

CHAPTER 5: MAGMATIC PROCESSES INVOLVED IN URANIUM DEPOSIT FORMATION

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INTRODUCTION

Magmatic processes entirely controlled uranium and other incompatible element fractionation during the early history of the Earth. Uranium in silicate magmas exhibits a strongly incompatible behaviour because of its large ionic radius and high valence, which prevents its incorporation into the structure of the main rock-forming silicates. As a result, during partial melting and crystal fractionation, U is preferentially fractionated into silicate melts. Such a behavior has several major geochemical, geophysical and metallogenic consequences:

(i) through geological time, U and other incompatible elements such as Th and K have been continuously transferred from the mantle to the Earth crust, and within the continental crust towards its upper part,

(ii) radiogenic heat production is maximized in the upper crust, and thus the distribution of radiogenic heat flux production may be used to delineate radioelement enriched crustal blocks (see Chapter 4),

(iii) felsic melts tend to be the most enriched in U, and thus granites and rhyolites represent primary sources of U for the formation of deposits.

However, despite the strongly incompatible behaviour of U, deposits dominantly resulting from magmatic processes are rare. In an average granitoid, with a U enrichment of 3 to 4 ppm, U is dominantly held within the crystal structure of accessory minerals such as zircon, apatite, monazite, titanite and xenotime from which U cannot be leached by most geological fluids.

The first mechanism of U enrichment were partial melting of the primitive mantle (U = 20.5 ppb, Table 5-1) followed by crystal fractionation during ascent of melts to the surface. Fractionation of basic melts did take place during the Hadean and Paleoproterozoic times, but to a limited extent because the Earth was relatively hot. Before the Mesoarchean (<3.2–3.1 Ga), these processes led to the formation of a relatively thin and dominantly mafic crust, composed essentially of komatiitic and tholeiitic basalt.

TABLE 5-1. AVERAGE U, TH, AND K ABUNDANCES IN THE EARTH'S MAJOR COMPONENTS AND IN CHONDRITES

	U (ppb)	Th (ppb)	K (ppm)	Th/U
Upper continental crust	2,800	10,700	28,000	3.8
Average continental crust	1,020	4,800	21,900	4.7
Primary upper mantle	20.5	85	40–80	3.92
Mid-ocean ridge basalt	47	120	600	2.55
Ocean island basalt	1,020	4,000	12,000	3.92
CI carbonaceous chondrites	7.4	29	545	3.92

Primary upper mantle and average continental crust U and Th (Javoy 1995) and K (Allègre 1987), upper crust (McLennan *et al.* 2001), mid-ocean ridge basalt, ocean island basalt, CI: carbonaceous chondrites.

Felsic rocks in the early Earth belonged to the classic tonalite-trondhjemite- granodiorite (TTG) suite with generally less than a few ppm U.

At such low concentrations, U essentially enters into accessory minerals, and uraninite saturation of the melts was never attained. Uraninite has never been described in pre-Mesoarchean plutonic rocks. The first uraninite-bearing granites appear during the Mesoarchean (Cuney 2010).

MAGMATIC DIFFERENTIATION

Uranium deposits related to magmatic fractionation processes may occur through extreme fractional crystallization, mostly of peralkaline magmas as well as by partial melting of U-enriched supracrustals. Extreme fractional crystallization of peralkaline magmas may lead to the formation of very large, low-grade U and Th resources, such as the Kvanefjeld deposit at Ilímaussaq, Greenland (Sørensen 2001, Bohse *et al.* 1974). Other major occurrences of this type are: Poços de Caldas, Brazil (Fraenkel *et al.* 1985), Bokan Mountain, Alaska (MacKevett 1963), Lovozero massif, Kola Peninsula, Russia (Balashov 1968), and the Kaffo Valley, Nigeria (Bowden & Turner 1974). Such an association can be also extended to the ultimate fractionation products of peralkaline complexes, namely carbonatites such as those found at Palabora, South Africa (Verwoerd 1986). The association of peralkaline complexes with U and Th enrichments reflects the high solubility of U and Th in high-temperature peralkaline silicate melts (see Chapter 3). Peralkaline granites and syenites are always enriched in U, Th, and other high field strength (HFS) elements such as Zr and REEs, as a result of the high solubility of these elements in highly depolymerized magmas and the low degrees of partial melting that produce peralkaline liquids. Their high solubility leads to continuous enrichment during magmatic fractionation. However, simultaneous enrichment of U, Th, Zr, REEs, Th, Nb, and Ta can lead to crystallization of complex Zr, REE, Th, Nb, and Ta minerals that can incorporate U. Uranium will not be able to crystallize independently as uraninite, or if so, only in very small amounts. Therefore, even if the U content of some deposits related