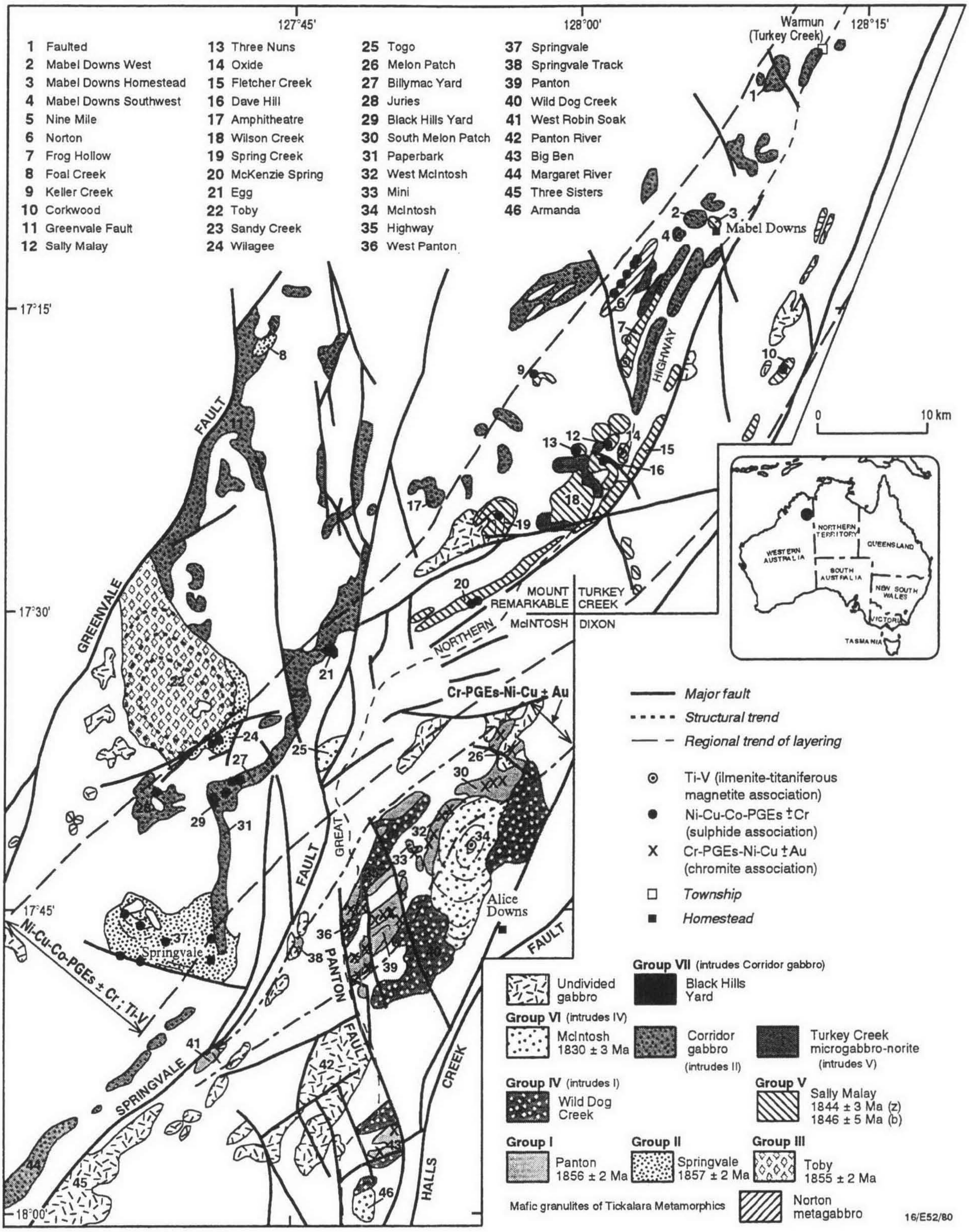

Figure 1. Regional distribution of Palaeoproterozoic layered mafic-ultramafic intrusions in the East Kimberley. The two northeast-trending 'metallogenic corridors' containing mineralised intrusions are indicated by the dashed lines.

(Corridor gabbro, Turkey Creek microgabbro-norite, Norton metagabbro) are not included in this classification (Fig. 1).

The intrusions have variable forms that range from folded sheets, through shallow dipping sheets or basins, composite multi-chambered bodies, funnel-shaped bodies, broad dyke-like bodies, to steeply plunging plugs. Metamorphic overprinting for most intrusions is from lower to upper amphibolite facies, but although the intrusions have been recrystallised to varying extents, primary igneous cumulus textures are generally well preserved. Laterally continuous mafic sills in the Tickalara Metamorphics locally form large bodies (Norton metagabbro) of granulite facies. The intrusions have thin chilled and contaminated margins, narrow contact aureoles, co-mingling net vein features at the contacts of gabbro and granite bodies, and comagmatic satellite intrusions in the country rocks indicating that they crystallised in situ, rather than being exotic blocks as proposed by previous investigators. U-Pb zircon and baddeleyite geochronology (Page & others, 1995) has shown that there was at least three major periods of emplacement, namely 1855 Ma, 1845 Ma, and 1830 Ma.

Group I mafic-ultramafic intrusions (the Panton intrusion is the type example) are coeval with groups II and III, and are the oldest (1855 Ma) layered bodies dated in the East Kimberley. Group I intrusions are strongly differentiated bodies that have the highest ratio of ultramafic to mafic cumulates. They have been deformed by tight, upright, northeasterly-trending folds to produce steeply dipping folded bodies up to 2 km thick, and are metamorphosed under lower amphibolite facies conditions. Group I intrusions are considered prospective for stratabound chromitite and/or sulphide-bearing layers enriched in PGEs, Cu, Ni, and Au near contacts between ultramafic and gabbroic zones, and for structurally controlled hydrothermal concentrations of PGEs, Cu, and Au in thick sequences of serpentinised olivine-rich cumulates. *Group II* intrusions (Springvale) are moderately dipping sheet-like mafic bodies that are coeval with, or slightly younger, than adjacent granite plutons of the Bow Granite Batholith. Minor Ni-Cu-PGE mineralisation occurs with disseminated chromite in plagioclase-rich cumulates (troctolite and anorthosite) in the upper parts of these bodies. *Group III* (Toby) is a high level evolved intrusion consisting largely of sub-ophitic gabbroic and doleritic rocks of non-cumulus origin. *Group IV* intrusions (Wild Dog Creek) form homogeneous sheet-like gabbroic bodies that intrude group I. Their high degrees of fractionation, lack of layering and cyclicity indicate that they have little potential for base- or precious-metal mineralisation. *Group V* intrusions (Sally Malay) are differentiated mafic-ultramafic bodies emplaced at 1845 Ma. They host massive and disseminated Ni-Cu-Co sulphides in embayments near the basal contacts of peridotite-troctolite-gabbro-anorthosite cycles (possible feeder conduit similar to the Voisey's Bay Ni-Cu-Co deposit in Labrador, Canada), and titaniferous magnetite layers in the upper mafic parts of the intrusions. The sulphide-rich nature and the pervasive faulting which characterise these intrusions also indicates potential for hydrothermally remobilised mineralisation within the intrusion and adjacent country rocks. *Group VI* intrusions (McIntosh) are large 1830 Ma funnel-shaped and smaller sheet-like gabbroic bodies that intruded various metamorphic rocks and group IV gabbros. The upper parts of the thicker and more fractionated intrusions of this group have potential for stratabound titaniferous magnetite layers, and remobilised Au-Pd-Cu mineralisation similar to the Platinova Reef of the Skaergaard intrusion, East Greenland (Bird & others, 1991). *Group VII* intrusions (Black Hills Yard) are steeply plunging troctolite and olivine gabbro plugs cutting biotite gabbros of the Corridor gabbro that have limited economic potential because of their small size. Mafic granulites in the Norton intrusion (Tickalara Metamorphics) host gossans along the basal contact and PGE mineralisation.

The regional distribution of the major mineralised intrusions (groups I, II, and V) defines two parallel northeast-trending corridors (Fig. 1) that can help focus exploration for Cr-PGEs-Ni-Cu±Au (chromite association) and Ni-Cu-Co-PGEs±Cr (sulphide association) mineralisation. The 'metallogenic corridors' indicate that magmatism migrated towards the northeast for these mineralised intrusions, and that PGE-sulphide associations appear to have become dominant with time relative to PGE-chromite associations.

Depths of emplacement and tectonic evolution

Geothermobarometric studies of reaction coronas between olivine and plagioclase in gabbroic rocks by Trudu & Hoatson (1996) have shown that the intrusions were emplaced into different levels of the crust at depths ranging from about 8 to 24 km (2.4 to 6.7 kb). The layered intrusions of greatest economic interest were emplaced into deeper crustal levels with time. The 1855 Ma group I intrusions (Cr-PGEs-Ni-Cu) crystallised at crustal levels of ~11 km, whereas the 1845 Ma group V intrusions (Ni-Cu-Co-PGEs) were emplaced at levels of 20 to 24 km.

The regional tectonic setting of the HCO has been the subject of considerable debate for many years. Hancock & Rutland (1984) postulated that the province developed in an intracontinental setting from initial extension where large areas of thinned Archaean crust were formed, followed by convergence. Tyler & others (1995) made analogies with modern continental margins, with the formation of felsic magmas being related to modern-style subduction processes in a magmatic arc environment, whereas Wyborn. L. (pers. comm., 1997) on the basis of the geochemistry of Kimberley granites has proposed a model involving underplating and extension. Proterozoic layered intrusions overseas, such as the Bushveld Complex and Great Dyke, are generally believed to have been emplaced in stable intracratonic environments in which rifting occurred (von Gruenewaldt & Harmer, 1992). The emplacement of the East Kimberley layered intrusions suggests multiple periods of rifting between 1855 Ma and 1830 Ma. The development of large mafic-ultramafic magma chambers (the McIntosh intrusion is 7.8 km-thick and covers an area of 84 km^2), laterally continuous rhythmic layering (over many kilometres in some cases) and internal cumulate textures also indicate that stable tectonism prevailed, at least locally, during the crystallisation of individual intrusions. Any model that attempts to explain the tectonic evolution of the HCO must take into consideration the emplacement of the Palaeoproterozoic layered mafic-ultramafic intrusions.

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The Cullen Event: A Major Felsic Magmatic Episode in the Proterozoic Pine Creek Inlier of Northern Australia

Elizabeth A. Jagodzinski and Lesley A. I. Wyborn

Australian Geological Survey Organisation
P.O. Box 378, Canberra, ACT 2601 AUSTRALIA

Based on new geochronological and geochemical data, volcanic and plutonic suites of the Cullen Suite and Jim Jim Suite, are now considered to represent a single large-scale felsic magmatic event , named the Cullen Event in the Pine Creek Inlier. The tectonic setting, geochemical signature and associated metallogenic styles of this Event are consistent with intracontinental rifting, with magmas derived from partial melting of a pre-existing lower crustal source. There are no petrographic, chemical or metallogenic similarities to magmas from subduction zones.

Field and Geochronological Data

Felsic magmas of the Cullen Event cover an area of at least 10 000 km2. Components of this event include the Cullen Batholith, the Grace Creek, Malone Creek, Jim Jim, Tin Camp and Nabarlek Granites as well as felsic volcanics of the Edith River Group, including the newly defined Gimbat Ignimbrite Member (Jagodzinski, 1992). The units overlie and intrude older basin sediments (~2100-1880 Ma) of the Pine Creek Inlier, which were affected by a major high T, low P compressional orogenic event (1870-1850 Ma). Previously considered to be synorogenic, the 1830-1820 Ma Cullen Event is now known to clearly postdate this event by ~ 40 Ma (Jagodzinski, 1992; Stuart-Smith et al., 1993), and is associated with an unrelated extensional tectonic event. Volcanics are subaerial, proximal ignimbrites, and interbed with braidplain and lacustrine sediments. They infill rift valleys which formed on the older basin sediments during reactivation of pre-existing basement structures (Jagodzinski, 1991, 1992). There is no strong preferred orientation to any of the plutons or volcanics although the intrusives are divided by a major north-northwest trending shear, the Pine Creek Shear Zone.

Geochemistry

On a regional scale, the magmatism is bimodal, consistent with the proposed extensional setting. The main geochemical characteristics differ significantly from magmas in modern subduction zones in either an island arc or continental margin environment. In particular, the granites are predominantly I-(granodioritic) type, and compared with granites associated with Phanerozoic subduction zones, they have much higher SiO2, K, Th and U values. The granites are so felsic that they cannot be derived by direct partial melting from the mantle, implying that they are derived by melting of a pre-existing continental crust. This is supported by Sm-Nd model source ages for volcanics of the Cullen Event, which range between 2000 and 2400 Ma. Multi-element, primordial-mantle-normalised abundance diagrams are Sr-depleted and Y-undepleted. This, as well as flat HREE patterns and negative Eu anomalies, indicate plagioclase stability and the absence of garnet/hornblende phases in the source region. The chemical compositions also indicate a high geothermal gradient operating throughout both the duration of the Cullen Event and derivation of the source of the magmas that are all part of this event. The multi-element primoridal-mantle normalised patterns also contrast the Sr-undepleted, Y-depleted signature typical of granites from modern subduction settings, reflect melts derived from a deeper, higher pressure garnet-rich source, such as the upper mantle region above the subducting slab in regions with lower geothermal gradients.

The Cullen Batholith has very little evidence for a restitic component, apart from some amphiboles, most minerals within the rocks are believed to be precipitated from a melt. As the magma contained little solid residue, the melt fractionated producing a suite of chemically heterogeneous plutons, with large chemical variations both within and between individual plutons. Plutons can be of three main types, zoned plutons, leucogranites and relatively homogenous adamellites. Fractionation is controlled by mineral phases such as hornblende, muscovite, apatite, zircon and allanite, and is characterised by exponentially increasing Rb, Y and U, and decreasing K/Rb, with increasing SiO2. Associated greisens, aplites and pegmatites indicate late stage magmatic fluid saturation. A wide contact aureole (up to 3 km) indicates high emplacement temperatures, with the granite introducing significant heat into the local environment.

Mineralisation

The Cullen Batholith has a close spatial association with mineralisation (Bajwah, 1994a, b; Stuart-Smith et al., 1993). The most fractionated leucogranites concentrate U, Sn and W, to form vein-hosted deposits within and near the granites. In contrast, Au and base metal deposits and occurrences occur some distance from the granite, and their distribution reflects host rock interaction with ore-bearing magmatic fluids. It is probable that vapour/brine separation occurs in the ore-bearing magmatic fluids, with Au and Cu preferentially carried in a sulphur-enriched phase, and precipitating in reductant-bearing units, while Pb and Zn are transported in chloride-enriched brines that react with carbonate hosts (Wyborn and Stuart-Smith, 1993). These styles of mineralisation provide yet another contrast with Phanerozoic skarn-type porphyry copper (± Au± Mo) systems, which are associated with subduction-related granites of markedly different chemistry and morphology.

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Crustal structure of the Fennoscandian Shield

Annakaisa Korja[1] and Toivo Korja[2]

[1]Institute of Seismology
P.O.Box 26, FI-00014
University of Helsinki, Finland

[2]Geological Survey of Finland
FI-02150, Espoo, Finland

The Fennoscandian Shield is bounded by the Caledonides in the west and by Phanerozoic sediments to the south. The Shield comprises an Archaean Domain in the east, Paleo-proterozoic Svecofennian in the central parts and Neoproterozoic Sveconorwegian in the southwest.

The crustal thickeness map (Fig. 1) indicates variations from 35 km to 65 km within the Archaean Karelian Province, the Palaeoproterozoic Svecofennian Domain and the Palaeoproterozoic Lapland-Kola Orogen. A strong correlation exists between the thickness of lower crustal high velocity layer and the crustal thickness indicating that the crustal thickness variations are due to variations in the thickness of the lower crustal high velocity layer. An important additional feature is that the upper and middle crust seem to be thinner in regions where the lower crustal high-velocity layer is thicker, so that velocity isolines (7.0 kms^{-1}) are deflected upwards in areas where the crust is thickest. In the Svecofennian domain, the thickest crust represents the Palaeoproterozoic Svecofennian orogeny but the regions of thinner crust relate rather to the Subjotnian extension. Similarly, the thinner parts of the Archaean Karelian Province may be Archaean or, more plausibly, result from Palaeoproterozoic extension, whereas conversely, the thickest parts are a product of Palaeoproterozoic Svecofennian processes (Korja et al., 1993, Korsman et al., 1997).

The cratonized Archaean crust experienced several extensional events between ca. 2.5 and 2.0 Ga that produced layered intrusions, mafic dykes and several electrically conducting volcanogenic and metasedimentary belts. A lower crustal high-velocity layer up to 8 km thick everywhere beneath the Karelian Province indicates that a mafic underplate may have formed as a result of these Palaeoproterozoic extensional events.

General results from the reflectivity data from the BABEL experiment (BABEL Working Group, 1993a, 1993b), supported by refraction seismic data (e.g. Korhonen et al., 1990; BABEL Working Group, 1993), indicate that in the regions of thinned crust the lower crust is characterized by strong subhorizontal reflectivity with sharp Moho reflections whereas in the regions of thickest crust the reflectivity decreases and becomes irregular in the lowermost crust. The deepest parts are almost transparent with occasional reflections at the Moho depth inferred from refraction data. This has been interpreted as indicating that prominent lithological layering in the deep crust becomes less pronounced in the deepest parts, consistent with a proposed mafic intraplate and underplate (Korja et al., 1993).

The isostatic compensation of the regions of thickest crust is not expressed topographically, nor are there any obvious gravity anomalies associated with the large variations in the crustal thickness. Gravity modelling of the crustal thickeness variations (Elo, 1997) indicates firstly that the thickness variations are mostly compensated within the crust by density variations and secondly that crustal roots are compensated in three plausible ways, each of which is supported by some observations: a) thinning of the upper/middle crust in the regions of thickest crust, b) the intrusion of mafic and denser material into the upper crust; c) tectonic emplacement of deeper and denser rocks to higher crustal levels. The thick Svecofennian crust has been preserved, because its density was increased by magmatic intra- and underplating during the compressional stage of the Svecofennian orogeny. The entire Svecofennian crust equilibrated soon after magmatic underplating.

In the Bothnian Sea area, the near vertical reflection data from the BABEL-7 and BABEL-1 profiles (BABEL Working Group, 1993a) have revealed two sets of cross-cutting listric faults in the crust (Korja, A. and Heikkinen, 1995, Fig.). The older of these sets of faults flattens at the mid-crustal detachment zone (35-40 km) whereas the younger set soles out at the crust-mantle boundary (48 km). Strong, sub-horizontal reflectivity in the

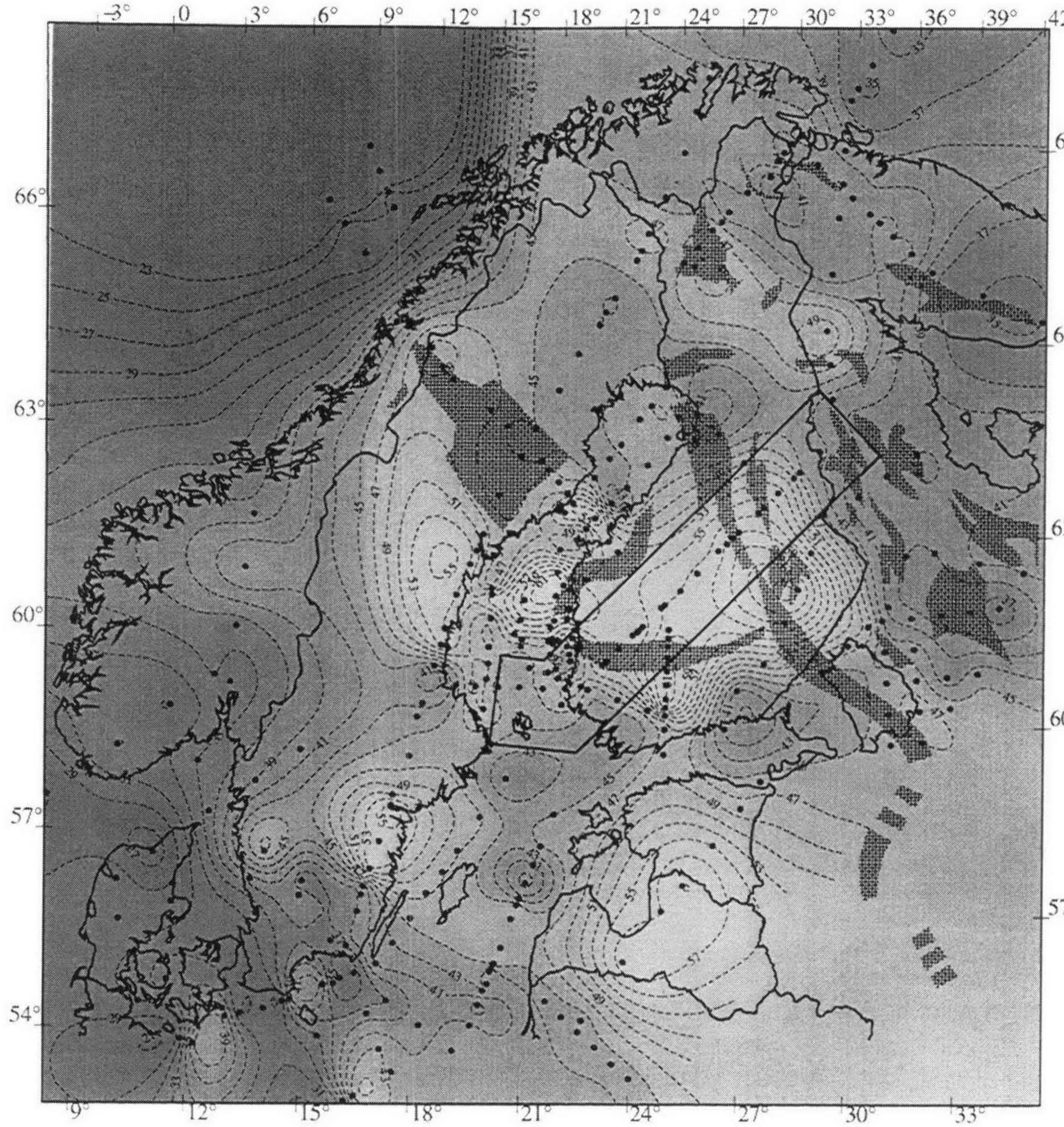

Fig. 1. Crustal thickness and crustal conductors in Fennoscandia. Contours represent depths to Moho (km) obtained from refraction seismic studies after Luosto (1991) and crustal conductors after (Korja and Hjelt, 1993). Compiled by Korsman et al. (1997).

lowermost crust is attributed to a combination of stretching and mafic sills associated with the Subjotnian diabase dykes and rapakivi magmatism. In the upper crust the rapakivi batholiths are transparent regions bounded by strong reflectors that image horst and graben structures. Nearly transparent regions, associated with high Vp beneath the Åland rapakivi batholith are attributed to coeval gabbro-anorthosites. The older set of listric shear zones has been related to extensional thinning of the crust during the collapse of the Svecofennian orogeny. The younger set of listric shear zones are parallel with major fault zones associated with rapakivi granites, anorthosites and sandstone basins. Thus the shear zones are interpreted as Subjotnian-Jotnian in age (Korja, A. and Heikkinen, 1995; Korja, A., 1995).

The rapakivi batholiths are associated with Bouguer gravity anomaly minima surrounded by anomaly maxima with a total range of 65-75 mGal. The Wiborg rapakivi granite anomaly, in south-eastern Finland has been modelled (Figs. 8 & 13 in Elo and Korja, A., 1993) as uplifted upper mantle, which produced a large Bouguer anomaly maximum around the batholith, and by low density rapakivi granites in the upper crust, which cause a minimum coinciding with the batholith. The model with uplifted upper mantle is consistent also with the refraction seismic data that indicates thinned crust beneath the Wiborg rapakivi batholith (Luosto,et al.,1990; Elo and Korja, A., 1993).

The velocity models along the GGT/SVEKA-transect were used to chose a collection of lower crustal lithologies to which thermal modelling was applied. P-wave velocities range from 7.0 kms^{-1} to 7.45 kms^{-1} in the lower crust beneath the transect. Possible rock types present therefore include anorthosites, mafic granulites, metapelites and, at the higher velocity end, pyroxenites (Holbrook et al., 1992; Rudnick and Fountain, 1995; Christensen and Mooney, 1995). Data from xenoliths indicate that the heat production of metapelitic granulite is about 10 times greater than that of mafic granulites (Rudnick and Fountain, 1995). According to thermal modelling along the GGT/SVEKA transect (Kukkonen, 1997), mafic and pelitic granulites have relative proportions of 55% and 45%, respectively, in the upper part of the lower crust in the Palaeoproterozoic Svecofennian Domain. Beneath the Archaean Karelian Province the mafic component dominates in lower crust with corresponding proportions of 75% and 25%, respectively. In the lowermost Svecofennian crust, below the depths of 40 km the mafic granulites seem to be the dominant constituent (>90%).

The location of major crustal conductors in the central and northeastern part of the shield, based on a compilation of magnetometer array, magneto-telluric and airborne electromagnetic data, is shown in Fig. 1. The close spatial coincidence of the near-surface and deep conductors suggests a genetic relationship and the deep electromagnetic data confirms the extension of near-surface structures to deeper crustal levels. Correlations between the airborne electromagnetic data for near-surface conductivity and geological data show that most of these consist essentially of graphite- and sulphide-bearing metasedimentary rocks.

The first conductive belt in the Lapland-Kola Orogen, in the northeastern part of the shield, is associated with the Palaeoproterozoic Pechenga and Imandra-Varzuga Belts (Fig. 1). These conductors are inclined and reach a depth of several kilometres, indicating that collisional tectonics was responsible for the present day geometry of conductors because a transportation mechanism is required to carry sedimentary material to the depth of tens of kilometres. The second belt of conductors extends from Northern Finland in the northeast to Russian Karelia in the southeast. This belt is comprised of shallow, sub-horizontal conductors that may represent exposed remnants of the sedimentary sequences within the Proterozoic rift basins in the Karelian Province. The conductive lithologies in both belts were deposited roughly between 2.1 and 2.0 Ga, indicating that they represent the latest Palaeoproterozoic rifting events and/or subsequent events in continental margin development (Korja and Hjelt, 1997).

The third set of conductors is detected along the boundary zone between the Archaean Karelian Province and the Svecofennian Island Arc Complex. This belt extends from the Bothnian Bay in the northeast to the Lake Ladoga region in the southeast. Further to the south the conductor gradually disappears beneath Phanerozoic sedimentary cover but can nevertheless be used to indicate the location of the Archaean-Proterozoic boundary (e.g. Korja et al., 1986; Korja and Hjelt, 1993; Korja and Koivukoski, 1994; Kovtun et al., 1997).

The fourth major conductor crosses southern Finland in a roughly east-west direction, before changing in trend towards the northwest and north in the Bothnian Sea region, from where it continues across the Bothnian region of western Finland to the Skellefteå district in Sweden, after which it crosses northern Sweden in a southeast-northwest direction, and finally disappears beneath the Caledonides (Fig. 1; Jones, 1981; Rasmussen et al., 1987; Korja, 1990). The deep conductor is associated with near-surface conductors by airborne electromagnetic surveys and with deep conductors imaged by magnetometer arrays (Pajunpää, 1986, 1987; Korja, 1990) The conductor has been modelled by one- and two-dimensional modelling techniques using data from the magnetotelluric profiles (Korja and Koivukoski, 1994, Pernu et al., 1989, Rasmussen at al., 1987). In southern Finland, the conductor, which is dipping to the north or is near vertical, coincides spatially with the southern part of the the Central Finland Continental Arc Complex. In Sweden ,the conductor dips to the north (northeast) beneath the Skellefteå volcanic belt and penetrates the whole crust to a depth of ca. 40 km.

The crust was thickened tectonically and by several magmatic under- and intraplatings. Tectonic thickening involved both under- and overthrustings as evidenced by crustal conductors as well as by reflection seismic. The presence of the thick high-velocity lower crustal layer, an abundance of seismic reflectors at the middle crust and a gradual decrease of reflectivity towards greater depths in the regions of thickest crust indicate magmatic under- and intraplating.

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Tectonic Setting of the Bergslagen Mining Region, South Central Sweden

Ingmar Lundström[1] & Rodney Allen

Geological Survey of Sweden
Box 670, S-751 28 Uppsala
Sweden

Volcanic Resources Ltd.
Morteveien 57, Hundvåg, 4085 Stavanger
Norway

The Bergslagen mining region is the westernmost and most intensely mineralized part of a Palaeoproterozoic supracrustal belt, which extends from south-central Sweden to southern Finland (Baker & Hellingwerf, Eds, 1988, Lundström & Papunen, Eds, 1986, Lundström, Stephens & Wahlgren, Eds, 1990).

Bergslagen contains hundreds of iron and base-metal sulphide deposits, but at the present time, only two Zn-Pb mines (Zinkgruvan and Garpenberg, with tonnages of c. 40 and 20 mt, respectively) still operate.

Most sulphide and iron ores occur in distal, subaqeous, rhyolitic volcanic ash-siltstones which are mostly situated in the upper part of the volcanic pile. The current interpretation is that there are both syngenetic, stratiform, ash-siltstone hosted Zn-Pb ores as well as stratabound, volcanic-associated, limestone-skarn hosted, Zn-Pb impregnation ores. The latter are commonly accompanied by extensive footwall alteration and are probably related to nearby subvolcanic intrusions. Extensive, synvolcanic Na-, K- and Mg alteration characterize large areas of Bergslagen.

Proximal, felsic pyroclastic rocks, their rapidly resedimented equivalents and subvolcanic intrusions characterize the lower part of the volcanic pile. The pyroclastic rocks are mainly pumice and glass rich ash-flow deposits erupted from large calderas. A comprehensive facies analysis is presented in Allen et al. (1996). In some places, these deposits are interbedded with mature, deltaic continent-derived sandstones, containing Palaeproterozoic to Archean (2.7-1.95 Ga, Claesson et al., 1993, Kumpulainen et al., 1996) detrital zircons. Evidence for an older, felsic, continental basement thus exists in the lower part of the volcanic stratigraphy. Although this basement does not outcrop, its existence is corroborated by geochemical and isotope data (eg. Valbracht et al., 1994).

The explosive character, abundant accretionary lapilli and many depositional features suggest that the volcanic rocks were initially erupted and deposited in shallow water to subaerial conditions. With time, the depositional basin subsided, probably due to an extensional tectonic régime. Thus, deep-water environments became successively more frequent and distal, frequently planar-bedded, volcanic ash-siltstone terminates the volcanic part of the stratigraphy. The supracrustal sequence is completed by the deposition of planar bedded, turbiditic mudstones. The volcanic rocks yield U-Pb (zircon) ages around 1880-1900 Ma. The tectonic setting of the area is interpreted as a subsiding continental platform or margin, possibly in a continental back arc setting.

Differentiated, I-type, granitoids (mostly 1870 to 1890 Ma) intruded the supracrustal sequence which was later folded and metamorphosed by low pressure metamorphism under low- to high-grade metamorphic conditions. This orogenic event affected rocks dated at about 1850-1840 Ma and was the local expression of the Svecokarelian orogeny. It was accompanied and followed by numerous 1840 to 1750 Ma old, S- and A-type granites, some of which are associated with tungsten and some sulphide mineralizations.

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Lithostratigraphy, geochemistry and tectonic setting of the Paleoproterozoic Kiruna Greenstone Group in northern Sweden

Olof Martinsson

Division of Applied Geology
Luleå University of Technology
S-971 87 Luleå, Sweden.

In Northern Norrbotten a Paleoproterozoic succession of volcanic and sedimentary rocks rests unconformably upon a 2.7-2.8 Ga Archean basement. This c. 10 km thick pile includes the Kovo Group, the Kiruna Greenstone Group, the Kurravaara Conglomerate, the Kiruna Porphyries and the Hauki Quartzite. The upper contact of the Kovo Group, the Kiruna Greenstone Group and the Kiruna Porphyries are characterised by minor unconformities and pebbles from these units are found in basal conglomerates in overlying units.

The Kiruna Greenstone Group contains economically important Cu-Au deposits, and it belongs to a large tholeiitic province extending over the northern and eastern parts of the Baltic Shield. Massive, amygdaloidal and pillowed basaltic lava is a dominant constituent in the Kiruna Greenstone Group, with komatiites, volcaniclastic rocks, clastic and chemical sediments occurring as minor but characteristic components. The greenstones were formed during a continental rift event which ended up with NW-trending ocean floor spreading. Conglomerate intercalated with WPB-type basalt and possibly evaporitic sediments in the stratigraphically lowest part of the Kiruna Greenstone Group marks the onset of the rifting event. A high degree of mantle melting, triggered by a rising mantle plume, is reflected by the succeeding komatiites. Later, MORB-type magmatism was generated by decompressional mantle melting due to strong crustal attenuation in response to the development of a NNE-directed failed rift arm.

Introduction

The Baltic Shield consist of an Archean nucleus with accreted and successively younger Prot-erozoic terrains towards the southwest (Gaál and Gorbatchev, 1987). In the northeastern part of the Shield, Paleoproterozoic greenstones are widely distributed and they contain economically significant Cu-Au deposits. From the extensive greenstone areas in northern Finland and Norway minor branches extend into the Norrbotten county in northernmost Sweden, with the well preserved Kiruna greenstones occurring in the westernmost part. Due to a remarkable similarity of the stratigraphy in the greenstone units from this region and the mainly tholeiitic character of the volcanic rocks, Pharaoh (1985) suggested them to be coeval and representing a major tholeiitic province. Based on petrological and chemical studies of the mafic volcanic rocks and associated sediments, a continental rift setting is favoured for the greenstones (e.g. Lehtonen et al., 1985; Pharaoh et al., 1987; Huhma et al. 1990; Olesen and Sandstad, 1993).

General geology

In northern Sweden a Paleoproterozoic succession of greenstones, porphyries and clastic sediments, rests unconformally upon a deformed 2.7-2.8 Ga old Archean basement. Stratigraphically lowest is the Kovo Group. It includes a basal conglomerate, tholeiitic lava, calcalkaline mafic to intermediate volcanic rocks and volcaniclastic sediments. The following Kiruna Greenstone Group is dominated by mafic to ultramafic volcanic rocks, and it is overlain by the Kurravaara Conglomerate, the Kiruna Porphyries and the Hauki Quartzite. The upper contacts of the Kovo Group, the Kiruna Greenstone Group and the Kiruna Porphyries are characterised by minor unconformities and pebbles from these units are found in basal conglomerates in overlying units. An albite diabase, intruding the lower part of the Kovo Group, is dated to 2.18 Ga (Skiöld 1986), and gives a minimum depositional age for this unit. The Kovo Group is suggested to be of Sumi-Sariolan age (2.5-2.3 Ga), while the Kiruna Greenstone Group is supposed to be of Jatulian age (2.2-2.0 Ga). The upper age limit is set by a mineral Nd/Sm-isochron age of 1.93 Ga for a mafic sill in the middle part of the greenstone succession (Skiöld & Cliff 1984).

The Kurravaara conglomerate is dominated by pebbles of porphyritic intermediate volcanic rocks (Frietsch, 1979). Up to 400 m thick arenitic beds, and locally volcanic rocks, occur intercalated in the conglomerate. The overlying porphyries in the Kiruna area are divided into two groups: the Porphyrite Group and the Kiruna Porphyries (Offerberg, 1967). The stratigraphically lower Porphyrite Group, consists mainly of porphyritic basalt and andesite with a calcalkaline affinity, and it may be stratigraphically equivalent to the Kurravaara Conglomerate. Chemically these rocks are different to the tholeiitic or slightly alkaline and partly bimodal Kiruna Porphyries volcanics (Martinsson and Perdahl, 1994), which have U-Pb zircon ages of 1.88-1.89 Ga (Welin 1987; Cliff et al., 1990). The Hauki Group, overlying the Kiruna Porphyries, consists mainly of quartzite, while conglomerate, graywacke and phyllite are less abundant (Frietsch, 1979).

The Paleoproterozoic pile of volcanic and sedimentary rocks were deformed and metamorphosed during the Svecofennian orogeny (1.9-1.8 Ga). Synorogenic 1.89-1.87 Ga intrusions of the Haparanda and Perthite-monzonite suites range in composition from gabbro to granite. Minimum melt granites and pegmatites, represented by the Lina suite, formed at c. 1.79 Ga (Skiöld et al., 1988).

The Kiruna Greenstone Group

At Kiruna a 2-4 km thick pile of mainly mafic volcanic rocks occur. These greenstones are the host rock to the Viscaria and Pahtohavare sulphide deposits, and they are currently attracting a number of exploration companies. A dominant component is basaltic lava, occurring as amygdaloidal flows in the lower part of the greenstone pile and pillow lava in the upper part. Less important are komatiites, tholeiitic tuff and andesitic to dacitic tuffaceous rocks. Minor constituents are black schist and carbonate rocks, occurring in the volcaniclastic units. These rocks have been defined as the Kiruna Greenstone Group. Based on petrographical and geochemical criteria this group was divided into six formations, which are from the stratigraphically lowest: Såkevaratjah Formation, Ädnamvare Formation, Pikse Formation, Viscaria Formation, Peuravaara Formation and Linkaluoppal Formation. The internal lithostratigraphy of the Kiruna Greenstone Group is consistent within an area of at least 70×60 km.

Conglomerates intercalated with amygdaloidal basalt dominates the lower part of the Såkevaratjah Formation in the central Kiruna area. Picrite, hematite stained arenite and a silicified carbonate rock occur as minor intercalations. Conglomerate pebbles are mainly derived from the underlying Kovo Group. Rapid lateral facies changes are recorded by the variation from coarse grained sediments of local derivation to thick dolomite units outside the central Kiruna area. This suggests the existence of small fault controlled basins, with deposition of sedimentary breccias and conglomerates adjacent to fault scarps, while precipitation of carbonates was restricted to low energy areas in the central part of the basins. A strong enrichment of Cl and Br in the basalt and conglomerate, and locally occurring silicified carbonate rocks with a nodular structure are suggested to be preserved records of a saline environment and the existence of former evaporites (Martinsson et al., 1997).

The Såkevaratjah Formation is overlain by peridotitic to basaltic komatiites, which constitutes the Ädnamvare Formation. The dominance for peridotitic komatiites with 27-30 % MgO corresponds to the maximum MgO content of liquid komatiites, and require a high degree of mantle melting (Nisbet et al., 1993). The overlying Pikse Formation consists of a thick pile of basaltic lava. Eruption of terrestrial to shallow water amygdaloidal lava flows developed a lava plain of regular thickness over a large area.

Volcaniclastic rocks with associated chemical and carbonaceous sediments in the overlying Viscaria Formation record a change in depositional environment, from earlier terrestrial to shallow water, to prevailing subaqueous conditions. A regional consistency of the detailed internal stratigraphy suggests stable conditions and the existence of a large basin. During this stage the volcanism was exclusively pyroclastic and partly bimodal. Within this mainly basaltic unit a 2-5 m thick dacitic ash tuff is recognised for a distance of more than 20 km in the upper part, and the lower part is dominated by andesitic tuff-tuffite. The basaltic volcanism was mainly of Surtseyan type which generated subaquatic pyroclastic massflows. Close to the top of the volcaniclastic pile the local occurrence of possible stromatolites and desiccation cracks indicate a temporary subaerial event.

The Viscaria Formation was succeeded by voluminous eruptions of pillowlavas in the overlying Peuravaara Formation. The rare occurrence of pyroclastic products and the small size of amygdules in the pillowlavas suggest eruptions at a water depth in the range of c. 500-2000 m (Dimroth et al., 1985; Cas, 1992). This demonstrata a rapid basin subsidence prior to eruption of the pillow lavas. A return to pyroclastic volcanism is recorded by the volcaniclastic Linkaluoppal Formation, which suggests basin shoaling. This uppermost unit of the Kiruna Greenstone Group is partly missing in the central Kiruna area due to erosion, and the Peuravaara Formation is unconformally overlaid by the Kurravaara Conglomerate (Martinsson, 1997).

Geochemistry of the Volcanic Rocks

Except for the komattites in the Ädnamvare Formation the volcanic rocks are mainly tholeiitic in character. The volcaniclastic Viscaria and Linkaluoppal formations also include minor calcalkaline andesitic to dacitic intercalations and two flows of picrite are found in the Såkevaratjah Formation. Tholeiitic flows of WPB-type are restricted to the lowest part of the greenstone pile. They are followed by basalt of LKT-type in the upper part of the Såkevaratjah Formation. The komatiites are succeeded by basalt of LKT and MORB character occurring as alternating flow units. Minor intercalation of intermediate to felsic tuff-tuffite in a dominantly basaltic volcaniclastic unit, gives the Viscaria Formation a slightly bimodale character. The basaltic tuff is chemically similar to the overlying pillowlava from the Peuravaara Formation. Both these units are slightly enriched in HFS-elements, resulting in a character transitional to E-MORB. The volcaniclastic Linkaluoppal Formation is dominantly of WPB-type. Minor tuff-tuffite of LKT and calcalkaline character are found in the lower part of this unit.

Tectonic Setting

The Paleoproterozoic Kiruna Greenstone Group is part of a volcano-sedimentary pile deposited upon an Archean granitoid-gneiss basement, demonstrating a continental setting. Characteristic for the greenstones is the mainly tholeiitic composition of the volcanic rocks, the occurrence of vesiculated terrestrial flows in the lower part and subaquatic pillowed flows in the upper part. The petrography and the chemistry of the Kiruna Greenstone Group and the temporal evolution of the volcanic rocks are suggestive of a rift setting. In particular, the association of WPB-basalt, coarse grained clastic sediments, and possibly evaporites in the lower part of the pile is indicative. Subsequent generation of MORB-type magma suggests strong crustal attenuation and decompressional mantle melting (Bailey, 1983). The associated andesitic to dacitic volcanism in the Viscaria Formation may have originated by crustal melting due to a high geothermal gradient. This evolution from early amygdaloidal WPB-basalt to late MORB-type pillow lava monitors the magmatic evolution in many flood basalt provinces (Marsh, 1987), and is consistent with the compositional changes reflected in volcanism on a lithospheric plate moving over an active mantle plume (Wyille, 1988). Eruption of komatiites and the thick pile of tholeiites may imply a generation close to the plume axis (Campbell et al., 1989). The Paleoproterozoic greenstones in Finland and northern Norway are similar in character to the Kiruna Greenstone Group, and existing age determinations suggests a main magmatic and rifting event at c. 2.1 Ga. A successful rifting of the Archean craton, along a line in a NW-direction from Ladoga to Lofoten, was accompanied by injection of NW-trending dike swarms (Vuollo, 1994), and the eruption of N-MORB pillow lava along the rift margin. The Kiruna greenstones, and dike swarms north of Kiruna, outlines a NNE-trending magmatic belt extending into northern Norway. This belt is almost perpendicular to the major rift, and may represent a failed rift arm caused by a triple junction south of Kiruna (Martinsson, 1997). The rapid basin subsidence, accompanied by eruption of MORB-type pillowlava in the Peuravaara Formation are suggested to be expressions of the development of this rift arm.

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The Northern Norrbotten Ore Province

Olof Martinsson

Division of Applied Geology
Luleå University of Technology
S-971 87 Luleå, Sweden.

The Northern Norrbotten ore province is a major producer of Cu and Fe ore in Sweden. Economically important deposits include apatite iron ores, stratiform Cu ores and epigenetic Cu-Au ores. Most mineral deposits in this area are hosted by Paleoproterozoic greenstones and porphyries. Stratiform sulphide deposits with Cu-(Zn-Pb) and uneconomic iron formations are found in volcaniclastic units of the greenstones, while apatite iron ores are restricted to the porphyries. Epigenetic sulphide deposits with Cu-Au-(Co) are found in both greenstones and porphyries and they are generally spatially related to shear zones. Scapolite, albite and K-feldspar are characteristic alteration minerals in most epigenetic sulphide deposits, while phyllosilicates may have formed extensive footwall alteration zones at stratiform deposits. Fluid inclusion and mineralogical data suggest a formation from highly saline ore fluids for most sulphide deposits in northern Norrbotten. The suggested existence of evaporitic sediments at the base of the greenstones may be metallogenetically important for this ore province by contributing to the formation of saline hydrothermal fluids for both syngenetic and epigenetic sulphide deposits.

Introduction

Northern Norrbotten is an important mining province, which is dominated by Fe- and Cu-deposits. Au is a minor constituent in some of the Cu-deposits, and metals found in subeconomic amounts include Zn, Pb, V, Ti and Co. Economically most important for the region are the apatite iron ores with an annual production of 31 Mt of ore from two operating mines, and a total production of about 1600 M ton from 10 mines over the last 100 years. Copper has been produced during the 17th and 18th centuries from several small deposits. During the years 1983-1997 copper ore was mined on a larger scale in the Kiruna area. From two mines about 15 Mt of ore with 2-3 % Cu have been produced. Sweden's largest sulphide mine, Aitik in the Gällivare area, produces annually 17 Mt of ore by open pit mining.

Geological setting

The Precambrian bedrock in the northern Norrbotten region include a c. 2.8 Ga Archean granitoid-gneiss basement, which is unconformally overlaid by greenstone, porphyry and sedimentary successions of Paleoproterozoic age. Stratigraphically lowest is the Kovo Group, comprising basal conglomerate, tholeiitic lava, calcalkaline mafic to intermediate volcanic rocks and volcaniclastic sediments. The following Kiruna Greenstone Group is dominated by mafic to ultramafic volcanic rocks, and they are overlain by the Kurravaara Conglomerate, the Kiruna Porphyries and the Hauki Quartzite. The Kovo Group is suggested to be of Sumi-Sariolan age (2.5-2.3 Ga), while the Kiruna Greenstone Group was related to a Jatulian (2.2-2.0 Ga) rifting event, and, more specifically, the development of a NNE-directed failed rift arm (Martinsson, 1997). Subduction related calcalkaline andesites from the Porphyrite Group are suggested to be of Svecofennian age and stratigraphically equivalent to the Kurravaara Conglomerate. The Kiruna Porphyries are partly bimodal and may have an intraplate setting (Martinsson and Perdahl, 1994).

The c. 10 km thick pile of volcanic and sedimentary rocks were deformed and metamorphosed during the Svecofennian orogeny (1.9-1.8 Ga). Synorogenic intrusion of the 1.89-1.87 Ga Haparanda and Perthite-monzonite suites range in composition from gabbro to granite. Minimum melt granites and pegmatites, represented by the Lina suite, formed at c. 1.79 Ga (Skiöld et al., 1988). Albitization and scapolitization of the Paleoproterozoic rocks is widespread, and locally intense, in northern Norrbotten. These alterations have been suggested to be related to felsic intrusions (Ödman 1957), or to be an expression of mobilised evaporites from the greenstone successions during metamorphosism (Frietsch et al., 1997).

Greenstone and porphyry hosted deposits

Most mineral deposits in northern Norrbotten are hosted by Palaeoproterozoic greenstones and porphyries. Stratiform-stratabound deposits of exhalative origin are confined to volcaniclastic units of the greenstones. They contain Cu-Zn-Pb and Fe in different proportions. Epigenetic deposits are mostly Cu-dominated, and they are found in both greenstone and porphyry environments. The apatite-iron ores are almost exclusively associated with the Kiruna Porphyries. The suggested existence of evaporitic sediments in the stratigraphically lowest part of the Kiruna Greenstone Group (Martinsson, 1997), may be a feature of great importance for the metallogeny of this area (Frietsch et al., 1997), as they may have contributed to the generation of highly saline hydrothermal fluids.

Stratiform-Stratabound Greenstone-Hosted Deposits

Two types of uneconomic iron deposits are found in the tuffitic upper parts of the greenstones, skarn-hosted iron formations and chert-banded iron formations (BIF). The skarn iron formations occur as stratiform lenses of magnetite, serpentine, diopside, actinolite and smaller amounts of Fe-sulphides. A typical size is 5-80 Mt with 30-40 % Fe and 1-3 % S. The BIF-deposits are mainly restricted to the eastern part of northern Norrbotten. They are dominated by silicate facies and have a thickness of up to 200 m. Generally the iron content is in the range of 15-25 %, slightly higher grade is encountered in locally developed oxide facies. The syngenetic sulphide deposits contain chalcopyrite, pyrrhotite, pyrite, sphalerite, galena and magnetite in varying proportions. The ore minerals occur disseminated or as massive intercalations in tuffite, black schist and carbonate rocks.

The largest, and the only economic, syngenetic sulphide deposit is the Viscaria Cu-ore at Kiruna. Since 1982, 12.54 Mt of ore with 2.29 % Cu has been produced. The ore is hosted by the Palaeoproterozoic Kiruna Greenstone Group, and it was formed in conjunction with basin subsidence and the onset of voluminous eruptions of MORB-type pillow lava. Ore grades of copper are mainly confined to the economically important A-zone, which is the uppermost of three stratiform ore horizons. Magnetite is a major constituent in all mineralised horizons, and the amount of sulphides increases upwards in the stacked ore sequence. A large alteration zone in the footwall to the A-zone becomes gradually narrower until it end c. 450 m stratigraphically below the ore. It is characterised by a high K/Na-ratio due to destruction of plagioclase and the formation of phyllosilicates.

The A-zone ore is generally 2-10 m thick and it has a length of 3700 m. Magnetite, chalcopyrite, pyrrhotite and some sphalerite occur disseminated or as massive ore in a calcite gangue. Lamination in mm-scale, with alternating layers of sulphide and magnetite, is common, and slumping structures in laminated ore are locally encountered. Fluid inclusions indicate a depositional temperature of c. 210 °C and a salinity of 30 eq. wt. % NaCl. These highly saline fluids are suggested to have formed by dissolution of evaporites at the base of the greenstone pile during seawater circulation. The exhaled ore fluid accumulated into a brine pool from which the ore constituents were precipitated during cooling and brine mixing with overlying seawater. The ore exhibits a pronounced Cu-Zn zoning with high grade Cu-ore confined to the fissure controlled exhalative vent. Compared to world wide Cu-Zn deposits, most features of the Viscaria Cu-ore are similar to those of the Besshi type. Regarding tectonic setting and ore genesis, Viscaria may represent an ancient analogue to the Atlantis II Deep metalliferous sediment in the Red Sea (Martinsson, 1997).

Huornaisenvuoma is a subeconomic Zn-Pb-Ag deposit of stratiform character in northeastern Norrbotten, with an estimated tonnage of 0.56 Mt with 4.6 % Zn, 1.7 % Pb and 12 ppm Ag. It is hosted by dolomite in the uppermost part of the greenstone pile. Magnetite, sphalerite, galena, pyrite, and locally chalcopyrite, occur enriched in layers within a mineralised sequence at the top of the dolomite. Calcsilicates are developed in the mineralised zone and in a restricted area in the footwall below the central part of the deposit (Martinsson, 1994).

Epigenetic Sulphide Deposits

Epigenetic Cu-Au deposits, hosted by Paleoproterozoic greenstones, are economically important in the northern part of the Baltic Shield. Most of these ores are characterised by an association to albitic rocks, shear zones and they have formed from saline hydrothermal fluids. The Pahtohavare Cu-Au deposits is the only economic deposit of this type in northern Sweden. It has produced 1.7 Mt of ore with 1.9 % Cu and 0.9 ppm Au, by underground and open pit mining during the years 1989-1997. In character it is similar to the Bidjovagge Au-Cu deposit in northern Norway (Bjørlykke et al. 1987).

The Pahtohavare deposit consists of two ore lenses, which are hosted by the volcaniclastic Viscaria Formation in the middle part of the Kiruna Greenstone Group. The ore lenses are situated in an open antiform adjacent to a shear zone, and they are in detail closely related to altered black schist, demonstrating both a structural and lithological control of the mineralization. The host rock to the ore is fine grained albite felsite, formed from black schist by albitization and graphite decomposition. Chalcopyrite and pyrite occur disseminated, in veinlets and in breccia matrix together with carbonates. Native gold occurs mainly as inclusions in chalcopyrite, which is reflected in the positive correlation between Cu and Au in the ores. A weak enrichment of U generally accompanies the sulphides. Extensive, and generally barren, biotite-scapolite alteration zones surrounds the ore-bearing albite felsites. Marialitic scapolite has characteristically developed porphyroblasts, veinlets and networks in a biotite rich matrix. Coarse grained ferro-dolomite and calcite veins are abundant in the Pahtohavare area. They may contain high grade ore, but are rarely mineralised outside the ore-related alteration zones (Martinsson, 1997).

Fluid inclusion data demonstrates a formation from highly saline solutions at a temperature of 350-500°C (Lindblom et al., 1996). Ore precipitation and host rock alteration was associated with brittle to ductile deformation. The introduction of hot and highly saline solutions in strongly fractured rocks caused early biotite-scapolite alteration and scapolite veining, which was followed by albitization and decomposition of graphite in black schists. However, carbon isotope data from carbonates in altered black schists indicate only minor contribution of isotopically light carbon of graphite origin. Pb isotopes of ore-related scapolite is similar to Svecofennian ore lead (Romer et al., 1996), suggesting a Svecofennian age of the ore. Alteration of black schist was an important process for ore precipitation by decrasing fO_2, which lowered the solubility of Cu and Au.

The porphyry-hosted sulphide deposits are generally Cu-dominated, and typical ore-related alterations include K-feldspar and scapolite. Magnetite is a common minor to major component in most of them, and bornite may be an important Cu-mineral. Co-rich pyrite is an important ore mineral in some deposits. The Aitik Cu-Au deposit is of major economic importance with an annual production of 17 Mt of ore with 0.38 % Cu and 0.2 ppm Au (Abrahamsson, 1997). Chalcopyrite and pyrite occur disseminated and in veinlets together with minor amounts of magnetite, barite, and apatite. Biotite, sericite, garnet, epidote, amphibole, tourmaline and scapolite are characteristic components in the strongly altered host rock, which mainly consists of garnetiferous quartz-biotite schist and sericite schist (Zweifel, 1976; Monro, 1988). The deposit is situated within a major NW-directed shear zone and both the foot- and hangingwall contacts are tectonically controlled. The footwall is intruded by a 1.88 Ga monzodiorite and 1.74 Ga pegmatites are cutting the ore and the hangingwall (Monro, 1988; Witschard, 1996). Sulphur isotope data from sulphides and barite indicate a temperature of formation around 500°C, and together with strontium isotope data a magmatic hydrothermal origin for the Aitik deposit is suggested (Yngström et al., 1986).

Apatite-Iron Ores

The genesis of the apatite iron ores have been a subject for discussion the last 100 years. However, the magmatic origin suggested for the Kiirunavaara deposit (Geijer, 1910), is favoured by most later workers. An U-Pb titanite age from a magnetite-titanite dike in the footwall to the Luossavaara deposit suggests the ore to be c. 1.89 Ga old (Romer et al., 1994). This is in accordance with crosscutting granitic dikes at Kiirunavaara which gives a minimum age of 1.88 Ga for the ore formation (Cliff et al., 1990). From the two producing mines, Kiirunavaara and Malmberget, c. 31 Mt of ore is produced annually. The grade of Fe varies from 55-68 % in the economic parts of the deposits, and the content of P is generally less than 1 %.

Based on morphology and mineral composition the apatite-iron ores in the Kiruna area may be divided in two major types, 1) breccia ore and 2) stratiform-stratabound lenses (Martinsson, 1994). Deposits of breccia type generally occur in a stratigraphically low position within the Kiruna Porphyries, or in the underlying Porphyrite Group. The breccia deposits consist of magnetite and varying amounts of actinolite, while apatite is rare. Ore related alterations are not prominent, but may include scapolite. The stratiform-stratabound lenses are confined to stratigraphically high positions within the Kiruna Porphyries. Hematite is a minor to major constituent, and they are rich in apatite. Actinolite is missing, carbonate and quartz together with apatite are the main gangue minerals. Sericite and tourmaline are locally common as alteration minerals in the host rocks. Generally the apatite iron ores are poor in sulphides. However, Cu-mineralizations are spatially related to some of them, with Gruvberget and Tjårrojåkka as the most important examples.

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Field Geological and Sm-Nd Isotopic Evidence for Delineating the Southwestern Margin of the Archaean Craton in Northern Sweden

Claes Mellqvist

Division of Applied Geology
Luleå University of Technology
SE-971 87 Luleå, SWEDEN
clme@sb.luth.se

On a regional scale, Sm-Nd isotopes have been used to delineate the Archaean-Proterozoic palaeo-boundary in northern Sweden. More detailed work is being performed in the Luleå and Jokkmokk areas. New results from c. 1.9 Ga plutonic rocks in the area close to Luleå show a distinct shift in the Sm-Nd characteristics, from strongly negative ε_{Nd} in the north to positive values in the south. At this shift Archaean rocks have also been found which are interpreted as representing the location of the southwestern-most edge of the Archaean craton. A tectono-magmatic model is made by using Sm-Nd data from c.1.8 Ga and c. 1.9 Ga granitoids. This model shows a possible scenario for the formation of the Archaean-Proterozoic palaeoboundary.

Introduction

Except for 2.1 to 2.3 Ga old mafic, rift-related volcanics, no igneous rocks with an age between 2.67 Ga and 1.95 Ga are known in northern Sweden (Skiöld et al., 1993). Sm-Nd isotopes can therefore be used to trace the influence of Archaean rocks in the source material of Proterozoic igneous rocks. The Archaean-Proterozoic palaeoboundary in northern Sweden was tentatively delineated on a regional scale by using Sm-Nd isotopes (Öhlander et al., 1993). A transition zone extending from the Luleå area at the coast of the Bothnian Bay in a WNW direction, separates the Archaean craton in the northeast from Proterozoic juvenile areas in the southwest. The Archaean rocks are generally covered by Proterozoic rocks, or have been reworked during the Svecofennian orogeny. When delineating the Archaean palaeoboundary, Öhlander et al.(1993) used Sm-Nd isotopes on c. 1.9 Ga old granitoids and metavolcanics and c. 1.8 Ga old granitoids, although the most frequent sample type was the 1.9 Ga old granitoids. The hypothesis was put forward that the Archaean palaeoboundary was formed when the juvenile volcanic-arc terrain to the south was thrust onto the Archaean continent. Similar models for the border of the Archaean craton in Finland have been presented, e.g. Lahtinen(1994), Lahtinen & Huhma (1997).

In order to obtain a more precise delineation of the border of the old craton at the present level of exposure, detailed studies have been performed in the Luleå and Jokkmokk areas (Fig.1). Since comparison of results based on different rock types is difficult, the strategy has been to use rocks of similar type and age occurring on both sides of the boundary. In this case this is c. 1.9 Ga old plutonic rocks of granodioritic to tonalitic composition.

New results from the Luleå area

The latest Sm-Nd data on the c. 1.9 Ga old plutonic rocks show that the boundary zone of the Archaean craton is isotopically very distinct near the town of Luleå. Here, plutons with an isotopic memory of old continental crust are separate from contemporaneous magmas with a juvenile mantle signature.

Rocks of Archaean age have recently been found in an area close to the town of Luleå (Lundqvist et al., 1996 and Wikström et al., 1996). These rocks were first identified in the Vallen-Alhamn area, approximately 30 km south of Luleå. Rocks with petrographical, chemical and Sm-Nd isotopical similarities were later found at localities further north. The locations of Archaean exposure coincide well geographically with the zone outlined by the abrupt change in Nd isotopic signatures of the Proterozoic plutons. This zone is believed to be identified on the aeromagnetic map

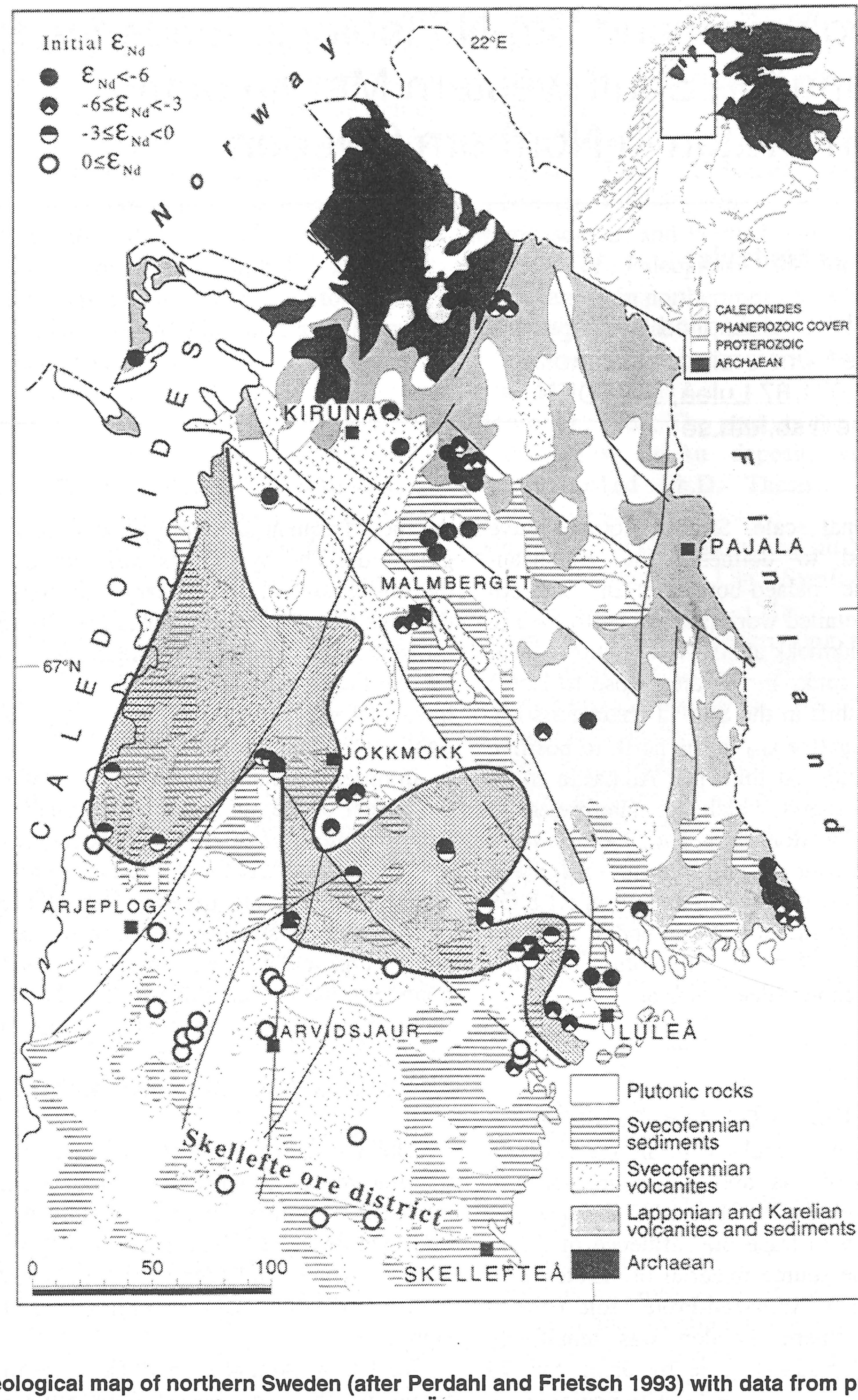

Fig. 1. Geological map of northern Sweden (after Perdahl and Frietsch 1993) with data from previous Sm-Nd isotopic work and the Luleå-Jokkmokk zone of Öhlander et al. (1993). Major fracture zones drawn from Öhlander and Nisca (1985).

as well. The Archaean exposures are characterized by low magnetism and border a high magnetic anomaly to the southwest which is mainly due to mafic to intermediate subvolcanic rocks. The Archaean rocks are of two types; one is an even-grained, foliated, sometimes gneissic plutonic rock, ranging in composition from granodiorite to diorite, and the other is a foliated, porphyritic granodiorite. Contact relationships between the Archaean and the Proterozoic rocks have been studied at a few localities. The results show that the Archaean exposures occur as megaxenoliths, some of them covering areas of a few square kilometres, with tectonic contacts towards the younger rocks. They commonly have an internal deformation pattern which is cut by later shear-zones. The contacts towards the surroundings are also heavily disturbed by later deformation.

Various types of smaller Archaean xenoliths have also been identified. One of these small xenoliths, approximately 10×20 m in size and consisting of porphyritic granodiorite, occurs at the Måttsundsberget hill. Here, it is surrounded by foliated 1.9 Ga old subvolcanic to volcanic rocks of intermediate to mafic compositions with identical deformation patterns. Another Archaean xenolith was observed at the Bälingsberget hill where the surrounding host rock has a plutonic character and is only slightly deformed. This xenolith is smaller (approximately 20 cm in diameter) with a rounded shape and internal foliation at an angle with the foliation of the host. Rounded xenoliths of Archaean age together with xenoliths of Proterozoic age are also found in some magmatic breccias in the Luleå vicinity. Thus the question arises whether the presence of the described remnants of Archaean rocks reflects the presence of the southwestern-most edge of the Archaean craton in the Luleå area. Such an inference is supported by the fact that no Archaean rocks have been found further to the southwest.

Tectono-magmatic modelling

A comparison between the Sm-Nd isotope characteristics of c. 1.9 Ga old granitoids and c. 1.8 Ga granitoids occurring in the same area shows that strongly negative initial ε_{Nd} values occur farther southwest in the potassic 1.8 Ga granitoids than in the calc-alkaline 1.9 Ga granitoids. Intrusions of the younger granitoids having distinctly negative initial ε_{Nd} values are surrounded by granitoids of the older group with positive values. The interpretation is that a juvenile volcanic-arc terrain to the south was forced and thrust onto the Archaean continent after the formation of the 1.9 Ga granitoids. Consequently juvenile rock masses were deposited upon the reworked Archaean continent. The 1.8 Ga granitoids were formed by remobilisation of continental crust. Partial melting at 1.8 Ga resulted in the intrusion of granitods carrying the Sm-Nd isotope signature of the Archaean continent into juvenile rocks. It is probable that the collision discussed here was part of a major accretion to the Archaean craton after the formation of the calc-alkaline 1.9 Ga granitoids, but before the formation of the 1.8 Ga granitoids.

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Some Key Geochronological Constraints for Palaeoproterozoic Crustal Evolution in Northern Australia

R.W. Page

Australian Geodynamics Cooperative Research Centre
Australian Geological Survey Organisation
P.O. Box 378, Canberra A.C.T. 2601 AUSTRALIA
rpage@agso.gov.au

The comparison of tectonic processes in pre-1800 Ma rocks between the Palaeoproterozoic provinces in northern Australia and Fennoscandia, a prime objective of the present workshop, is to a large extent dependent on the reliability of the geological age framework.

Both the accuracy and precision of geochronological input into tectonic models have improved considerably in the past several years, due both to better instrumentation and to improved geological insights/interpretations from strategically applied geochronological research. In Australia, the main advances have come from application of SHRIMP studies which allow in situ U-Pb dating of zircon and other high uranium minerals. This approach has allowed us to become far more confident of the temporal framework of Palaeoproterozoic crustal development

This review highlights some of the critical geochronological constraints now available for the pre-1800 Ma Palaeoproterozoic terrains of northern Australia.

Late-Archaean Crustal Elements in Northern Australia

Prior to discussing the geochronology of the earliest recognised Palaeoproterozoic depositional sequences in northern Australia, it is useful to summarise the isotopic evidence relevant to the late-Archaean basement.

There are only a few areas in the North Australian Craton where late-Archaean basement is seen to underlie Palaeoproterozoic sequences. The two best known examples which have been dated at 2500 to 2600 Ma are the Rum Jungle and Nanambu Complexes, both in the Pine Creek Inlier (Page,1988). Recently, two small late-Archaean granitic gneiss terrains having SHRIMP U-Pb ages of 2514±3 Ma, 2510±22 Ma, and 2504±4 Ma (Billabong complex and Browns Range Dome) have been identified in The Granites-Tanami region (Page et al., 1995). Crust of this general age is considered to form the basement underlying the supracrustal sequences of the Tanami Complex.

The existence of Archaean basement beneath and between most of the Palaeoproterozoic rocks of the north Australian craton is now less conjectural. In addition to the direct dating evidence outlined above, there is mounting indirect evidence (from Nd isotopic model ages, and U-Pb ages of inherited zircon components in igneous rocks and detrital zircon components in sedimentary rocks) suggesting the involvement of late-Archaean crust in the formation of other Palaeoproterozoic terrains.

New Sm-Nd studies have led McDonald et al. (1996) to suggest that there is an Archaean tonalite-trondhjemite-granite suite within the Kalkadoon-Leichhardt Belt of the Mount Isa Inlier. This could have important implications for interpreting the crustal evolution in the Mount Isa Inlier. However, Page and Sun (in press) indicate that although this unit has abundant Palaeoproterozoic and Archaean inheritance, it has a zircon U-Pb crystallisation age of ~1860 Ma, and may therefore represent Palaeoproterozoic rocks that have been recycled to a large degree from as yet undiscovered Archaean crust.

Metasediments in the Browns Range Dome region of the Tanami province include arkose with detrital zircons ranging in age from 2470 Ma to 3460 Ma, with several clear age groupings. The youngest of these groupings has an age of 2507±22 Ma, representing a maximum depositional age for the arkose sequence. Groupings at 3050±30 Ma, 3140±10 Ma, ~3270 Ma, and ~3410 Ma mimic ages for inherited zircon in the nearby gneiss and leucogranite (Page et al., 1995). Detrital zircons with similar ages, and some dated at 3600 Ma, are also present in sandstones of the Saunders Creek Formation (with Nd TDM model age of 3460 Ma), at the base of the Palaeoproterozoic Halls Creek Group in the East Kimberley. These zircon data reinforce the view that rocks as old as early-mid Archaean might be components of the unexposed lower crust of northern Australia.

Pre-1900 Ma Magmatism

The apparent dearth of depositional or magmatic activity between 2500 and 2000 Ma is broken with the advent of bimodal igneous activity in the interval 1920-1905 Ma. So far, this event has only been recognised in the Halls Creek Orogen of Western Australia, although there is circumstantial evidence for this event in detrital and xenocrystic zircon suites in the Tennant Creek Inlier (Compston, 1995). The dated felsic units in the Halls Creek Orogen occur in the core of antiformal domes, and range from metarhyolite (Ding Dong Downs Volcanics and equivalents 1907±6 Ma, 1912±3 Ma) to high-level granitoids (Sophie Downs Granite 1912±5 Ma; Junda Microgranite 1913±5 Ma). Relatively early and volumetrically minor magmatic events of comparable age are reported in the Svecofennian of northern Sweden (e.g. Norvijaur tonalite 1926±12 Ma; Skiöld et al. 1993).

1880-1850 Ma magmatism and metamorphism

All of the Palaeoproterozoic terrains of northern Australia were affected by a major crust-forming event in the interval 1880-1850 Ma. This was termed the Barramundi orogeny and considered to be largely ensialic in character by Etheridge et al. (1987). Best estimates for the age(s) of this event are derived from zircon SHRIMP ages of 1890±8 Ma in the Yaringa Metamorphics (Mount Isa) and 1885-1870 Ma in the Pine Creek region (Page and Williams, 1988). In the eastern Pine Creek Inlier, migmatitic granitoids and granulitic rocks in the Nimbuwah Complex have U-Pb zircon ages of 1866±8 Ma and 1886±5 Ma (Page et al., 1980). However, the Barramundi event(s) is generally not well dated, and in fact may be diachronous across the various Palaeoproterozoic terrains of northern Australia (e.g. the Hooper Orogen in the East Kimberley occurred at ~1850 Ma). The Barramundi orogeny is a useful term for grouping the widespread magmatic and metamorphic rocks of this age, although the orogeny is clearly more complex than originally envisaged, and may in part involve subduction-related tectonics (e.g. Tyler et al., 1995; Zhao and Bennett, 1995; Myers etal., 1996).

The best geochronological control on the earliest supracrustal rocks of this age comes from dating of the South Alligator Group in the Pine Creek Inlier, which contains carbonaceous and iron-rich sediments that interfinger with felsic volcanics (Gerowie Tuff and Mount Bonnie Formation) dated at 1884±3 Ma and 1877±11 Ma (Needham et al., 1988). This part of the Pine Creek Inlier sequence is comparable in age to bimodal volcanism and clastic and calcareous deposition within the Biscay Formation, in which a firm age of 1880±3 Ma has been established for this lower part of the Halls Creek Group in the East Kimberley. Volcanic units interspersed with younger turbidites the Halls Creek Group are also closely age-constrained, with SHRIMP U-Pb magmatic ages of 1857±5 Ma, and 1848±3 Ma suggesting a long-lived volcanic evolution (Page, Blake & Tyler, in prep.). The older (~1860 Ma) volcanic and depositional ages in the Halls Creek Group closely overlap the range of ages reported by Compston (1995) for pre-Barramundi orogeny turbiditic sequences (Warramunga Formation) in the Tennant Creek region.

Voluminous granitoids were also produced in this time interval, both in the King Leopold and Halls Creek Orogens of the Kimberley (1850-1865 Ma), and in the Tennant Creek, Pine Creek, and Mount Isa Inliers. Numerous other small basement inliers across northern Australia (e.g. under the Mesoproterozoic McArthur Basin) consist of magmatic products that can also be identified as having the same age as these major Barramundi orogeny suites (Wyborn et al., 1987).

The complexity of the events that must have contributed to this orogenic episode(s) is better discerned from recent more precise dating work in the Kimberley (Page et al., 1995; Page and Hoatson, in prep.). This has established two major thermal events associated with regional metamorphism and emplacement of granitoids and gabbros. These events are at ~1850 Ma and ~1830 Ma in the East Kimberley (Halls Creek Orogen). In the West Kimberley (King Leopold Orogen), contemporaneous magmatism and high-grade metamorphism may be slightly older, as all but one of six dated granitoid plutons have coherent crystallisation ages between 1858±5 Ma and 1864±4 Ma, close to the age of metamorphism in nearby migmatitic gneisses which formed at 1861±5 Ma.

1840-1820 Ma magmatism

SHRIMP U-Pb zircon results provide moderately good geochronological control on the syn- to post-orogenic felsic magmatism at this time in the Pine Creek and Tennant Creek regions. Rocks dated in the Pine Creek region include felsic volcanics in the Edith River Group (1825±4 Ma), granitoids in the Litchfield Complex (1840±5 Ma), Jim Jim Granite (1838±7 Ma), and rocks of the Cullen Batholith, within which various plutons have been dated at between 1820-1835 Ma (Stuart-Smith et al., 1993).

Fractionated leucogranites in the Cullen Batholith are considered to be associated with Au mineralisation, and the age agreement of zircons dated from a quartz-sulphide vein at 1817±16 Ma is consistent with this model (Compston and Matthai, 1994). Micas associated with and considered to be related to the Tennant Creek Au deposits have 40Ar/39Ar ages in the same range (1825-1830 Ma). In the Halls Creek Orogen, several younger granitoids of the Bow River Batholith were also emplaced at this time, and have U-Pb zircon ages ranging from 1832±3 Ma (Mabel Downs Granodiorite) to 1821±4 Ma (Sally Downs Tonalite). This Batholith, which also includes younger undeformed granitoids with ages between 1810-1790 Ma, does not appear to be associated with any significant mineralisation.

This post- or late-Barramundi orogeny magmatism may be comparable in tectonic terms to the post-Svecofennian Revsund-type granitoids (1780-1800 Ma), collectively referred to as the Transscandinavian Igneous Belt (Skiöld et al., 1993).

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Introduction to the Finnish Precambrian with Special Emphasis on the Palaeoproterozoic Svecofennian Orogeny

Matti Pajunen*, Toivo Korja, Kalevi Korsman, Petri Virransalo and the GGT/SVEKA Working Group

*Geological Survey of Finland
P.O. Box 96, FIN-02151 Espoo, Finland
matti.pajunen@gsf.fi

The Finnish Precambrian, in the central part of the Fennoscandian Shield, records an informative cross-section of Precambrian history from Mesoarchaean (>3.1 Ga) to Mesoproterozoic (ca. 1.2 Ga) (Simonen 1980). Eastern and northern Finland are dominated by the Archaean Karelian Province, a continental terrain formed by a complex history between >3.1 and 2.65 Ga. In the west the Karelian Province borders the Palaeoproterozoic Svecofennian Domain, an arc complex which evolved between 1.93 and 1.80 Ga. Since Hietanen's (1975) plate tectonic scheme several evolutionary models for the Svecofennides have been presented (discussed in Lahtinen 1994). In northern Finland a continent-continent collision at ca. 1.9-1.87 Ga between the Archaean Inari Complex and the Karelian-Lapponian terrain resulted in the Lapland Granulite Belt which represents an uplifted, laminated lower crustal sequence of khondalites and enderbites (ca. 2.0-1.9 Ga old) (Korja et al. 1996). The Karelian-Lapponian terrain in central Lapland includes rifting-related Palaeoproterozoic Lapponian greenstone belt formations and sedimentary sequences. Central Lapland was later reactivated during an extensional event at about 1.8 Ga ago resulting in the Central Lapland Granitoid Complex.

This introduction focuses on the evolution of Svecofennides in southern Finland by presenting the results of the Global Geoscience Transect programme, GGT/SVEKA (Korsman et al. 1997). The 840 km long transect area extends from the eastern Archaean continent to the Mesoproterozoic rapakivi area in the southwest representing cross-section of the bedrock in both space and time. The programme was carried out by researchers from several organizations in Finland and was set up to link geological and geophysical information on the crust and lithosphere with a view to constructing a tectono-evolutionary crustal model. Tectonic, metamorphic and magmatic history dated by isotopic methods was interpreted and fixed with the aid of geophysical features. According to refraction seismic data (SVEKA) the crustal thickness varies considerably, reaching exceptionally high values, up to 65 km, in southern Finland. It is about 45 km in the easternmost part of the Archaean Karelian Province and in the rapakivi areas. How this exceptionally thick crust was formed and how it was preserved without collapse is an interesting question. One of the aims of the project was to explain the significance of the crustal conductors shown by the magnetotelluric resistivity data. Gravity and magnetic data were used to establish upper crustal structures as well as to explain anomalies in the middle and lower crust.

Archaean Karelian Province in eastern Finland

The eastern Archaean Karelian Province records a multistage development. It is composed of trondhjemitic-tonalitic-granodioritic granitoids and migmatites (TTG) and of greenstone belts (Luukkonen 1992). The earliest crustal remnants are in excess of 3.1 Ga in age (Paavola 1986). A widespread dynamothermal anatexis and amphibolite facies metamorphism occurred at about 2.84 Ga. The greenstone belts consist of mafic metavolcanic rocks associated with minor ultramafic, komatiitic and acid volcanites. Magmatism took place at about 2.79-2.7 Ga. Metasedimentary sequences include meta-arkoses, mica and black schists and quartz-banded iron ores. Subsequent compression deformed the greenstone belts into narrow, isoclinally folded zones. The early-generated crust was overthrusted and intruded by massive tonalite-granodiorite magmatism at 2.7-2.65 Ga and was affected by epidote-amphibolite to amphibolite facies and even locally granulite facies metamorphism. Cratonization of the continent can be inferred from the more zonal character of the later deformations and by the more granitic character of magmatism. Termination of an Archaean orogenic cycle was commonly expressed by intrusion of anorogenic-like granite batholiths, which, indeed, took place in the Karelian Province during Palaeoproterozoic time at 2.44-2.35 Ga (Luukkonen 1992).

Palaeoproterozoic reactivation of eastern Archaean Karelian Province

Extension and rifting events in the northern end of the eastern Archaean continent initiated at 2.45 Ga with the intrusion of a number of layered mafic to ultramafic intrusions. Widely distributed, 2.2-1.97-Ga mafic dykes in the Archaean craton area show that the extensional stages were episodic and prolonged (Vuollo 1994).

Jatulian epicontinental sediments, including conglomerates, quartzites and arkoses, were deposited discordantly on the craton and were associated with mafic volcanism. The transgressive Jatulian sequence is overlain by the passive margin Lower Kalevian metaturbidites. These formations exist only east of the Archaean-Svecofennian suture zone. An extensive segment of the Palaeoproterozoic formations in the Karelian terrain, including the Outokumpu nappe, are allochthonous and were overthrust from the southwest during the Svecofennian orogeny (Koistinen 1981). Geophysical studies also confirm that these sequences are underlain by Archaean crust (Korja 1995). According to geochemical characteristics the 1.95-Ga Jormua Ophiolite Complex represents an early opening of an ocean basin (Kontinen 1987 and Peltonen et al. 1997). It however is also allochthonous, transported to its present position by Svecofennian overthrusting. According to the carbon isotopic evidence even earlier break-up of the craton at about 2.1 Ga ago is suspected (Karhu 1993).

Palaeoproterozoic metamorphism and metasomatism at 1.85 Ga in the zones cutting the Archaean craton progressively increase in grade in the western part of the craton up to amphibolite facies conditions. Pressure estimates suggest uplift and exhumation of about 15 km since 1.85 Ga. The distribution of Palaeoproterozoic overprinting by metasomatism-metamorphism indicates a widespread thermal event. Isotopic studies on Archaean granitoids and greenstone belts and contemporaneous magmatic activity in the Karelian terrain also indicate such a late thermal event. The seismic deep structure of the crust with the distribution of the high-velocity layer in the lower crust below the reactivated areas may indicate a disturbance in the lower crust and lithospheric mantle at that time. It is suggested that this disturbance is related to magmatic underplating into the lower crust after the Svecofennian-Archaean collision (see later) (Pajunen and Poutiainen 1977).

Svecofennian Domain

The Svecofennian Domain in southern Finland is divided from the Archaean Karelian Province by a strongly sheared zone. Palaeoproterozoic magmatic rocks east of the zone always show an Archaean component in their geochemical characteristics, whereas in the Svecofennian granitoids the Archaean component does not exist (Huhma 1986). The Svecofennian Domain is dominantly composed of marine sediments, metagreywackes, with minor clastictic sediments, limestones and volcanic rocks. A vast amount of syn- and post-tectonic granitoids characterizes this Domain. The main geotectonic units are the Pyhäsalmi Primitive Island Arc near the craton boundary, Central Finland Continental Arc and Southern Finland Sedimentary-Volcanic Complex.

The transitional crust of the Pyhäsalmi Primitive Island Arc (1.93-1.91 Ga) was developed over a SW dipping subducting plate and is characterized by strongly metamorphosed low-K-tholeiites, basaltic andesites and low-K-rhyolites and by comagmatic gneissic tonalites and by psammites and high-Al-turbidites.

The northern part of the Central Finland Continental Arc is composed of the Central Finland Granitoid Complex that formed of a groups of older (1.89 Ga) per- or meta-aluminious granitoids that have within-plate affinities and a group of younger granitoids that was formed by mixing of within-plate magma and ca. 2.0 Ga crustal material. Evidence for the latter exists only as a 2.0 Ga zircon population in metasediments (Lahtinen and Huhma 1997). These granitoids also cut the Pyhäsalmi Arc. In the south the Granitoid Complex is bordered by the low-grade Tampere Schist Belt with mature island arc volcanic rocks (1.904-1.889 Ga) and turbiditic sediments. Some Pb-Pb isotopic model ages of ca. 1.95 Ga for volcanic rocks underlying the island arc volcanites imply rifting of the protocrust prior to 1.91 Ga (Lahtinen and Huhma 1997). The southernmost part of the Continental Arc, a high-grade

Psammitic Migmatite Belt, is characterized by tonalitic-granodioritic migmatization, MOR-type ultramafic rocks (>1.9 Ga) and basic and ultramafic cumulates (1.89 Ga) (Peltonen 1995). The Psammitic Migmatite Belt forms an important crustal conductor that dips below the Central Finland Granitoid Complex.

The Southern Finland Sedimentary-Volcanic Complex, coinciding approximately with the Pelitic Migmatite Belt, has sharp border with the Continental Arc. It is characterized by migmatitic pelitic metaturbidites with granitic leucosomes, islan arc volcanic rocks (ca. 1.89-1.88 Ga) and an amphibolite-carbonate rock-quartzite association of unknown age. Some volcanic rocks showing MORB-characteristics evidence of an early rifting (Ehlers et al. 1986).

The latest stages of Proterozoic crust formation in southern Finland are presented by the intrusion of postorogenic rapakivi granites at about 1.67-1.54 Ga ago (Rämö 1991) and the deposition of Mesoproterozoic, Jotnian graben-type sediments (Kohonen et al. 1993), which were intruded by Postjotnian diabase dykes at 1.27-1.25 Ga ago (Simonen 1980).

Evolution of the Svecofennian orogeny

According to tectonometamorphic studies (compiled in Korsman et al. 1997) the main phase of collision between the Svecofennian Domain and the Archaean Karelian Province that caused crustal thickening was initiated already before 1.9 Ga and continued up to ca. 1.885 Ga. The effects of this collision can be found in a broad zone up to 150 km into Archaean, where the Svecofennian metamorphic overprint is identifiable.

The crust of the northern complexes, the Pyhäsalmi Primitive Island Arc and Central Finland Continental Arc, stabilized earlier than the southern part as confirmed by tectonic-metamorphic and isotopic studies (Korsman et al. 1988 and Vaasjoki and Sakko 1988). The thermal peak was achieved at 1.883 Ga in the Pyhäsalmi and Central Finland arcs, when post-tectonic orthopyroxene-bearing granitoids were intruded. In contrast within the east-west trending Pelitic Migmatite Belt continuous thermal activity produced granitic magmatism, migmatization and locally granulite facies metamorphism that lasted for a long time, from about 1.885 to 1.81 Ga. The highest heat flow was attained late with respect to collisional tectonic structures and it occurred in the transect area at a coherent pressures of about 5 kbar. The southward decrease in ages in the Svecofennian terrain is also seen in the ages of post-tectonic granitoids, 1.883 Ga near the craton boundary, 1.87 Ga in central parts and 1.8-1.78 Ga in southern Finland.

Understanding of the high Svecofennian heat flow was sought by linking the regional variations in crustal densities and thicknesses to geological processes in the lower crust. As shown by tectonic-metamorphic observations the amphibolite to granulite facies, high-T/low-P, metamorphism was connected with magmatic activity, especially in the areas of the thickest crust. The variation in crustal thickness is caused mainly by variations in the thickness of the seismic high velocity layer in the lower crust. This high velocity layer continues eastwards below the Archaean continent up to the area where the crust is of its normal thickness and the Svecofennian thermal effect is minimal. Comparison of observations of tectonic-metamorphic and magmatic evolution on the present erosion level with the geophysical structure of deeper crust suggests that the high heat flow was due to magmatic under- and intraplating into the lower crust. Density also increased in the upper crustal levels as a result of mafic magmatism. The thin crust in the south-western part of the profile is related to extensional evolution that produced graben formations in the upper crust and extensive lower crustal granitic rapakivi magmas.

Summarizing, the thick crust was formed by collision and was strongly modified by later magmatic under- and intraplating processes causing delamination in the inhomogeneous lower crust. Of course, the present Moho boundary is also partly reflecting phase transitions during the cooling in the deeper crustal levels. But why has the thick crust been preserved up to the present without collapse? The problem is explained by high densities of the crust caused by under- and intraplating-related mafic magmatism even into the upper crust, and by continuing compression caused by collision in the south between the Central Finland Continental Arc with the Southern Finland Sedimentary-Volcanic Belt. The crustal conductors associated with the Psammitic Migmatite Belt at the southern border of the Central Finland Continental Arc are interpreted as an accretionary zone between these two complexes.

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Overview of the Svecofennian Mafic Magmatism in Finland

Petri Peltonen & Hannu Huhma

Geological Survey of Finland
FIN-02150 Espoo, Finland
(petri.peltonen@gsf.fi)

The bedrock of the Fennoscandian (Baltic) Shield represents a deeply denuded remnant of Precambrian mobile belts of different ages. Finland occupies the central part of the Shield which can be divided into two major geological units: (1) the late Archaean (3.1-2.5 Ga) cratonic nucleus, comprising TTG-suite gneisses and greenstone belts, with its metamorphosed and deformed Palaeoproterozoic volcano-sedimentary cover deposit, and (2) the Palaeoproterozoic Svecofennian accretionary island arc terrane to the southwest. In this presentation I shall review the tectonic settings of major mafic-ultramafic igneous provinces and formations of the Finnish part of the Fennoscandian Shield. Emphasis is given both to the petrogenesis and metallogeny of mafic-ultramafic supracrustal and plutonic rocks formed during the prolonged rifting and break-up of the late Archaean Karelian craton between 2.45-1.95 Ga as well as those generated during the Palaeoproterozoic Svecofennian orogeny between 1.93-1.85 Ga.

The Karelian Province

The northern part of the Archaean (Karelian) craton - particularly in Finnish Lapland - records prolonged and episodic history of sedimentation, rifting and magmatism throughout the Palaeoproterozoic. The Palaeoproterozoic (2.53- ca. 2.0 Ga) Lapland greenstone belt, apparently related to NW-SE trending intra-cratonic rift, is the largest mafic-ultramafic igneous province in Finland (Sorjonen-Ward et al. 1997). A bimodal sequence of crustally contaminated komatiitic and felsic volcanics dated around 2.5 Ga unconformably overlie the Archaean basement and represent the onset of the rifting (Räsänen et al. 1989). Continued rifting of the Archaean crust resulted in the widespread emplacement of gabbro-norite layered intrusions between 2.45-2.39 Ga. These layered complexes host chromite deposits at Kemi and in Koitelainen, PGE mineralization at Penikat, Cu-PGE occurrences at Portimo layered complex and Ni-Cu-PGE deposits in Koillismaa (Alapieti 1989). The total ore reserves of the Kemi deposit, which is the most important chrome mine in Europe, are around 70 Mt with an average grade of 26% Cr_2O_3 for the open pit ores. Mining of the Kemi chromite started in 1966 and 19 Mt of ore have been mined so far, with current annual production being about 1 Mt. No economic PGE deposits have been discovered from these layered complexes. Terrigenous clastic sediments discordantly overlie layered intrusions, with further episodes of mafic magmatism recorded as sporadic lavas and sills dated at around 2.2 Ga, 2.1 Ga, and 2.05 Ga. This latter phase includes the vast - although low-grade - Keivitsa polymetallic Cu-Ni-PGE-Au deposit (Mutanen, 1997). Sedimentation in Lapland continued in a variety of depositional environments prior to the onset of the volumetrically most significant phase of volcanism in the belt between ca. 2.2-2.0 Ga ago when pyroclastic komatiites, iron-rich tholeiites and Mg-rich tholeiites were erupted. Recent studies imply that the western subarea of the Lapland greenstone belt (Kittilä) is allochthonous in nature which is compatible with the presence of serpentinites having ophiolitic characteristics close to the eastern margin of the area (Hanski 1997). The Lapland greenstone belt is distinct from the Archaean greenstone belts in eastern Finland (Kuhmo and Ilomantsi belts) and elsewhere, because of the lithofacies of the komatiites: spectacular pyroclastic (Saverikko 1985) and shallow-water outcrop structures imply depositional environment truly distinct from that of Archaean komatiites which are characterised by thick lava flows and subvolcanic cumulates. Apparently due to such unfavourable environment for the segregation and accumulation of immiscible sulphide liquid, no economic magmatic sulphide deposits have been located from the komatiitic rocks of the Lapland greenstone belt. Instead, mafic-ultramafic volcanics host several syntectonic Au deposits. One of such occurrences, the Geluk-type komatiite-hosted Pahtavaara deposit, went into production in 1996 with estimated reserves of 1.3 Mt ore with an average grade of 3.4 g/t Au (Korkiakoski & Kilpelä 1997).

In Lapland, rifting ceased after the formation of the uppermost volcanic units of the greenstone belt. At the site of the present SW margin of the Archaean craton, however, rifting continued until true oceanic crust formed, as recorded by a sequence of 1.97 Ga tholeiitic dykes, 1.96 Ga old peralkaline granites and 1.95 Ga fragments of Red Sea -type oceanic crust and upper mantle (Jormua and Outokumpu Ophiolites; Peltonen et al. 1996). Major oceanic basin developed west of the newly formed passive margin to became closed some 50 Ma later due to the continental collision between juvenile Svecofennian arc terrane and Archaean craton. At this stage fragments of oceanic crust and upper mantle became thrust onto the craton. Within the Outokumpu nappe, altered and metamorphosed upper mantle fragments (now serpentinites) are associated by the well-known Cu-Co-Zn-Au ores of the Outokumpu-type (Peltonen & Kontinen, this volume).

The Svecofennian Province

The Svecofennian province represents predominantly juvenile crust generated in a rapid succession of igneous activity, uplift, erosion and redeposition between 1.93-1.85 Ga (Huhma 1986). Numerous plate tectonic models have been presented to explain the lithological and metallogenic zonation of the Fennoscandian Shield and the Svecofennian domain in particular (e.g. Hietanen 1975, Gaál 1990, Lahtinen 1994, Nironen in press.). The Svecofennian domain consists of at least three arc complexes of different ages. The oldest one, ca. 1.92-1.93 old Pyhäsalmi island arc complex adjacent to the craton margin comprises a bimodal volcanic sequence of basalts, andesites and rhyolites. This volcanic suite hosts several important VHMS deposits (e.g. Pyhäsalmi Zn-Cu-S deposit operated since 1962). Mafic members of this bimodal suite are subalkaline, low-K tholeiitic basalts or basaltic andesites, with primitive IAT affinity (Ekdahl 1993, Kousa et al. 1994). The greatest part of the Svecofennian domain in Finland, however, is slightly younger in age and several distinct, but broadly coeval, supracrustal belts can be delineated. The mafic metavolcanic rocks of the Tampere Schist Belt, for example, are geochemically similar to modern volcanic-arc rocks at continental margins or mature island arcs (Kähkönen 1989). The metavolcanic rocks of the Häme Belt, in turn, grade from older intermediate island arc-type volcanics into younger mafic ones with gradually increasing geochemical within-plate affinity (Hakkarainen 1994). The southern-most Uusimaa Belt is bimodal the mafic members being primitive tholeiites that show within-plate or MORB affinities (Lindroos & Ehlers, 1994).

Economically, the most important episode of Svecofennian mafic magmatism occurred close to the peak of the ongoing Svecofennian orogeny. At this stage, tholeiitic mantle derived melts intruded supracrustal rocks giving rise to diverse group of gabbroic, pyroxenitic and ultramafic intrusions. At uppermost crustal levels these magmas crystallised as large gabbroic arc complexes, locally showing prominent igneous layering. These complexes are synvolcanic with basaltic to andesitic volcanic rocks of mature arc affinity. No promising magmatic sulphide occurrences are known from this type of intrusion. At mid-crustal levels - as represented by the tonalite-migmatite zone of southwestern Finland - this magmatism gave rise to small ultramafic cumulate bodies which are best depicted as boudins, lenses or pipes that "float" in polydeformed paragneisses. They have been interpreted as representing middle crustal expressions of the same Svecofennian arc magmatism that resulted in the emplacement of the gabbroic arc complexes at higher crustal levels, and the formation of large ultramafic cumulate complexes in the lower crust and at the crust-mantle boundary region. It has been proposed that such ultramafic cumulate bodies represent remnants of originally more extensive synorogenic conduits for basaltic arc-type magmas (Peltonen 1995a). This particular type of ultramafic cumulate bodies host several small Ni-Cu(-PGE) mineralizations, some of which have been exploited. Mineral and sulphide composition of mineralised intrusions require that sedimentary sulphides played an important role in ore genesis: when ascending magmas encountered sedimentary formations containing abundant sulphidic black schists, they assimilated external sulphur, which led to the formation of an immiscible sulphide phase in the magma (Peltonen 1995b). Finnish production and smelting of nickel has been based on mining of such Sveofennian deposits: the Kotalahti mine produced 12 Mt ore grading 0.66% nickel and 0.26% copper from 1957-1987; the Vammala mine produced 7.6 Mt ore grading 0.67% nickel and 0.42% copper between 1974-1994; and the Enonkoski mine produced 6.7 Mt ore grading 0.78% nickel and 0.21% copper from 1984-1994. The only currently operating nickel mine is Hitura which has produced over 8 Mt ore grading 0.55% Ni and 0.20% Cu since 1965.

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The Outokumpu Association

[1]Petri Peltonen & [2]Asko Kontinen

[1]Geological Survey of Finland
FIN-02150 Espoo, Finland
(petri.peltonen@gsf.fi)

[2]Geological Survey of Finland
FIN-70211 Kuopio, Finland

The Outokumpu mining district in eastern Finland has become famous for its Precambrian massive sulfide ore deposits with economic grades of Cu, Zn, Co, Au, and Ag (e.g. Gaál & Parkkinen, 1993). The sulphides are intimately associated with a suite of rocks comprising serpentinites (including minor pyroxenites and gabbros), dolomite rocks, quartz rocks, calc-silicate rocks (skarns), and black schists. In order to denote this rock suite, Gaál et al. (1975) introduced the term Outokumpu Assemblage, and believed that the carbonate and quartz rocks represent chemogenic sediments which originally deposited above the serpentinite and which in turn were overlain by black schists. The Outokumpu Assemblage can be traced as a complexly deformed, discontinuous, sinuous ribbon of lenses, sheets and pods for nearly 250 km in the Karelian metasediments comprising the Outokumpu nappe complex overthrusted onto the Archaean continental margin from the west (Wegmann 1928, Väyrynen 1939). Considering the entire belt as a single stratigraphic horizon and allowing for crustal shortening, one can reconstruct the paleoenvironment of the Outokumpu Association as a 300-400 km long, NW-trending rift zone (Gaál, 1990).

The Origin of the Serpentites

Koistinen (1981) was the first to propose that serpentinites of the Outokumpu Association represent fragments of ancient ophiolites. Soon after, this view was strengthened by the description of the Jormua mafic-ultramafic com-plex in the northwestern extension of the same belt as a well-preserved early Proterozoic ophiolite, accompanied by equal U-Pb zircon ages of the Outokumpu and Jormua gabbros (Kontinen, 1987). Recent work has, however, shown that Jormua Ophiolite is a "composite body" formed in a tectonic setting related to continental break-up and contains distinct fragments of both Proterozoic oceanic crust and suboceanic mantle, and - most likely Archaean - subcontinental lithospheric mantle (Peltonen et al., 1996, in press). Thus, some ophiolite blocks in Jormua are like oceanic crust formed at slow-speading oceanic ridges, i.e. they are composed of upper mantle peridotites intruded by gabbroic pods, spectacular sheeted dykes complex and overlying pillow lavas, massive lavas and hyaloclastic tuffite. In contrast, one large block is truly distinct from the others: harzburgites and chromitite pods are absent and peridotites consist mainly of only slightly or moderately depleted lherzolites which are extensively intruded by clinopyroxene + amphibole ± garnet cumulate-textured dykes derived from OIB-type alkalic melts. Importantly, too, this block is not spatially associated with any oceanic crustal units but is bounded by slices of country rocks, mainly of Archaean basement. Such evidence led Peltonen et al. (in press) to suggest that this block represents fragment of the subcontinental lithospheric mantle, and that the Jormua Ophiolite thus represent an almost contiguous sequence across an ancient ocean-continent transition, recording the accretion of a mantle diapir to the Archaean subcontinental lithospheric mantle during the formation of an incipient oceanic basin ca. 1950 Ma ago (Peltonen et al., in press). This finding provides an additional perspective to interpret the origin of the Outokumpu serpentinites. Because sheeted dykes are not known to occur in any of the Outokumpu-type bodies and lavas are only a minor component of the rock assemblage, being completely unknown from the major serpentinite (and ore) occurrences such as Outokumpu and Vuonos, it is well possible that the protolith of most of the Outokumpu serpentinites actually represent upper mantle fragments exposed in the zone of extensive crustal thinning without any associated volcanic activity. Younger analogues for such bodies, known as "orogenic lherzolites", are common in zones of extensive lithosphere thinning such as Ivrea (Italy), the French Pyrenees, and southern Spain and Morocco (Ronda, Beni Boussera ultramafic massifs). The major difference between "orogenic lherzolites" and ophiolitic peridotites is that the former represent old subcontinental lithospheric mantle, while the latter represent fragments of uppermost oceanic mantle diapir or lowermost ultramafic cumulates of the oceanic crust.

The ores

The Outokumpu Association host eleven massive Cu-Zn-Co±Au±Ag deposits, three of which have been mined producing a combined total of 50 Mt ore averaging 2.8% Cu, 1% Zn and 0.2% Co. The mining history of the district spans the period 1913-1988 involving the exploitation of a few low-grade disseminated Ni-Co sulphide occurrences, too. The ore rocks at Vuonos and Outokumpu are present as thin, up to several kilometers long sheets following the carbonate- and quartz-rich marginal zones, that actually surround the serpentinite bodies. The host rock of the Cu-Zn-Co ores is usually quartz rock and structurally the ore can be divided into layered and brecciated types - apparently the latter representing sulphides mobilised during the regional polyphase deformation (Koistinen, 1981). The massive ore contains some 50% sulfide minerals the quartz being the only major gangue mineral. Mineralogically the massive ore can be divided into pyrite- and pyrrhotite-types, the large Outokumpu (Keretti) ore being of predominantly pyrite-type whereas Vuonos and Luikonlahti are predominantly of pyrrhotite-type. The main ore minerals, in addition to iron sulfides, are chalcopyrite and sphalerite and in the Outokumpu (Keretti) also cubanite. In the pyrrhotite-type ores cobalt pentlandite is the main carrier of the Co, but in pyritic ore cobalt is included in pyrite. Numerous accessory minerals have been described from Outokumpu including e.g. stannite, cobaltite, mackinawite and siegenite. The mineral assemblage pyrrhotite-pentlandite-chalcopyrite-sphalerite, in turn, characterises nickel disseminations locally present in quartz rocks and skarns between serpentinites and enclosing mica/black schists. The Outokumpu Association is also well-known because of the abundance and variety of chrome-bearing minerals: these include chrome diopside, chrome tremolite and uvarovite in the skarns, fucsite and chromite in the quartz rocks, as well as kaemmererite, tawmawite (chrome epi-dote), chrome tourmaline, and eskolaite (Cr_2O_3).

Deposit modelling

During the past decades of mining in Outokumpu, several diverse theories have been presented to explain the origin of the Outokumpu-type ores. It is apparent that these ideas - reviewed also in this presentation - reflect not only the progress made in geological observations in Outokumpu, but also the trends of general geological and ore geological thinking. These include origin from hydrothermal fluid or "ore magma" derived from the nearby Maarianvaara granite (Trüestedt, 1921) or from some unexposed magma body beneath the ore field (Vähätalo, 1953), metasomatic-metamorphic segregation of sulphur and metals from serpentinites (Mäkinen, 1921) or from the surrounding black schists (Saksela, 1957). In recent years the volcanic-exhalative model, introduced by Borchert (1954) and its variations has been in fashion (e.g. Mäkelä 1974, Peltola 1978, Gaál 1990, Loukola-Ruskeeniemi, 1992).

Much of the validity of these models depends on what is considered to be the origin of the carbonate and quartz rocks - intimate components of the Outokumpu Assemblage. Two contrasting hypothesis have been presented: (1) either the carbonate and quartz rocks represent pervasively carbonatised and silicified peridotites or serpentinites (Haapala 1936, Auclair et al. 1993), or (2) quartz rocks and dolomites represent chemogenic sedimentary rocks (e.g. cherts) deposited on ocean floor from hydrothermal solutions - a view favoured since early 1960´s (e.g. Vähätalo 1953, Peltola 1960, Huhma & Huhma, 1970). An interesting viewpoint is, however, provided by the recent study of Kontinen (in preparation). He has observed that the quartz rocks, for example, contain abundant chromite which is similar in habit and composition to that of serpentinites. Moreover, Cr and Ir abundances of the quartz rocks are equal to their mantle abundances (2000-3000 ppm and 4-5 ppb, respectively) strongly implying that the quartz rocks, instead of being metamorphosed cherts, actually represent residual rocks derived through pervasive leaching of a peridotite or serpentinite precursor. During this alteration chromite (major host of Ir) was the only primary mineral to survive. Such an origin would also explain why the serpentinite lenses of the Outokumpu district, regardless of their size, are frequently surrounded by a thin shell of dolomite and quartz rocks against black or mica schists. It is apparent that if the sedimentary origin of the carbonate and quartz rocks is discarded, the volcanic-exhalative model for the Cu-Zn-Co±Au±Ag ore is no longer credible. We are in the very beginning of preparing an up-to-date model for the ore deposition which would take into account these new observations, but we are impressed by the similarities the Outokumpu Assemblage bears with the Eastern Metals serpentinite-associated Ni-Cu-Zn-Co-Au deposit in Appalachians (Quebec, Canada). The Eastern Metals deposit provides an unique opportunity to study a deposit similar to Outokumpu, but which has not been overprinted by supergene alteration or regional polyphase deformation and metamorphism. In that particular deposit it has been excellently demonstrated that serpentinites were initially transformed into talc-carbonate schists, and then, in a progressively more oxidizing and lower temperature environment, into quartz-carbonate rocks, and finally into siliceous rocks containing over 90 et.% SiO_2 (Auclair et al., 1993) - a probable origin of the Outokumpu Association lithologies as well.

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Age and Regional Significance of Deformation Episodes in the Svecofennian Province South of Skellefte

R.W.R. Rutland*[1] , T. Skiöld[2] and R.W. Page[3]

[1]Australian Geodynamics CRC
Research School of Earth Sciences
Australian National University
Canberra, ACT 0200, Australia.

[2]Laboratory for Isotope Geology
Swedish Museum of Natural History
Box 50 007, S-104 05
Stockholm, Sweden

[3]AGCRC, Australian Geological Survey Organisation
PO Box 378, Canberra, ACT 2601, Australia

The immediate structural and stratigraphic setting of the Skellefte ore district is well established (Weihed et al 1992, Billström and Weihed, 1996, Allen et al., 1997); and many attempts have been made on the basis of geophysical and geochemical evidence to relate the Skellefte volcanism to an island arc system overlying a northward or westward dipping subduction zone, (e.g. Korja et al., 1993, Schneider et al, 1996, Perdahl and Frietsch, 1993). Weihed et al.(1992) who reviewed earlier subduction related interpretations noted, however, that the structural evolution of the regions north and south of the Skellefte district was still poorly known.

The Skellefte area lies between two quite distinct Svecofennian tectonic zones. The northern zone displays major shear zones with distinct NNW, N, and NNE trends. The sequence in the Skellefte district itself shows a relatively simple pattern of upright folding with E-W or ESE-WNW trends. The granitic belt immediately to the south has a strong foliation with similar trends. Further south, in the southern zone, the dominant trends are NE-SW or WSW-ENE. The deformation is much more complex while the metamorphic grade increases from NW to SE with 'schollen' migmatites dominating the coastal areas.

Interpretation of the relationship between these zones and the intervening Skellefte Zone has led to the view (Rutland et al., in preparation) that the southern high-grade region displays an earlier structural and stratigraphic history than is seen in the Skellefte district and the northern zone. A key element in the proposed correlation between the zones is the identification of a major near E-W ductile shear-zone in the southern zone which cross-cuts the dominant NE-SW trending migmatites. This Bygdeträsk Shear-zone is regarded as coeval with the E-W trending folding in the Skellefte Zone and is allocated to D_2, while the earlier deformation in the migmatites of the southern zone is allocated to D_1.

It follows from this correlation that the principal deformation (D_1) in the migmatites of the southern zone is older than any observed in the Skellefte zone. It then also becomes possible that the stratigraphic age of the metasedimentary sequence in this zone is greater than that of any part of the Skellefte sequence.

The present study was undertaken to test this hypothesis which has wide implications for regional tectonic interpretations.

Geochronological Studies

The D_2 Bygdeträsk shear-zone exposes progressively deeper crustal levels moving from west to east. In the west it is a discrete shear-zone, with some retrograde features and it remains discrete as far as the western margin of the Rönnskär map sheet where some felsic veining and garnet crystallisation follow its development. Further east, in the area north of Lövånger, it is a diffuse zone separating areas to north and south which retain the earlier D_1 NE-SW trends. New migmatites are developed in this zone.

Samples were collected for dating from minor granitoid intrusions which can be related to the deformation phases in their host rocks at three localities It therefore becomes possible to allocate minimum ages to deformation phases which are cut by intrusions and maximum ages to deformation phases which also deform intrusions.

Locality 1 is a road-cut on Route E4 at Rävelsberget, about 8 km. north of Lövånger. Two similar felsic intrusions were dated from this locality, which is representative of areas with the D_1 NE-SW trend unaffected by the D_2 E-W deformation.

The migmatites at this locality show very strong deformation with steep south-westerly plunges. One of the intrusions, however, clearly cross-cuts the migmatite fabric and consequently provides a minimum age for the principal deformation in the migmatites.

The U-Pb-ion probe data was acquired on the NORDSIM instrument (Cameca SIMS 1270) machine in Stockholm. The high-uranium zircon population, often found in late to post-orogenic intrusions, consist dominantly of slender, zoned igneous grains, some of which contain inherited cores ranging in age from 1900 Ma to 2030 Ma. The magmatic (emplacement) age for the intrusions is provided by a well-defined cluster of ion-probe results at circa 1860 Ma.

This age is supported by four TIMS U-Pb monazite ages from the same rocks which indicate monazite crystallisation also at circa 1860 Ma.

Locality 2 is on the shore at Sillhällorna, about 1 km. south-east of the Chapel on the Bjuroklubb peninsula.

A number of coastal outcrops north of Lövånger demonstrate that the early N-E trending fabric of the migmatites is modified to varying degrees to a new W to NW trending fabric related to shear-zones. In pelitic lithologies migmatisation accompanies the shearing and a new N-W trending fabric can completely obliterate the earlier fabric. At Locality 2 a prominent shear-zone cuts quartzo-feldspathic metasediments, and granitic veins are commonly emplaced along the shears. It is inferred from the relationships in both pelitic and quartzo-feldspathic lithologies that this granitic phase is effectively the same age as the shear zones.

Conventional (TIMS) U-Pb zircon work had indicated two zircon ages of circa 1860 Ma and circa 1890 Ma, which we believe are geologically meaningful. The magmatic age of the felsic intrusion has now been substantiated from a coherent set of very precise ion-probe zircon data at 1858±3 Ma.

Locality 3 is at Trehörningen, about 2 km. south of Vallen. It lies within the superposed Bygdeträsk shear-zone. The cross cutting nature and younger age of this shear-zone, are clearly demonstrated by the aeromagnetic map data. The shear zone is broadly near E-W but it has been deformed by younger N-E trending shear zones and at this locality has a NW-SE trend. The foliation is approximately vertical and a strong near vertical lineation can be observed, parallel to the axes of amphibolite boudins. However, the pelitic lithologies also display folding of the foliation on near horizontal axes. The vertical linear structure is interpreted to be the refolded linear fabric of the migmatites outside the shear zone while the folding on horizontal axes is attributed to the formation of the shear zone.

A number of small tonalitic bodies can be shown to be deformed by the folding associated with the shear-zone. They are interpreted as having been emplaced after the main deformation in the migmatites outside the shear-zone but before the deformation which produced the shear-zone.

The U-Pb ion probe data reveal a complex history for this zircon population which has an age range between circa 2720 Ma and 1800 Ma. The youngest group of data (from zircons of magmatic aspect) suggests a zircon crystallisation age of close to 1880 Ma and we interpret this as the magmatic (emplacement) age of the intrusion. This result therefore pushes back the minimum age for the first (NE trending) deformation, D_1, to 1880 Ma.

Some other zircons have similar magmatic aspect to the youngest group but are dated between 1895 and 1900 Ma. The magmatic character gives a strong indication that they can be attributed to the melting event associated with the D_1 migmatisation which preceded the intrusion. Given the similar ages obtained from inherited zircon populations at Localities 1 and 2, it can be inferred that the D_1 migmatisation occurred between 1895 and 1900 Ma.

The older zircons between circa 2720 and 1930 Ma are interpreted as inherited components from the source rocks to the host metasediments. This is consistent with the presence of thin xenoliths of metasediment within the granodiorite. The youngest of these detrital components dated at circa 1930 Ma provides a maximum age for deposition of the sedimentary sequence.

Conclusion

In summary, the dates obtained above indicate:

- The second E-W main deformation episode, D_2, can be confidently dated at about 1860 Ma from Locality 2.
- The first NE -SW main deformation phase, D_1 and migmatisation in the migmatites is known

to be older than 1860 Ma from both Localities 1 and 2, and older than 1880 Ma from Locality 3.

- It is possible that ages of approximately 1890 and 1900 Ma obtained at Localities 1 and 2, and approximately 1895-1900 Ma obtained from Locality 3, correspond to the migmatisation associated with D_1.
- Earlier zircon populations at Locality 3 indicate that the original sediments included Archaean source material and that the maximum age of sedimentation was circa 1930 Ma.

If the correlation of the main E-W deformation phase with the E-W deformation in the Skellefte zone, advanced by Rutland et al. (op. cit.), is accepted, these results suggest that the latter deformation took place at about 1860 Ma. This fits well with the chronology summarised by Billström and Weihed (1996), where the stratigraphic age of the folded succession is bracketed between 1890 and 1870 Ma.

Rutland et al. (op. cit.) infer from the regional relationships that since the early NE trending deformation (D_1) in the migmatites in the southern zone is absent in the Skellefte domain, it should be older than the deposition of the Skellefte succession. The inferred age of about 1890-1900 Ma agrees well with this proposition. Moreover, the felsic intrusion at Locality 3, which also gives a minimum age of 1880 Ma for the early migmatisation, agrees well in age with the early Jörn phase of intrusion in the Skellefte district.

The presence of a major episode of deformation and metamorphism between 1890 and 1900 Ma, has previously been demonstrated in the Pihtipudas and Pyhajarvi areas in Finland (Aho, 1979), in a volcanic zone often broadly correlated with the Skellefte district. There, an older sequence of intrusives and associated volcanic rocks which has been strongly deformed and metamorphosed, is dated between 1900 and 1920 Ma. Younger felsic and mafic volcanics and associated granitoids are dated between 1890 and 1860 Ma and are much less deformed. Kousa et al (1994) recognise that only the younger of these phases of volcanism is equivalent in age to the Skellefte sequence.

Thus it can be concluded that in both the Skellefte area and in Finland, a major period of deformation and metamorphism preceded deposition of the Skellefte sequence and its equivalents. There must be a major unconformity at the base of this sequence, although direct evidence of this has not been published.

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The Tectonic Setting of Granitesin the Halls Creek and King Leopold Orogens, Northwest Australia

S. Sheppard*, T.J. Griffin and I.M. Tyler

Geological Survey of Western Australia
100 Plain Street, East Perth, W.A. 6004 Australia
s.sheppard@dme.wa.gov.au

The geology and chemistry of granites in the Halls Creek and King Leopold Orogens in the Kimberley region of Western Australia, suggests that they were generated during northwestward subduction of oceanic crust beneath the Kimberley Craton. The end of subduction was marked by oblique collision of the Kimberley Craton with, and suturing of it to, the North Australian Craton to the southeast. This conclusion contrasts with previous interpretations of the orogens as an ensialic rift.

Regional geology

The Halls Creek and King Leopold Orogens (Fig. 1) form a continuous belt about 600 km long, and are just two of a number of orogenic belts in northern Australia that formed during the Palaeoproterozoic. The orogens were reactivated several times during the remainder of the Proterozoic and in the Palaeozoic. Palaeoproterozoic rocks within the Halls Creek Orogen comprise the Lamboo Complex, and can be divided into three parallel zones (Fig. 1): the Hooper Complex in the King Leopold Orogen is a continuation of the western zone in the Lamboo Complex (Tyler et al., 1995). The three zones in the Lamboo Complex had separate histories until c. 1820 Ma when they were intruded by massive monzogranite and syenogranite plutons. Previous correlations of older units between the three zones are no longer regarded as valid, and the zones may represent tectonostratigraphic terranes (Tyler et al., 1995). Volumetrically, felsic magmatism in the two orogens is dominated by the 1865-1805 Ma Bow River batholith, and the 1865-1850 Ma Whitewater Volcanics. However, numerous tonalite and subordinate trondhjemite sheets that comprise the Dougalls suite are also an important part of the granitic magmatism in the Lamboo Complex.

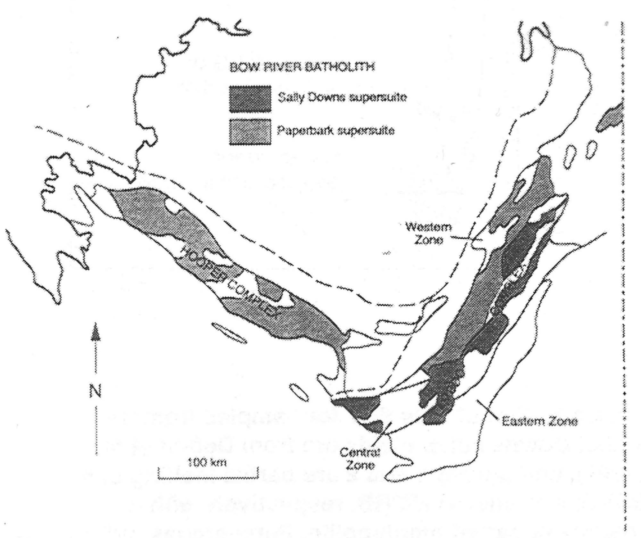

Figure 1. Distribution of the Bow River batholith and its component supersuites.

Bow River Batholith

The Bow River batholith is subdivided into the 1865-1850 Ma Paperbark supersuite and the 1835-1805 Ma Sally Downs supersuite.

Paperbark supersuite

Granites of the Paperbark supersuite are restricted to the Hooper Complex and the western zone of the Lamboo Complex. The Paperbark supersuite is dominated by K_2O-rich granodiorite, monzo-granite and syenogranite, and it is comagmatic with the Whitewater Volcanics. The granites have high K_2O, Rb, Y, Th, and K_2O/Na_2O, and low Sr, Sr/Y and K/Rb, and thus resemble other Palaeo-proterozoic granites from northern Australia (Fig. 2).

Sally Downs supersuite

The Sally Downs supersuite mainly intrudes the central zone of the Lamboo Complex, but is absent from the Hooper Complex of the King Leopold

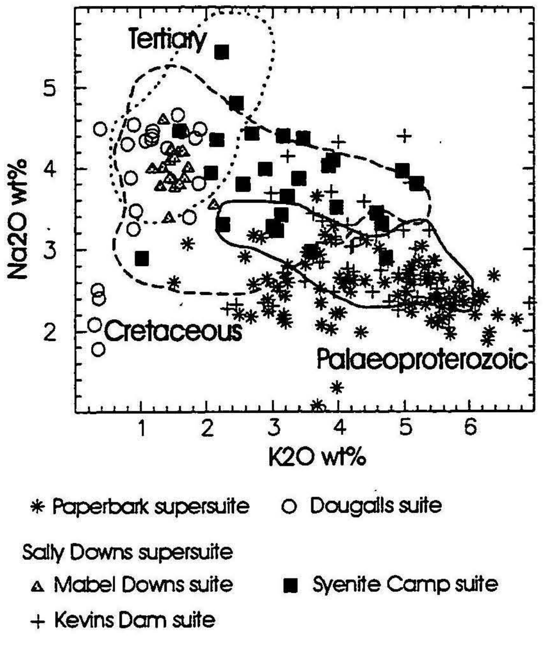

Figure 2. Whole-rock K_2O vs Na_2O for granites in the Lamboo Complex compared with Palaeoproterozoic granites of north Australia (Wyborn, 1988), and subduction-related granites from the Cretaceous (Atherton and Sanderson, 1985; Chappell and Stephens, 1988) and the Tertiary (Whalen et al., 1982; Whalen, 1985).

Orogen. It consists of three distinct suites; the Mabel Downs suite (dominantly tonalite), the Syenite Camp suite (Y-depleted granodiorite and monzogranite), and the Kevins Dam suite (K_2O-rich monzogranite and syenogranite).

Tonalites of the Mabel Downs suite were intruded along the eastern edge of the central zone. They are characterized by high Al_2O_3, Na_2O, Sr, Sr/Y (Fig. 3), and low Y and Yb, in conjunction with moderately to strongly fractionated REE patterns that lack a negative Eu anomaly. These features are shared by Archaean high-Al tonalite-trondhjemite-granodiorite suites and their Phanerozoic equivalents. Granites in the Syenite Camp suite have mantle-normalised patterns that are Sr-undepleted and Y-depleted, in contrast to many Palaeoproterozoic granites in northern Australia. Their major and trace element abundances are similar to the Coastal Batholith of Peru and the western part of the Peninsular Ranges Batholith (e.g. Fig. 2). Massive granites of the Kevins Dam suite intrude all three zones in the Lamboo Complex, and they have compositions similar to the Paperbark supersuite.

Dougalls Suite

In the central zone, coeval with, and to the east of, the Paperbark supersuite, turbiditic sediments and mafic volcanics formed an oceanic island arc. These rocks were intruded by extensive tonalite and trondhjemite sheets of the Dougalls suite and metamorphosed at high grade at c. 1850 Ma. Samples from the Dougalls suite are characterised by low K_2O, Rb, LREE, Th, U and Rb/Sr, and high Na_2O, CaO and K/Rb. The most primitive tonalites have low Sr/Y ratios, moderate Al_2O_3 and Y contents and weakly fractionated REE patterns similar to low-Al tonalite-trondhjemite-granodiorite suites. The Dougalls suite resembles Phanerozoic tonalites and trondhjemites found in island arcs, or along continental margins related to subduction or subsidiary back-arc spreading.

Interpretation

Granites of the Paperbark supersuite were intruded into the margin of the Kimberley Craton between c. 1865 and 1850 Ma, probably in an extensional basin setting behind an oceanic island arc to the east. The granites may be the products of partial melting of intermediate high-K calc-alkaline rocks, perhaps represented by early, dismembered tonalite intrusions in the supersuite, or of a mixture of late Archaean continental crust

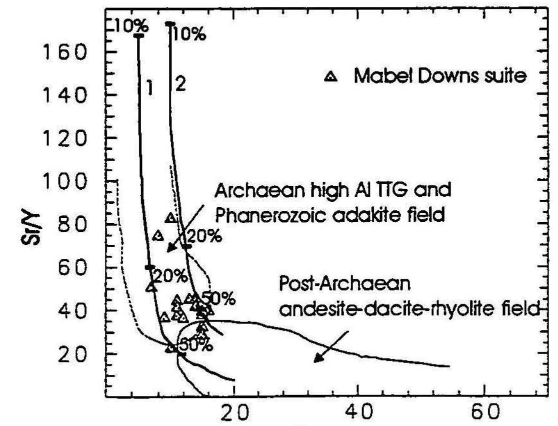

Figure 3. Plot of Y vs Sr/Y for samples from the Mabel Downs suite. Fields are from Defant et al. (1991), and curves 1 and 2 are partial melting of N-MORB and altered MORB, respectively with a residue of garnet amphibolite. Percentages indicate degrees of partial melting. Curves are from Defant and Drummond (1990).

and recently underplated mafic magma. The Mabel Downs suite was intruded between c. 1835 and 1820 Ma into the new continental margin that formed after accretion of the island arc to the Kimberley Craton, above a west-dipping subduction zone. The Sr and Y contents of the Mabel Downs suite can be reconciled with partial melting of a MORB-like source in thickened continental crust with a garnet amphibolite residue (Fig. 3). The Syenite Camp suite was formed by partial melting of amphibolitic source rocks at ≤10-12 kb in an Andean-type setting. Collision of the Kimberley and North Australian cratons at c. 1820 Ma was followed by intrusion of K_2O-rich monzogranite and syenogranite of the Kevins Dam suite.

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Timing of Palaeoproterozoic Events in North Sweden with Comparisons to Other Parts of the Baltic Shield

Torbjörn Skiöld

Swedish Museum of Natural History
P.O. Box 50 007, SE-104 05 Sweden
torbjorn.skiold@nrm.se

The earliest Palaeoproterozoic rocks are called Sumi-Sariolan in Russia and parts of eastern Finland, the lower Lapponian in northern Finland and with the local name Kovo Group (Martinsson, 1997) in northern Sweden. They have been discordantly deposited upon the eroded Archaean basement gneisses and compose debris from these gneisses with intercalations of mafic volcanics (Meriläinen, 1980). In the Kuhmo-Suomussalmi terrain this sub-unit appears in an E-W trending strike-slip fault system (Luukkonen, 1992) whick also carries deposits of andesites. Indirect and minimum dates of these supracrustals are supplied by layered gabbro with ages of about 2.44 Ga. New dates from the Belomorides and Kola (Amelin et al., 1995) on such supposedly plume-related intrusions also indicate another magmatic peak at about 2.50 Ga. Layered intrusions of about this age also appear subconcordantly upon a 2.67 Ga old (e.g. Öhlander et. al, 1987) Archaean gneiss dome and below the rift-related Peräpohja Shist Belt immediately to the north of the Bothnian Bay (Perttunen, 1989).

It seems that there are no major influxes of granitoids in the early Palaeoproterozoic. Despite a 2.4 Ga Gabbro-granite association in the Belomorides (see Bogdanova and Bibikova, 1993), the whole period leading to the upstart of the Svecofennian orogeny at plus 1.9 Ga, is in its first stages probably characterized by thermal upwellings in the upper mantle leading to the breaking-up and separation of parts of the old craton.

Rifting and near-continent sedimentation,

Approx. geochronologiocal summary from c. 1.86 Ga and backwards:

1.86 Ga	Ductile E-W shearing, migmatization and plutonism in the Bothnian Basin
> 1.86 Ga	Accretion of arcs to the Archaean craton
> 1.88 Ga	Ductile deformation in north Sweden
1.882 ±1.004Ga.....	Kiruna Iron Ores and host porphyries
1.86 -1.91Ga	Voluminous Svecofennian plutonism and volcanism
1.92 - 1.93Ga	Mainly tonalites in NW - SE trending zone
< 1.93 Ga	Lapland Granulite Belt towards SW
> 1.95 - 1.90Ga	Kalevian turbidite sedimentation
1.96 Ga	Late rift stage and MORB volcanics
> 1.95 Ga	Jatulian/KGG/middle-late Lapponian supracrustals
	Pillow lavas and shallow marine sediments,
	Black schists and dolomites (key horizon!),
	depleted mantle with positive $\varepsilon Nd = 3$ to 4
2.02 - 2.06 Ga	Felsic horizons in Kittilä and Kiruna
c. 2.20 Ga	Transecting albite-diabase dykes, feeder(?)-dykes
< 2.33 Ga	Early Jatulian tholeiites
	Orthoquartzites (key horizon!)
c. 2.3 - 2.5 Ga	Sumi-Sariolan, basement debris in the form of arkoses and conglomerates
2.44, 2.50 Ga	Mafic layered intrusions on Archaean basement

associated with mafic extrusions because of extensional processes in the mantle, is a mature process during the following Jatulian time; named middle and upper Lapponian in N Finland, and different local volcanosedimentary names in Sweden and Norway. The older age limit for the Jatulian sub-unit is poorly constrained, but it ought to have started before 2.2 Ga ago. Mafic dykes with zircon ages close to 2.2 Ga have transected the Runkaus basalts of Peräpohja (which themselves indicate an imprecise Sm-Nd age of 2.33 Ga; Huhma et al., 1990), and other pre-Jatulian formations in Finland and Sweden. Some of these dykes have a within-plate character and may have been acting as feeder dykes to the early Jatulian sequences. Thus, although there have been a number of transgressions and considerable local variations in the early Proterozoic until c. 2.0 Ga, the general impression is mostly that of a continuous evolution in extensional regimes.

The lower parts of the mafic volcanics at Kiruna and corresponding successions in northern Sweden (KGG) are dominated by tuffs and pyroclastic flows and are considered to be time equivalents to the middle-upper Lapponian in north Finland. Both these sub-units have been deposited upon arkoses, ortho-quartzites and micaceous sediments including carbonate-bearing rocks. The upper and more extensive parts of the KGG constitute tholeiitic lava flows and mafic intrusions. Some are terminated by pillow lavas, and the flows are often separated by black schists. Komatiites or picrites also occur. In a few circumstances, felsic parts of the Finnish mafic volcanics have given zircon ages in the 2.06 - 2.02 Ga age interval (Huhma, pers. comm., 1993). The Jatulian volcanics are mainly tholeiitic basalts, and the sedimentary strata are comprised of mature detritus. They were deposited in stable, shallow marine to fluviatile environments (Sorjonen-Ward, 1989). Dolomites interlayering the Jatulian sequences show enrichment of the C-13 isotope, while later carbonates at some stage after c. 2.1 Ga ago return to values normally obtained for sedimentary deposits (Kahru, 1993). A 1.93 Ga Sm-Nd isochron (Skiöld and Cliff, 1984) has been obtained from metabasalt between the Viscaria A and B-zone ores below the pillow lavas close to the city of Kiruna. The age marks the crystallization of secondary minerals and is a minimum age for the original rock formation.

A number of drill core samples from mafic lava horizons in the upper part of the KGG and above the Viscaria Cu-ores have been investigated for Sm and Nd isotopes. Their chondrite-normalised REE patterns vary from LREE depletion to enrichment which narrow to about 10 in the HREE field. These tholeiites have positive initial εNd values between +1 and +4 (n=17) and their magmas have obviously ascended through deep-reaching fractures with no or very limited crustal contamination. No whole-rock isochron have been calculated from the data, but together with observations from Finland (Perttunen, 1989; Huhma et al., 1990) they clearly demonstrate the widespread depleted character of the pre-2.0 Ga Jatulian mantle below this continent.

Preliminary zircon age data are presented on a felsic ash tuff appearing in the above mentioned Viscaria A-zone. The felsite is laminated, with graded bedding on the mm-scale, and grades into black schist with increasing content of graphite (Martinsson, 1997). Near the felsite appear silicified chemical sediments which together with the pronounced albitization of the rocks reflect the effects of later circulating solutions. Zircon morphologies are quite variable and indicate maximum 207/206Pb ages of 2.13 Ga for oval shaped brownish zircons. However, crystallization of beautiful elongated, transparent zircons seems to have taken place at about 2.04 Ga and may represent the time of volcanism. If so, the results would be in agreement with a correlation between these occurences at Kiruna (KGG)and similar lithologies assigned to the Kittilä Group (Hanski et al., 1997) in northwest Finland.

The occurences of ophiolites in eastern and northernmost Finland (Hanski, 1995) indicate the accretion of c. 1.95 Ga oceanic crust at some later stage to the shield. These rocks mark another stage of rupturing of the old craton and the tapping of a depleted upper mantle with MORB affinities (Peltonen et al., 1996). They are overlain by Kalevian turbidites, which are generally considered to mark the initial flysch-type sediments of the Svecofennian Orogen.

From about 1.95 Ga ago the Lapland Granulite Belt started to form along thrust planes towards west and south. We see remnants of such nappe transports in the Belomorides. Also the contemporaneous occurrences of granitiods north of the belt are indications of a collisional plate tectonic environment. At about this time a belt is formed running NNW - SSE from Lake Ladoga into northern Sweden with a concentration of igneous rocks emplaced in the time interval 1.93-1.92 Ga and which do not seem to occur anywhere else in the Shield. The Belt depicts the border for exposed Archaean rocks in the east and acts as a transition zone for initial εNd values indicating essentially juvenile magmatism to the south.

Conglomerates with felsic volcanic inlayers are discordantly deposited on top of the Jatulian KGG in Sweden. That volcanic horizon has been dated(Skiöld et al in prep) to 1881±5 Ma which constitutes the maximum age for the following suite of volcanic to subvolcanic Ore Porphyries containing the Kiruna Iron Ores. Likewise unpublished zircon ages from a dyke crosscutting the Kiruna Iron Ores indicate an age of 1883 ±4 Ma. Granophyres and granites with a tectonic position similar or younger than the Ore Porphyries have ages of 1880±3 Ma (Cliff et al., 1990) and 1879±7 Ma (Skiöld and Öhlander, 1989). From these indirect datings we may conclude that the Kiruna Ore Porphyries (which I have unsuccessfully tried to date at several localities) have an age of 1882±4 Ma. In contrast to the Jatulian mafic volcanics, the Ore Porphyries have markedly negative initial εNd ratios (Nyström pers. comm., 1994) indicating ensialic assimilation of old continental crust.

One way of modelling the geodynamics of the Svecofennian Orogen starts with the marine depositions outside a passive rifted margin to the southwest followed by the progressive accretion of arc sequences towards the Archaean craton in the northeast (e.g. Öhlander et al, 1993; Nironen, in press). These processes continued untill about 1860 Ma ago. In northern Sweden the degrees of deformation in rocks as a criteria for their ages works to the extent that an age of 1880 Ma seems to separate severely foliated rocks from those of relatively brittle deformation or suffering from the intrusion of younger plutonics. However, in the Bothnian Basin to the south of the Skellefte ore district, plastic migmatite-forming deformations occur at 1860 Ma. E-W shearing, migmatism and plutonism cross cut earlier fabrics (Rutland et al. 1997).

Thermal reactivation, thrusting and exhumation reset some of the softer radiometric clocks in the central Archaean rocks of the Shield during Paleoproterozoic time. K-Ar ages from Finnish Karelia give c. 1850 and 1800 Ma for hornblendes and biotites respectively (Kontinen et al., 1992) as a result of orogenic thrusting towards the east. U-Pb titanite ages in the Russian Belomorides span the 1880 - 1800 Ma time interval, while the corresponding ages in Russian Karelia are in excess of 2600 Ma (Skiöld et al., 1997). The differences on the Russian side are considered to reflect thrusting towards the west in connection with the Lapland Granulite Belt. In that process the Archaean rocks of the Belomorian terrain were brought up through their titanite closing temperatures leaving the Karelian rocks unchanged with respect to titanite ages. The differences in tectonic style in the White Sea and the Baltic Sea regions are very evident. While 1.9 Ga igneous rocks are non-existent in the Belomorian, the Svecofennian Orogen of the Baltic Sea region has produced great volumes during a short time interval with characteristics which in many aspects match modern subduction-related processes.

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An Overview of the Early Proterozoic Depositional and Deformational History on and Adjacent to the Karelian Craton, Fennoscandian Shield

Peter Sorjonen-Ward

Geological Survey of Finland,
02150 Espoo
FINLAND

The Fennoscandian - or Baltic - Shield can be conveniently visualized in terms of three major NW-SE trending crustal domains that have shared a common history since about 1.8 Ga, namely the Paleoproterozoic Svecofennian Province exposed in southwestern Finland and Sweden, the Kola-Lapland Province occupying the Kola Peninsula and northernmost Norway and Finland, and the intervening Late Archaean Karelian Province, covering eastern Finland and the adjoining Republic of Karelia (Gaál and Gorbatschev 1987),. The Kola-Lapland Province, to the NE of the Karelian craton, records the amalgamation at around 1.9 Ga of several distinct crustal units of both Proterozoic and Archean age, and is more characteristic of collisional tectonic processes. In contrast, the Svecofennian Province, to the SW of the Karelian craton, and covering more than half a million square kilometres, is entirely early Proterozoic in age, and indicates relatively rapid formation and accretion of new crust between about 1.97-1.86 Ga. The Karelian Province preserves a varied and episodic record of early Proterozoic magmatism and sedimentation and although these sequences have generally been metamorphosed to at least greenschist facies, they have experienced relative minor deformation; this region therefore provides a useful framework for interpreting the early Proterozoic evolution of the remainder of the Shield.

Convergence in the northern part of the Karelian province: two-sided collision

The northern part of the Karelian province is dominated by the Lapland greenstone belt, which is the most extensive mafic and ultramafic sequence in the Shield. Results of the POLAR seismic refraction profile indicate that the greenstones overlie Archean basement at a depth of 5-8 km (Gaál et al., 1989), which is consistent with regional structural studies suggesting that the belt was deformed as a thin-skinned foreland fold and thrust belt related to both Svecofennian convergence from the SW and broadly coeval emplacement of the allochthonous Lapland granulite belt from the NE (Ward et al., 1989). Further support for this interpretation comes from the recent recognition of depleted mantle characteristics in a suite of highly deformed serpentinites (Hanski, 1995) indicating that part of the greenstone sequence is indeed allochthonous. If thrust from the SW, in relation to Svecofennian convergence, a minimum age is constrained by the post-thrusting 1.89-1.88 Ga Haparanda plutonic suite.

SHRIMP zircon data and Sm-Nd model ages indicate that at the protoliths to at least some of the Lapland granulite belt had a relatively brief crustal residence age before deformation and metamorphism. Deformed pyroxene-bearing granitoids with zircon ages of 1.93 Ga are considered to provide a minimum constraint on age of the granulite protoliths but cannot directly provide information on the maximum age of granulite metamorphism or thrusting. U-Pb monazite and SHRIMP zircon data nevertheless indicate that the cooling and exhumation of the granulites commenced as early as 1.91 Ga. Until more precise data become available, the relative timing with respect to Svecofennian collision remains uncertain and this may preclude the granulites from being a potential source for Svecofennian sediments. Given the similarities in zircon provenance age data, a plausible alternative is that the granulite protoliths and the Svecofennian sediments were both derived from a common, as yet unidentified source and exhumation of the Lapland granulites post-dated the earliest stages of the Svecofennian orogeny.

Rifting and convergent deformation in the transition zone between Karelian and Svecofennian provinces

The southwestern margin of the Karelian Province was strongly deformed and imbricated during NE-directed thrusting associated with the Svecofennian Orogeny at about 1.9 Ga.Alpine tectonic paradigms were already applied to the transition zone between the Karelian and Svecofennian provinces by Wegmann (1928), and the essential elements of this synthesis remain valid. In many respects the comparison is very apposite, with similarities in the nature of the pre-collisional passive margin sedimentary sequence, as well as in the emplacement of nappes onto the foreland, including units analogous to the Ultrahelvetic nappes - with attendant ophiolitic fragments - the monotonous flysch of the Helvetic and Penninic nappes. The extent of basement reworking, of the initiation of extensional deformation during ongoing convergence, and the distribution and timing of post-thrusting granitoid emplacement are also remarkably similar.

Instead of collision with another continental plate however, the Svecofennian orogeny records the formation and accretion of juvenile crustal material. The earliest collisional event recorded is constrained by the ages of the Outokumpu assemblage and Jormua ophiolite (1.97-1.95 Ga), and the earliest Svecofennian intrusions, which are 1.93-1.91 Ga in age and have participated in all stages of Svecofennian deformation, together with enclosing supracrustal gneisses. These intrusions have at times been regarded as a Svecofennian basement but this is at variance with local intrusive relationships and isotopic data, as well as geochemical similarities with some of the associated 1.92 Ga volcanics (Lahtinen, 1994). Various permutations of the plate tectonic paradigm, including reversal of subduction polarity following initial collision, curvilinear subduction zones or a further arc-arc collision and accretion are invoked to explain the most extensive phase of volcanism, magmatism and deformation in southern and western Finland between 1.89-1.86 Ga and will not be considered or evaluated further here (Hietanen, 1975, Park et al., 1984, Ward, 1987, Gaal, 1990, Ekdahl, 1993, Lahtinen, 1994); the most comprehensive attempt to date to integrate geochemical and isotopic data with regional evolution is found in Lahtinen (1994).

The Karelian Province and Lapland between 2.5-2.1 Ga - predominantly mafic magmatism and extension.

The Lapland greenstone belt is by far the largest mafic-dominated terrain preserved in Finland and despite its Paleoproterozoic age the overall dimensions, deformation style, metamorphic grade and abundance of ultramafic volcanism are more reminiscent of the late Archean Abitibi and Norseman-Wiluna greenstone belts than other greenstone units within the Fennoscandian Shield. A sequence of bimodal komatiitic and felsic volcanics dated at around 2.5 Ga unconformably overlie the Archean basement and represent the onset of rifting, followed by widespread emplacement of gabbro-norite layered intrusions between 2.45-2.39 Ga. Terrigenous clastic sediments discordantly overlie these layered intrusions, with field relationships commonly recording angular discordances, suggesting tilting of the layered intrusions during ongoing extensional deformation (Ward et al., 1989).

A prolonged phase of stable terrestrial to shallow margin sedimentation is recorded in eastern Finland by a well-defined sequence of orthoquartzites and related sediments including glacial deposits and paleoregoliths, now represented by kyanite quartzites up to 80 m thick (Kohonen and Marmo, 1992). Fluviatile orthoquartzites are followed by an arkosic quartzite sequence, which records a transient episode of basement rejuvenation and rifting associated with a very distinctive suite of approximately 2.2 Ga differentiated mafic sills found over large parts of eastern and northern Finland.

The Karelian Province between 2.1-2.0 Ga - rifting and passive margin subsidence

Further episodes of mafic magmatism in Karelia are recorded by mafic and minor volcanics, sills and dykes dated at 2.10 Ga, and 2.05 Ga. These broadly coincided with rifting and subsidence of the Karelian craton margin, recorded by coarse clastic turbidites, carbonates, iron formations and finer-grained graphitic schists, the latter hosting the extensive, though low grade Ni-Cu-Zn deposits. Coarse clastic lithofacies recording an Archean basement provenance are intercalated with mafic volcaniclastic deposits dated at 2.1 Ga and are interpreted as proximal, partly channelized prograding fan sequences correspond to the initial phase of rifting and basin subsidence. Karhu (1993) found that carbonates in the Fennoscandian Shield record a significant isotope excursion, from more normal $\delta^{13}C$ values of around 0 per mil at 2.4 Ga to +12 per mil by 2.2 Ga, and back to 0 per mil by 2.0 Ga, which is believed to require the burial of a large amount of carbon, in the form of organic matter, in marine sediments at that time; this is consistent with the presence of extensive black schists horizons within the rifted margin sedimentary sequences of this age in eastern Finland.

The Karelian Province at 1.97-1.95 Ga - renewed rifting at a passive margin and ocean basin formation

Rifting culminated in the formation of oceanic crust at 1.97 Ga, fragments of which were subsequently thrust back onto the Karelian craton as the Jormua complex, which is renowned as one of the few examples of Paleoproterozoic ophiolites (Kontinen, 1987) and the Outokumpu assemblage, which is perhaps best known for its Cu-Co-Zn deposits and chromian skarns. Mafic to ultramafic magmatism was widespread throughout Karelian and Kola Provinces at this time - a major province of picritic flood basalts is present around Lake Onega in eastern part of the Karelian craton, while a suite of picrites, hosting important nickel deposits at Pechenga, in the Kola Peninsula, also have an age of 1.97 Ga. Hanski (1992) found that the chemistry and isotopic characteristics of the Pechenga ferropicrites were also consistent with a plume-related origin.

The recent recognition of a suite of serpentinites with depleted harzburgitic mantle affinities within the greenstone sequence in central Finnish Lapland (Hanski 1995) indicates that a large part of this sequence must be allochthonous, presumably emplaced onto the Karelian craton at some time prior to 1.89 Ga (Ward et al., 1989). It is hence conceivable that the Lapland greenstones, the Jormua ophiolite and the Outukumpu assemblage each represent the deformed remnants of a formerly contiguous segment of oceanic crust being consumed towards the southwest and emplaced across the Karelian province foreland; this would imply an allochthon of dimensions 50-100 km in width and greater at least 500 km in length, which for example, is an area similar in magnitude to that of the Oman ophiolite. It is tempting therefore to speculate as to whether a plume was responsible for extensive mafic and ultramafic magmatism at the scale of the Fennoscandian Shield in the period preceding the 1.97-1.95 Ga rifting process.

Transition from divergent to convergent tectonics - passive margin or arc-derived sediments and incipient collision between 1.95-1.90 Ga.

Sediments associated with and overlying the Outokumpu assemblage and Jormua ophiolite are dominated by remarkably monotonous meta-psammites preserving sporadic evidence for deposition by medium- to thick- bedded mass flows and coarse, chaotic debris flows and high-density turbidity currents in a submarine canyon or channelized inner fan environment. Although most of this monotonous sequence is allochthonous, its widely transgressive nature is indicated by local preservation of basal depositional contacts in several places - conformable transitions from the underlying 2.1-2.0 rift phase deposits, as well as the Jormua ophiolite and the Outokumpu assemblage have been recognized, while in the southern parts of the region, towards the orogenic hinterland, they locally rest directly on Archean basement retaining paleoregoliths.. These relationships may be attributed to flexural uplift at the 1.97-1.95 Ga rifting margin, if not the incipient stages of attempted subduction of the passive margin, with the development of flexural bulge and consequent erosion of the earlier rift-phase passive margin deposits, prior to general subsidence and transgression.

These metasediments occupy a crucial position in any attempt to establish the relationship between the Karelian and Svecofennian provinces. So far, however, attempts to establish their provenance using petrography, chemistry, Sm-Nd model ages and detrital zircon have yielded ambiguous results, neither confirming or pre-cluding derivation from the Svecofennian arc terranes (Ward, 1987). However, the textural immaturity and chemical homogeneity of the sediments seems most compatible with an origin as a major submarine fan system prograding over the subsiding passive margin, just prior to collision, with sediments being derived from the craton, or a more distant orogenic source (Kontinen and Sorjonen-Ward, 1991).

Such a tectonic setting also appears to be in accord with the geometry of compressive deformation, which is most readily interpreted in terms of the Outokumpu assemblage and overlying sediments having been thrust over the Karelian foreland during imbrication and attempted subduction of the craton margin. Moreover, there is no evidence for any contemporaneous arc activity within the craton itself, as would be expected if subduction took place beneath it, nor are there any intrusive or volcanic units within the Svecofennian fold belt known to be coeval with the Outokumpu assemblage or the Jormua ophiolite; the oldest dated Svecofennian arc-related lithologies are some 30 Ma younger than the Outokumpu assemblage, (Huhma, 1986; Lahtinen, 1994). A minimum age constraint for deposition of the sediments (and hence a maximum age for the onset of deformation) is however given by 1.94-1.92 Ga U-Pb SHRIMP ages for detrital zircons - a result which does indeed approach that of the earliest Svecofennian intrusions and volcanics (Lahtinen and Huhma, 1997).

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Chemical and Isotopic Features of Palaeoproterozoic Mafic Igneous Rocks of Australia: Implications for Tectonic Processes

Shen-su Sun

AGSO PO Box 378, Canberra, ACT 2601, Australia (ssun@agso.gov.au)

Geochemical and isotopic features of Palaeoproterozoic mafic - ultramafic igneous rocks of Australia can be interpreted in terms of tectono-magmatic processes which have modified the mantle source regions of these rocks. Samples from Capricorn Orogen, Tennant Creek, Arunta Inlier, Mount Isa Inlier, East Kimberley and Gawler craton have trace element spidergram patterns showing strong 'subduction zone component' signatures, even though many of them were emplaced in an intraplate setting. When combined with time constraints derived from isotope studies, the geochemical data can be used to support the idea that plate tectonic processes operated in various parts of Australia during Palaeoproterozoic. For example, it is possible that mantle sources of 1860-1830 Ma mafic - ultramafic intrusions in East Kimberley have been modified by subduction zone processes, although the exact time when this subduction zone modification took place is not well defined. Thus, it is not clear from geochemical studies whether subduction was contemporaneous with the emplacement of these intrusions.

Geochemical and isotopic information for mafic rocks should be integrated with interpretations of granite geochemistry, SHRIMP zircon U-Pb ages, regional tectonic analysis and interpretation of geophysical data in order to evaluate alternative tectonic models.

Introduction

The lack of undoubted ophiolites, accretionary complexes, oceanic island arcs and tectonic indicators, such as boninites, in the Palaeoproterozoic of Australia has dampened the acceptance of plate tectonic models for this time period.

Neodymium TDM model ages for Palaeoproterozoic felsic igneous rocks in Australia are commonly in the range of 2.2-2.3 Ga, with a large portion having Archaean model ages (e.g., McCulloch, 1987; Sun & others, 1995; Page & Sun, in press). This is in contrast to many other parts of the world where T_{DM} model ages $\leq$ 2.0 Ga are common (e.g., Patchett & Arndt, 1986; Lahtinen & Huhma, 1997). Initial εNd values for Australian samples are commonly in the range of 0 to -4, compared with 0 to +4 for samples from other parts of the world. If plate tectonics had operated in Australia during Palaeoproterozoic, recycling of larger amount of sediment derived from Archaean crust through subduction zone at cordilleran environments is a possible explanation for the difference. If some of the felsic igneous rocks were derived from melting of mafic rocks underplated in the lower crust through mantle plume activities or subduction zone processes, the underplating is probably related to major thermal and magmatic events observable in the geological record. Ages of such events are likely to be considerably younger than those neodymium TDM model ages if the underplated materials have been affected by an older crustal component.

In this review trace element spidergram patterns of Palaeoproterozoic mafic igneous rocks from various parts of Australia, including many intraplate basalts and lamprophyres, are examined. The patterns commonly show strong signatures of 'subduction zone component'. Modification of magma sources in the asthenosphere and lithosphere mantle by subduction zone processes can be suggested. Pb, Sr and Nd isotope data for these rocks are used to place constraints on the timing of this modification.

Geochemical arguments for subduction during Palaeoproterozoic

Hynes and Gee (1986) first suggested that some of the early Proterozoic, subalkaline mafic and ultramafic igneous rocks of Narracoota Volcanics, occurring in a thrust slice on the northern margin

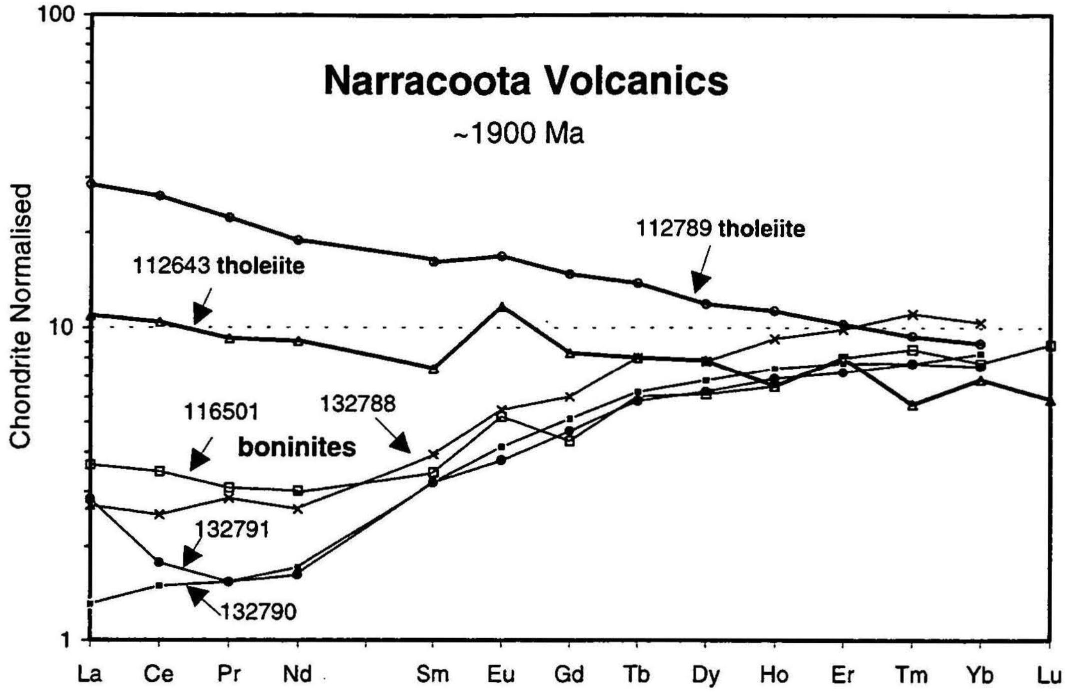

Figure 1. Chondrite normalised rare earth element patterns of tholeiites and boninites of Narracoota Volcanics of Capricorn Orogen, Western Australia (Pirajno & others, 1995; Pirajno, personal communication, 1997).

of Yilgarn Block, are boninites. This suggestion is further supported by chemical data reported in a recent study by Pirajno & others (1995; personal communication, 1997). These ~1.9 Ga volcanic rocks in the Bryah basin have very low TiO_2 contents (commonly <0.3%), high Al_2O_3/TiO_2 ≥40, and low Ti/V (7.1 - 8.6), Ti/Sc (27 - 37) and Ti/Y (100 - 160). Their chondrite normalised rare earth element (REE) patterns show strong depletion from heavy REE towards middle REE with reduced depletion of the light REE, some even being light REE enriched (Fig. 1). These features are not shared by MORB or komatiites of any ages but are characteristics of boninites generated from a very refractory mantle source which has been modified by subduction processes. In addition, composition of spinels from these rocks have very high Cr number (89 - 96) and low Ti contents (~200 ppm) supporting the view that these rocks are boninites with refractory mantle sources (W. R. Taylor, personal communication, 1997). In contrast, tholeiites and calc-alkaline rocks of the Killara Formation in the nearby Yerrida basin to the southeast were laid down in a stable rift environment, associated with chemical sediments and evaporites, on Archaean basement. These tholeiites have light REE enriched patterns with strong Nb depletion (Pirajno, personal communication, 1997) similar to basalts of back-arc basins. Geochemical and mineralogical data for the two suites of volcanic rocks are thus consistent with a subduction-related tectonic model (e.g., Tyler & Thorne, 1990; Myers & others, 1996).

In the Tennant Creek Inlier, Duggan and Jaques (1996) reported chemical and mineralogical data for ~1700 Ma shoshonitic lamprophyres intruding Warramunga Group metasediments. The major and trace element characteristics of these lamprophyres generally conform to orogenic ultrapotassic rocks. High incompatible-element contents and trace element spidergrams show strong 'subduction zone component' with Nb and Ti depletion and Th enrichment. Duggan and Jaques suggested that the mantle source region of these rocks experienced metasomatic enrichment involving a major crustal component, presumably through subduction zone processes. These lamprophyres are phlogopite-bearing, rich in K and Rb. However, a low $^{87}Sr/^{86}Sr$ initial value of 0.701 ± 0.002 (Black, 1977) indicates that the suggested subduction zone enrichment took place in the Palaeoproterozoic not very long before emplacement of the lamprophyre.

Based on petrological and geochemical studies of mafic and felsic igneous rocks several authors have proposed that an active continental margin with subduction zones existed in the Arunta Inlier during the Palaeoproterozoic (e.g., Foden & others, 1988; Sivell, 1988; Zhao & Cooper, 1992). There is also evidence for Precambrian rocks of apparent intraplate settings, with chemical and isotopic compositions suggesting that the mantle sources of these rocks have been modified by subduction processes in the Palaeoproterozoic. For example, some tholeiites of the ~1770 Ma Hart Range Metamorphic Complex have chemical compositions of intraplate basalts, with high TiO_2 (~1.8 %), Zr (~180 ppm), light REE (La ~30 ppm), but they also have strong Nb depletion (Nb/La ~0.5). The ~1150 Ma Mordor Complex in the southern Arunta is composed of alkaline mafic - ultramafic rocks showing strong Nb depletion and Th enrichment. These rocks have low initial εNd values (-9.5 to -11.6) and high initial $^{87}Sr/^{86}Sr$ (~0.711). Their Pb isotope compositions have the high $\mu(^{238}U/^{204}Pb)$ character of crustal lead. Nelson & others (1989) proposed that subduction zone enrichment involving sediments during Palaeoproterozoic is a preferred explanation for these isotope features. The 1080 Ma Stuart mafic dykes in the southern Arunta have geochemistry similar to island arc tholeiites. They have a whole rock Sm-Nd isochron age of 1853 ± 189 Ma (Zhao & McCulloch, 1993). Zhao & McCulloch interpreted it as the age of mantle source and suggested that the lithospheric mantle source of Stuart Dykes was modified by subduction zone processes during the Palaeoproterozoic.

In the Mount Isa Inlier ~1770 Ma continental flood basalts (Eastern Creek Volcanics) up to 4 km thick are exposed in the Leichhardt River Fault Trough. Those of the Cromwell Metabasalt Member, in the lower part of the sequence are more fractionated (with high TiO_2 close to 3 %) than those of the Pickwick Metabasalt Member in the upper part. Both have very similar trace element spidergram patterns (e.g., Wilson, 1982) with light REE enrichment and low Nb/La (~0.5) and Nb/Th (~2), which are unlike those of oceanic intraplate tholeiites with Nb/La >1.0 and Nb/Th >10. Their spidergrams are similar to those of some of Karoo basalts of South Africa. The genesis of the Karoo basalts has been discussed (e.g., Duncan & others, 1997) in terms of mixing of mantle plume with subduction-influenced mantle prior to decompression melting or contamination of plume melts with lower continental lithosphere. A similar interpretation may apply to Eastern Creek Volcanics.

A variety of 1860 - 1820 Ma mafic and felsic rocks occur in the East Kimberley. To the east of the Halls Creek Fault light REE-enriched basaltic andesite to rhyolite in the Butchers Gully Member of the Olympio Formation have intraplate-type chemistry and show no Nb depletion or Th enrichment. In contrast, 1855 to 1830 Ma layered mafic-ultramafic intrusions west of the Halls Creek Fault have trace element spidergram patterns showing distinct Nb depletion and Th and Ba enrichment, suggestive of a 'subduction zone component. Sills of 1790 Ma Hart Dolerite in the Kimberley Basin to the west have similar trace element features to the earlier mafic intrusions. Also in the East Kimberley the 1200 Ma Argyle lamproite and diamonds have chemical and C isotope compositions (Jaques & others, 1989a, b) and Pb isotope composition (Sun & others, 1986) suggesting that their mantle source regions contained crustal material, probably introduced in the Palaeoproterozoic. The 20 Ma lamproites of West Kimberley have Pb isotopic compositions characterised by low $^{206}Pb/^{204}Pb$ (17.2 - 17.9) and high $^{207}Pb/^{204}Pb$ (15.7 - 15.8) and $^{208}Pb/^{204}Pb$ (37.8 - 38.6). Nelson & others (1986) proposed that these lamproites have high μ early history probably developed in Archaean upper crust and then followed by low μ history starting in early Proterozoic through subduction of sediments. Large negative Eu anomalies observed in these lamproites lend support to the sediment subduction model.

Studies of melt inclusions in relic igneous minerals, using ion microprobe and laser ablation ICP-MS techniques are necessary to obtain more data on primary magmatic features, such as 'mobile' elements (e.g., K, Rb, Cs, Sr, Ba, Pb, and U) and water contents of the igneous rocks. Through such studies, the question of whether some of the Narracoota Volcanics are truely subduction related boninites may be answered. Better defined initial Sr isotope ratios for ~1700 Ma shoshonitic lamprophyres near Tennant Creek may be obtained from analysis of igneous apatite, and could be used to constrain the time when subduction modification occurred in the mantle source of these rocks.

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Nature and Significance of Palaeoproterozoic Alkaline Magmatism in Northern Australia

Wayne R. Taylor

RSES, Australian National University
Canberra, ACT 0200 Australia
wayne.taylor@anu.edu.au

Primitive alkaline magmas provide a window to their upper mantle source regions and hence can be useful in deducing the tectonic evolution of the ancient continents. In the North Australian craton and Capricorn orogen, primitive alkaline magmatism of Palaeoproterozoic age falls into two types: (i) ultrabasic-to-basic alkaline basalts and lamprophyres of intraplate affinity derived from asthenospheric sources, and (ii) potassic lamprophyre magmas of shoshonitic to transitional lamproite affinity typical of those associated with subduction-zone-modified litho-spheric sources. In the former case, some examples of felsic alkaline volcanism (such as the Brockman volcanics discussed below) can be unequivocally linked to an alkaline basaltic parental magma of mantle derivation. In northern Australia (and also in other Precambriam terranes world-wide) there are two important periods of alkaline magmatism at ca.1.83 Ga and ca.1.67 Ga which includes magmatism of either ultrabasic-to-basic intraplate type or shoshonitic type depending on location. In northern Australia other periods of Proterozoic alkaline magmatism are known at 0.8-0.9 Ga (kimberlite, lamprophyre and carbonatite), 1.1 Ga (lamproite), ca.1.25 Ga (carbonatite), and 1.87 Ga (trachyte). Some examples are discussed in more detail below.

ca. 1.87 Ga Intraplate Volcanism

Brockman alkaline volcanics and related rocks[Halls Creek mobile zone]

The Brockman alkaline volcanics (Taylor et al., 1995a,b) comprise a cogenetic suite of silica-saturated trachyandesite, trachyte and trachydacite lavas and associated volcaniclastic rocks some of which host REE and niobium mineralization. The trachyte-dominated volcanics, which include well-preserved flows, pillow lavas and pillow breccias were probably erupted from a small shield volcanic complex in a rift-related basin in a shallow marine setting. The geochemistry of the least differentiated lavas (initial ε_{Nd} >+2.6 and Nb/La ~1.2) indicate the Brockman volcanics were mantle-derived with the mantle source having affinity to OIB sources. The tectonic setting differs from Tertiary intraplate trachyte-dominant complexes, such as those in eastern Australia, in that the Brockman volcanics were developed on thin continental crust, and were produced at the beginning of a cycle of turbiditic sedimentation which took place prior to compressional orogeny. The closest modern tectonic analogue is probably the mid-Tertiary intraplate volcanic province of southern New Zealand which developed in small extensional basins in a shallow-marine, continental shelf setting. Volcanism pre-dated (by ~10 Ma) compressional orogeny associated with propagation of the modern plate boundary through New Zealand. The Brockman volcanics may have similarly been in proximity to an early Proterozoic plate boundary that propagated through the Halls Creek mobile zone coinciding with peak metamorphism at 1854 Ma (Page & Hancock, 1988).

ca. 1.83 Ga Potassic Magmatism

Mt Bundey olivine-mica lamprophyres [Pine Creek Inlier, N.T.]

Primitive, incompatible-element-rich olivine-mica lamprophyres (1831±6 Ma) are associated with the post-tectonic composite syenite-granite Mt Bundey pluton (Sheppard & Taylor, 1992). They have some geochemical characteristics in common with West Kimberley lamproites (e.g. high Ba and LREE contents) and contain high-pressure mantle xenocrysts and trace amounts of diamond. Geochemically they most closely resemble post-collisional potassic and ultrapotassic magmas of Mediterranean Europe. In particular, they have a negative Ta-Nb-Ti 'subduction-type' signature and HFSE abundances that are closely similar to those found in jumillite (a type of transitional lamproite) from southeastern Spain. The low initial Sr-isotope ratio of the lamprophyres (0.7038) constrains source enrichment to 2300 Ma before mantle partial melting.

Benmara alkaline complex [Murphy Inlier, N.T.]

The Benmara alkaline complex is a newly recognized zoned, ring complex of shoshonitic affinity dated at 1830±7 Ma. Located at the western end of the Murphy Inlier, it intrudes basement metasedimentary rocks of the 1.87 Ga Murphy Metamorphics and granitoids of the 1.85 Ga Nicholson granite (Hanley, 1996). The Benmara complex comprises ultramafic, mafic and felsic rocks ranging from pyroxenite and kentallenite (olivine monzogabbro) to shonkinite and trachyandesite. The form and composition of the complex is similar to some Alaskan-type intrusions. In a downthrown block to the west, 1832±18 Ma shoshonitic mica lamprophyre (minette) dykes and sills intrude the metasedimentary basement. They are associated with thick flows of high-K andesite of undetermined age and with anomalous concentrations of microdiamonds (Ong, 1991; Lee et al., 1994). The microdiamonds are dominated by fibrous cubes with C-isotopic compositions (-15 to -28‰) indicative of a recycled crustal carbon source. The minette sills are are of similar age and composition to potassic lavas and dykes of the Baker Lake group, Churchill Province, Canada (Peterson, 1994). These rocks include a 1832±28 Ma minette dyke containing abundant microdiamonds (MacRae et al., 1995). A post-collisional setting for their genesis has been inferred (Peterson, 1994).

Gifford Creek shoshonitic monzogabbro andmonzodiorite [Capricorn Orogen]

Monzogabbro and monzodiorite intrusions that are chemically identical to shoshonitic lamprophyres (e.g. Nb/La 0.09, Al/Ti 20) comprise part of the ca.1.83 Ga basement of the Gifford Creek complex (Pearson, 1996).

ca. 1.67 Ga Alkaline Magmatism

Coanjula nephelinite and alkali basalt diatremes [Murphy Inlier, N.T.]

Nephelinite and alkali basalt diatremes of 1665 Ma age, some containing mantle-derived xenoliths and megacrysts, intrude Murphy metamorphics basement rocks (Ong, 1991; Lee et al., 1994). Although they are amongst the oldest, best preserved alkali basalt diatremes known, their form and geochemistry are essentially identical to Tertiary to recent alkali basalt diatremes of intraplate affinity.

Ultrabasic lamprophyre sills, Gifford Creek complex [Capricorn Orogen]

A swarm of primitive, ultrabasic lamprophyre sills of 1679±6 Ma age intrude the ca.1.83 Ga granitoid basement of the Gifford Creek complex in the Gascoyne Province (Pearson, 1996). They are associated with an extensive zone of country-rock fenitization. Geochemically the sills are similar to some South African Group I kimberlites and their initial ε_{Nd} of +2.8 indicate derivation from a depleted mantle source. Their tectonomagmatic affinity is continental intraplate.

Shoshonitic lamprophyres [Tennant Creek Inlier]

Two suites of shoshonitic mica lamprophyres (minettes) of 1664±16 Ma age intrude basement metasediments of the Tennant Creek Inlier (Duggan & Jaques, 1996). The rocks show strongly negative mantle-normalized Nb-Ti anomalies and have Nb/La <0.5 similar to shoshonitic lamprophyres from ancient and modern settings.

Discussion

Although the presence of shoshonitic lamprophyres and intrusive complexes of 1.83 and 1.67 age in Palaeoproterozoic Australia cannot be correlated directly with active subduction processes, it seems likely from analogy with Phanerozoic shoshonites that these magmas were derived from a source that included a subduction-zone modified component, ultimately of recycled crustal origin. This view is supported by the intimate association of mica lamprophyres and microdiamonds having a recycled crustal carbon signature in the Murphy Inlier. This situation is identical to that in the Churchill Province of Canada in which 1.83 Ga minette magmas are inferred to have been produced in a post-collisional setting following earlier subduction (at ca.2.0Ga) and continental collision (Peterson, 1994). In this case the subduction zone was remote from the eventual site of shoshonitic magma production and a similar situation may apply in northern Australia. The question then arises as to how such processes might be recognised in the geological record of this region.

In at least two cases, 1.67 Ga alkaline basaltic magmas of asthenospheric origin have intruded terranes which recorded a 1.83 Ga shoshonitic event. This may reflect replacement (by delamination?) of anomalous potassic lithosphere in at least some parts of northern Australia between these times.

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Metallogeny of the Palaeoproterozoic Skellefte District, Northern Sweden. Relationship Between Ore Deposition and Tectonic Evolution

Pär Weihed

Centre for Applied Ore Studies (CTMG)
Luleå University of Technology, S-971 87 Luleå, Sweden
per.weihed@sgu.se

The Skellefte district occurs within a Palaeoproterozoic (mainly 1.90–1.87 Ga) magmatic province of low to medium metamorphic grade in northern Sweden. The district contains over 85 pyritic Zn-Cu-Au-Ag massive sulphide deposits, gold lode deposits, and subeconomic porphyry Cu-Au-deposits. The massive sulphide deposits are located within, and at the top of a felsic-dominated volcanic sequence attributed to a period of extensional continental or island arc volcanism (cf. Weihed et al. 1992, Allen et al. 1996, Weihed and Mäki 1997). The massive sulphide deposits range in style from deep water sea-floor ores, through sub-seafloor replacement ores, to shallow water and possibly sub-aerial sub-volcanic replacement ores. Two quartz vein-hosted gold lode deposits are located in the eastern part of the Skellefte district at Björkdal and Åkerberg. The Björkdal deposit is currently one of the largest gold mines in Europe and is characterised by a quartz vein system at the contact between a granodiorite-tonalite and surrounding supracrustal rocks. The gold at Åkerberg is hosted by a zone of narrow parallel quartz veins in a gabbroic intrusion.

Introduction

The Skellefte district is a loosely defined area, approximately 150 by 50 km in size, which forms part of the Svecofennian c. 1.90–1.85 Ga supracrustal sequence and associated intrusive rocks in the northern part of Sweden. It is extremely rich in mineral occurrences and today 5 underground mines and one open pit are in operation on VMS-ores, the Björkdal gold-lode

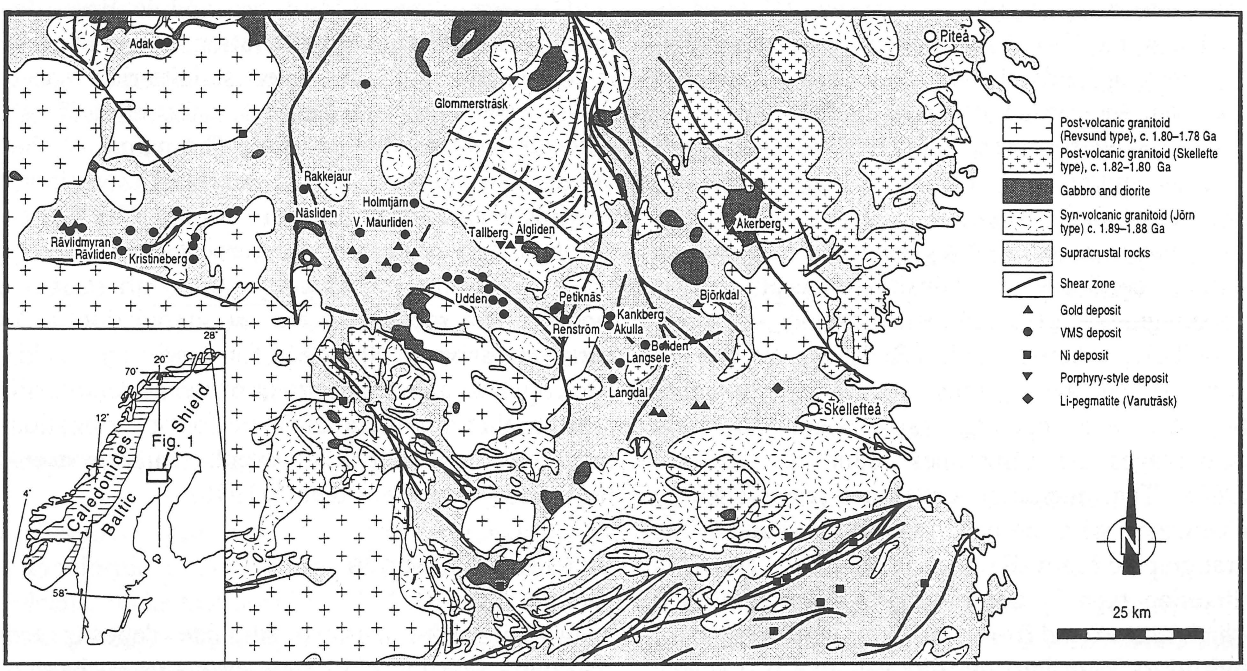

Figure 1. Geology of the Skellefte District with all ore deposits and major mineralization indicated.

deposit is mined in an open pit operation, and the Åkerberg gold-lode deposit is mined underground. Apart from VMS and gold-lode deposits, subeconomic porphyry-type Cu-Au deposits and Ni-deposits occur in the district.

VMS Deposits

Over 85 massive sulphide occurrences are known in the Skellefte district of which 6 (Långdal, Kankberg, Åkulla Östra, Renström, Petiknäs and Kristineberg) are currently mined. This makes the Skellefte district the most important mining area in Sweden today. The total tonnage, which exceeds 160 million tonnes, also makes the Skellefte district one of the largest early Proterozoic VMS districts in the world.

In general the Skellefte district VMS ores are poor in Pb, rich in Zn, As, and Au, and some deposits have a high contents of Sb and Hg. However, compositions vary between individual deposits. Some extremely gold-rich deposits stand out, for instance the Boliden (average 15.5 g/t Au) and Holmtjärn (average 7.4 g/t Au) deposits. The Boliden deposit was also extremely As-rich. Typically the deposits contain 3–5% Zn, 1–2% Cu, and less than 1% Pb. Well developed Cu-rich stringer zones can be identified in several deposits, e.g. Näsliden (Svenson 1982) and W. Maurliden (Allen et al. 1996). Disseminated pyrite is found in the stratigraphic footwall of several deposits, e.g. Långsele, Långdal, and Holmtjärn (Allen et al. 1996). In many deposits, e.g. Kristineberg, Petiknäs South, Långsele, Rävliden, and Rävlidmyran, sphalerite is concentrated towards the hangingwall. Chalcopyrite is concentrated in the stratigraphically lower part of the ores in the latter four deposits as well as in Udden.

The Skellefte ores are all associated with sericite and/or chlorite alteration zones in greenschist facies rocks or equivalent alteration assemblages in rocks of higher grade. Almandine, biotite, cordierite, andalusite, staurolite, and cummingtonite are commonly found in lower amphibolite facies (cf. Rickard 1986). In carbonaceous host rocks actinolite, andradite, diopside, and epidote calc-silicate alteration assemblages are sometimes found (cf. Rickard 1986). The alteration envelopes are generally asymmetric with the most intense alteration in the stratigraphic footwall (Allen et al. 1996). The main alteration type is quartz-sericite-pyrite alteration which may extend from 100 m to 1 km along strike and up to 2 km into the footwall like in Långdal and Långsele (Allen et al. 1996). An inner chloritic zone exists at Långsele and Rävlidmyran, while calc-silicate rocks of a hydrothermal origin (Allen et al. 1996) are present in Rävliden, Rakkejaur, Renström, and Långdal. Allen et al. (1996) also describe alteration zones which extend into the hangingwall at Holmtjärn.

Although most deposits are situated within the uppermost part of the Skellefte Group volcanic rocks, some deposits are situated both deeper in the stratigraphy and at the base of the Vargfors Group rocks (cf. Rickard & Zweifel 1975, Rickard 1986, Allen et al. 1996). The time period for the deposition of all massive sulphide deposits is probably less than 10 million years, possibly less than 5 million years and, according to Allen et al. (1996), corresponds to a period of intense extensional arc volcanism.

Epithermal VMS-gold deposit

The Boliden Au-Cu-As deposit was one of the first discovered in the district, and it has attracted a continuous interest since then due to its significant size and high gold grade, average 15 ppm (cf. Bergman Weihed et al. 1996). The Boliden ore can be divided into massive ore, with arsenopyrite- and pyrite-dominated lenses, and vein ore comprising a quartz-chalcopyrite-sulfosalt dominated assemblage occurring in brecciated parts of the arsenopyrite bodies and quartz-tourmaline veins mainly in host rocks below the massive ore. The gold is associated with the vein ore which mainly is found in deformational structures in the massive ore.

The ore zone is vertical and oblique to lithological contacts. The alteration around the ore is symmetric with an inner sericite-rich zone, locally containing abundant andalusite, and an outer chlorite-dominated zone. The nature of the alteration is consistent with leaching of cations, leaving a silica-alumina-rich residue. The cross-cutting nature of the ore with respect to host rocks, the hydrothermal alteration pattern with strongly leached host rocks, and the ore association with early massive sulphides followed by gold, chalcopyrite, and sulfosalts in brittle structures all indicate that a modern analogue for ore formation may be a high-sulphidation epithermal environment (Bergman Weihed et al. 1996).

Gold-lode deposits

Although the massive sulphide deposits are unusually rich in gold (some even extremely rich like Boliden and Holmtjärn), it was only in 1988 that the first gold lode deposit (Björkdal) was opened for mining. This was rapidly followed by the discovery of the Åkerberg deposit which opened in 1989. Today several promising targets are evaluated mainly in the eastern- and westernmost parts of the district, namely Ö. Åkulla, Barsele, and Ersmarksberget. All known gold deposits in the Skellefte district are associated with quartz veins that vary in width from 1 mm to 1 m. The gold is nearly always associated with sulphides like pyrite, arsenopyrite, pyrrhotite, and chalcopyrite and weak alteration halos with sericite, chlorite, biotite, carbonate, and sulphides are present in many of the deposits. Shear zones and faults are present and probably related to gold deposition in Vinliden (Bergström 1996), Middagsberget and Fäbodliden (Öhlander & Markkula 1994).

The Björkdal gold deposit is situated at the contact between a granodiorite-tonalite, dated at c. 1905 Ma (cf. Billström and Weihed 1996) and the overlying volcanic rocks. The granodiorite-tonalite is cross-cut by a system of quartz veins with at least three different trends of which two are mineralised (Albino & Weihed 1997). The whole Björkdal area is strongly deformed and the area around the mine is characterised by internally well preserved megalithons which are bordered by shear zones. Mapped displacements are in the order of a few meters in the mine area. The gold in the mine is associated with brittle quartz veins which probably formed in tension fractures associated with low-angle thrusting towards the southeast. Gold has probably been transported by magmatic fluids as $Au(HS)^{2-}$-complexes. Late hydrothermal fluids redistributed the gold under high pressure and temperature in the range of 200–300 °C (Broman et al. 1994).

Gold in the Åkerberg deposit is confined to a c. 20 m wide zone of cm-wide parallel quartz veins within a gabbroic intrusion. The host gabbro probably belongs to the c. 1.89–1.87 Ga old Jörn suite and the ore is cut by late pegmatites related to 1.80 Ga Skellefte type intrusive rocks. The grade is reported at c. 3 g/t and the proven tonnage is around 1 million tonnes.

Porphyry-type deposits

Porphyry-type deposits occur in tonalitic to granodioritic plutons which are coeval with the volcanic rocks that host the VMS deposits. One of these porphyry-type deposits, the Tallberg deposit, has been studied in detail (Weihed 1992). The host for the mineralization is a tonalitic I-type, and/or magnetite series, intrusion dated at 1.89 Ga. This intrusion is chemically similar to modern volcanic-arc granitoids (VAG) and has a high primary content of Cu and Fe. In the mineralized area the tonalite is intruded by quartz-feldspar porphyries also of a tonalitic composition. These porphyries are associated with a strong hydrothermal alteration of a mixed phyllic-propylitic type. The alteration minerals are chlorite, sericite, quartz, epidote, and calcite. Ore minerals are magnetite (early), pyrite, chalcopyrite, and sphalerite which occur both in stockwork and as disseminations. Lithogeochemical investigations indicate a metal zonation with

$$(Mo+Cu+S) \rightarrow (Zn+Pb \pm Au) \rightarrow (As)$$

from the centre outwards. Late shear zones adjacent to postmineral dykes have high Au contents (2–3 g/t), which is interpreted as gold mobilized from the porphyry-type deposit.

Ni-Deposits

Ni deposits have been known in the Skellefte district since the 1930's. The Lainijaur deposit consists of massive ore at the base of, and disseminations within, a fractionated gabbroic phacolith (Martinsson 1987). The 3500 m long and up to 100 m wide NE-striking Älgliden dyke contains disseminated and partly massive copper and nickel. Deposits associated with ultramafic rocks are concentrated in a c. 100 km long WSW-striking belt with shear zones near the coast south of Skellefte. This area is dominated by migmatized metasedimentary rocks and late- to post-orogenic granites. Komatiitic to tholeiitic ultramafic volcanic rocks and small ultramafic intrusions occur within the metasedimentary rocks. Ni occurs as disseminations, in veins, or in brecciations within, and at the contacts between, the ultramafic intrusions and the country rocks but locally also entirely outside the ultramafic rocks. Most common ore minerals are pentlandite, pyrrhotite, and chalcopyrite. Nilsson (1985) interpreted the WSW-striking shear zones as reactivated older structures which may have influenced the emplacement of the ultramafic rocks and associated Ni mineralizations.

Metallogeny in relation to tectonic evolution

The submarine mainly felsic arc volcanism in the Skellefte district probably occurred over a time span of maximum 15 million years between 1890 and 1875 Ma. At the end of the volcanic stage, when the arc was under extension, VMS ores were emplaced on the seafloor or as replacement deposits in sub-seafloor environments. Porphyry-style deposits (eg. Tallberg) may have formed at c. 1.885 Ga during the early stages of volcanism by magmatic hydrothermal systems that never reached the sea floor. After extension, possibly related to accretionary processes, large areas of the submarine arc were uplifted and hydrothermal activity in shallow to subaerial settings developed epithermal VMS and gold deposits like the Boliden deposit. It is still not well understood how the ultramafic hosted Ni-deposits south of the district fit into a tectonic model, but large crustal scale shear zones may have developed during crustal thickening. These then acted as passageways for mantle-derived utramafic magmas. During peak metamorphic conditions in the Skellefte district at c. 1.83 to 1.80 Ga, deformation and metamorphism caused remobilization (and possibly addition) of gold which was precipitated in structural and chemical traps to form quartz-vein hosted gold deposits. The recognition of slightly older crust south and west of the Skellefte district, e.g. Knaften (Wasström 1993 & Geol. Surv. of Sweden, unpublished data), is of vital importance for understanding the Svecokarelian orogenic processes and hence also the metallogeny of the Fennoscandian shield.

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Episodic Crustal Magmatism in the Proterozoic of Northern Australia - A Continuum Crustal Heating Model for Magma Generation

Lesley A.I. Wyborn[1], Alison Ord[2], Bruce Hobbs[2] and Mart Idnurm[1]

[1]AGSO, GPO Box 378
Canberra ACT 2601 Australia

[2]AGCRC, CSIRO, PO Box 437
Nedlands WA 6009 Australia

A proposed conductive heating model of the crust provides an explanation of the enigma that in Proterozoic provinces of Australia the temperature of formation of the granitic magmas increases with time whilst that of the mafic magmas appears to decrease. In this model an initial heat pulse applied to the base of the crust in the early Proterozoic creates a thermal anomaly that can last for at least 80 Ma years, depending on the initial thickness and the thermal conductivity of the crust, the presence of an underplated layer, the thickness of sediment overburden and the time constant for heat input from the mantle. The migration and emplacement of magmas generated by this heating is controlled by intraplate rather than interplate events. Major mineralisation episodes associated with granites are predictable, and metamorphic deposits are expected to occur only late in the thermal history of a province.

The field and laboratory observations

The Proterozoic shield of Australia can be subdivided up into major magmatic provinces each of which occurs within major fold belts or orogenic domains. In each of these provinces magmatic events have been synchronous with basin formation (occurring either contemporaneously with the early rift phase of sedimentation or just prior to turbidite sedimentation), or they may immediately follow a major compressional event. On the basis of outcrop characteristics, thin section petrography, SiO_2 distribution and multi element, primordial-mantle-normalised trace element patterns, Wyborn et al. (1992) divided the Australian Proterozoic granites into 5 main groups. Group 1, which is usually the oldest, comprises restite-rich I-(granodioritic) types that are Sr-depleted and Y-undepleted. Typical examples are the Tennant Creek Granite of the Tennant Creek Block, the Kalkadoon Granodiorite of the Mount Isa Inlier and the granites of the King Leopold Inlier, all of which are in the age range of between 1870 to 1850 Ma. Group 2 comprises I-(granodioritic) types which, like Group 1, are Sr-depleted and Y-undepleted but, unlike the latter, show strong evidence of magmatic fractionation: they generally have low concentrations of Ca, Sr, and the incompatible elements. Examples include the granites of the Granites-Tanami region and the Cullen Suite of the Pine Creek Inlier. Group 2 granite ages are mostly around 1850-1800 Ma. Group 3 contains I-(granodioritic) types that are Sr-depleted, Y-undepleted and enriched in incompatible elements. Granites of this group are widely referred to as 'anorogenic'. Wyborn et al. (1992) divided Group 3 into three subgroups. Subgroup 3_1 has very high levels of Zr, Nb and Y and includes felsic volanics of the Argylla Formation of Mount Isa and the Myola Volcanics of the Gawler Craton, both emplaced at around 1800-1780 Ma. Subgroup 3_2 is enriched in F, but the levels of incompatible elements such as Y, Zr, and Nb vary from high in some members to low in others. This group is characterised by a narrow SiO_2 range. Examples include the Sybella Granite of the Mount Isa Inlier and the Mount Swan and Ennugan Mountains Granites of the Arunta Block, which range in age from 1760 Ma to 1640 Ma. Subgroup 3_3 is not as enriched in F as Subgroup 3_2 and has a high $Fe_2O_3/(FeO + Fe_2O_3)$ and a wide SiO_2 range. Granites of the two other groups identified by Wyborn et al (1992) are rare in the Proterozoic of Australia. Group 4 is the rarest of the groups and consists of I-(granodioritic) types, but is distinctly Sr-undepleted and Y-depleted - a signature which is most commonly found in granites associated with subduction in island arc or continental margin settings. Group 5 comprises the S-type granites. It is significant that Groups 1, 2 and 3 are all I-(granodioritic) type and are dominated by the Sr-depleted, Y-undepleted magmas.

Significance of the dominance of I-(granodioritic) types in the Australian Proterozoic

Igneous events in the Proterozoic are either predominantly mafic or felsic, or consist of both types: igneous events dominated by intermediate rocks do not exist (Wyborn et al., 1987). Histograms of the igneous rocks of each province are bimodal and contrast the more unimodal plots of subduction-related terrains in which the SiO_2 maxima are at either 52 wt % (island arcs) or ~65 wt % (continental margins). In a global review of I-type granites, Chappell and Stephens (1988) classified them into three types: (1) M-type derived from partial melting of subducted oceanic crust or of mantle material overlying oceanic slabs in island arc terrains; (2) I-(tonalitic) type, which is characteristic of the Cordillera of South America, and is probably derived from M-type rocks of basaltic to andesitic composition; and (3) I-(granodioritic) type which is believed to be formed from I-(tonalitic) type. Proterozoic granites are predominantly I-(grano-dioritic) type which, by inference, are derived in at least one if not two stages of melting. Further, in view of the silica distribution, the I-(granodioritic) types must be derived by melting of pre-existing crust, rather than from a mantle source. This is supported by Sm-Nd data which give Nd TDM model source ages that are predominantly in the range 2.1 to 2.8 Ga. However, there are no signif-icant amounts of I-(tonalitic) type material in the exposed rocks of this age in Australia, it has been postulated that the source(s) of the voluminous Proterozoic I-(granodioritic) types was underplated as a result of Archaean to earliest Proterozoic mantle events (Wyborn et al., 1992).

Significance of the Sr-depleted, Y-undepleted signature

The multi-element, primordial-mantle-normalised abundance diagrams contain two distinct patterns. These are either Sr-depleted and Y-undepleted or are Sr-undepleted and Y-depleted, implying that the source has equilibrated with plagioclase and garnet respectively. As I-(granodioritic) type granites have a multistage history then those that are Sr-depleted and Y-undepleted could never have equilibrated with garnet during the melting event(s). Proterozoic crustal thicknesses range from 30 to 50 km (Drummond and Collins, 1986). Therefore the geothermal gradients must have been extremely high during these melting event(s), otherwise a garnet residue would be expected. Those I-(granodioritic) types with a Y-depleted signature must have equilibrated with garnet at some stage in their history. However, this signature may not necessarily relate to the contemporaneous melting event: it could well be the signature of the event that formed the source for that magma. The Y-depletion signature could imply that the magmatism occurred in a contemporaneous subduction zone. Alternatively, it could signify a lower geothermal gradient, or other differences in the conditions that operated at the time of forming the source(s).

Implications of the changes in composition with time

Within each of these major Australian magmatic provinces the inferred crustal temperatures in the source regions of the granites increase with time. The restite-rich granites of Group 1 are inferred to form early by the break-down of quartz, albite, K-feldspar and water at the source: biotite and hornblende are restite within these melts and there are abundant xenoliths. Group 2 consists of the fractionated granites that are low in incompatible elements. These granites have rare restitic amphiboles and it is possible that in their source regions dehydration melting of biotite has occurred. The Subgroup 3_1 granites are probably formed by dehydration melting of F-enriched biotites, whilst Subgroups 3_2 and 3_3 may involve dehydration melting of amphibole. Temperatures of formation of some of the Group 3 granites have been inferred to be as high as 1000°C. The systematic progression through these groupings indicates that the source region temperature has increased progressively with time.

In contrast, temperature of formation of the mafic igneous rocks generally appears to decrease with time, with the mafic igneous rocks from 2000 to 1870 Ma dominated by high Mg-tholeiites (Woodard Dolerite of the Kimberleys and the Zamu Dolerite of the Pine Creek Inlier), whilst continental tholeiites dominate after 1870 Ma (Hart Dolerite of the Kimberleys, Oenpelli Dolerite of the Pine Creek Inlier and the Eastern Creek Volcanics of the Mount Isa Inlier). The exceptions are the ~1700-1650 Ma high Fe-tholeiites of Broken Hill and the Soldiers Cap Group of the Mount Isa Inlier.

Duration of individual magmatic events

Australia-wide Palaeoproterozoic to Mesoproterozoic magmatic events have an episodic distribution with time. There is a period of quiescence from about 2400 Ma to 1880 Ma, followed by a series of major magmatic events from about 1880 Ma to

1500 Ma, which are in turn followed by a period of inactivity until about 1350 Ma. Although within any one province the period of magmatism is 70 Ma to 360 Ma, individual magmatic events are episodic. Each episode is of relatively short duration, generally less than 20 Ma and some episodes with tight age control are less than 10 Ma.

Role of tectonic processes in triggering the emplacement of magmas

Magmatism is commonly regarded as being caused by major interplate events such as subduction of a cold oceanic crust in either a continental margin or island arc environment, or impingement of a mantle plume on the lower crust. Other cited causes are extension (either within a continent or at oceanic ridges) and continent/continent collision. A significant feature of the Australian Proterozoic is the continent wide abundance of magmatic activity. If all of the magmatism occurred at interplate boundaries then the sheer number of magmatic events would require a large number of small plates. Alternatively many of the magmatic events may be the products of intraplate activity.

Intraplate events may be caused by changes in the direction or rate of plate motion (Loutit et al., 1994) rather than by plate collisions or within-plate extensional episodes. Intraplate events are regarded as the key in the evolution of sedimentary basin development and deformation (Green et al., 1992) , and have been held responsible for major episodes of mineralisation and fluid migration. For example, in northern Australia, a significant intraplate event at 1640 Ma has been suggested to have been responsible for major movements of basinal brines which produced regionally extensive alteration and resulted in the formation of the world class HYC Pb-Zn deposit (Idnurm et al., 1993; Loutit et al., 1994). Although this event resulted in large scale fluid migration, it did not produce regionally extensive angular unconformities. Similarly, intraplate events may provide the clue to the abundance of magmatic episodes in the Proterozoic of Australia.

Palaeomagnetic data have proved useful for determining the time of the major intraplate events. The apparent polar wander path (APWP) for northern Australia displays two modes of absolute plate motion. The first is characterised by periods of relative kinematic stability, represented by straight or nearly straight segments on the APWP. The second mode is characterised by instabilities that reflect geologically rapid changes in plate motion. This mode is represented on the APWP by hairpin bends and points of inflection. A detailed APWP is not available for Australia from ~2000 Ma to 1720 Ma, but the APWP from ~1720 Ma to ~1500 Ma indicates a series of relatively long periods when the direction of plate motion remained constant, separated by intervals of changes of direction. As noted by Loutit et al. (1994), most major magnetic overprints due to alteration, several major Pb-Zn deposits, and major magmatic events coincided in time with the hairpin bends and inflection points on the APWP for Northern Australia. For example, the HYC 1640 Ma event described above coincides with a major 180° bend in the APWP. Loutit et al.(1994) and Green et al. (1992) have argued that the periods of greatest migration of basinal fluids coincide in time with these bends. By the same argument, significant volumes of magmas could have also escaped to the surface at such times. Thus magmatic events may not necessarily result directly from major interplate interactions, but rather from intraplate tectonism. These intraplate events may reflect forces acting a large distance from the major plate boundaries. Palaeomagnetism provides insights also into the episodic nature of magmatism. It appears from the shape of the APWP that magma emplacement coincided in time with the hairpin bends or inflection points. However, the melting of the lower crust/mantle did not necessarily result from the same tectonic processes as the emplacement of the magmas, as that would not allow sufficient time to heat the crust to the melting temperature. Instead subsequent intraplate tectonism may have allowed already existing melts to migrate into the upper crust. Intraplate tectonism may have also enhanced melting by causing decompression.

Alternative causes for the melting events

Recent thermal modelling may provide new insights into the mechanism of generating melts within the lower crust. If heat is applied to the base of the crust as a result of a major mantle heating event, then the crust will heat up by conduction. The thermal time constants for crust 25, 30, 40 and 50 km thick are 19, 27, 48 and 76 Ma (Upton et al., 1997) and hence, the time lag between early mantle magmatic activity and then crustal metamorphism and major crustal melting is of the order of a minimum of 25 Ma and given Proterozoic crustal thicknesses could be at least 70 Ma if not longer. Modelling has also shown that the addition of sediments and thickening of the crust will cause further rises in temperature.

The proposed model shows that it takes considerable time for sufficient heat to be conducted to the crust to generate the types of magma consistently observed in the Proterozoic of Australia. Thus the tectonic processes that helped generate the melt are not necessarily the same as those that allowed for its migration into the upper crust. Intraplate forces acting on the crust can facilitate the escape of magmas to the surface, with the composition of the melt being simply dependant on the temperatures existing in the lower crust at the time of the 'escape' rather than on the contemporaneous tectonic setting.

Metallogenic implications

A consequence of the present model is that, in any province it is predictable that mineralisation derived from plutonic fluids will be associated with Group 2 and Group 3_3. Group 1 granites being restite-rich, cannot give a greater concentration of any element than that contained in the initial melt or restite and hence are rarely mineralised. Similarly the fluorine rich granites (Groups 3_1 and 3_2) are rarely mineralised. In contrast Groups 2 and 3_3 that have been formed by dehydration melting of biotite and hornblende respectively appear to have the greatest metallogenic potential. Granites of type 3_3 are rare in the Proterozoic of Australia having only been documented in the Gawler and Mount Isa provinces where they are associated with significant Cu+Au±U mineralisation. The rarity of the granite-types may reflect the difficulty in actually attaining these high temperatures in the lower crust. Further, the conductive model predicts that as the thermal anomaly will only reach mid crustal and higher levels late in the history of the province, metamorphic deposits will occur late in the history of any province.

Conclusions

The present model, which is preliminary, goes a long way to explaining the observation that in any Proterozoic province, granites are emplaced at progressively higher temperatures with increasing time. The model does not require subduction and implies instead that much of the Proterozoic magmatism may result from intraplate activity. Further modelling will be carried out to more fully determine the actual length of time that the thermal anomalies can exist in the crust, and what influence variables such as the original crustal thickness, the thickness of the sedimentary overburden, the composition of these sediments (viz. conductivity contrast between black shale vs pure quartz sand), and the thickness of the mafic underplated layer can have on the process.

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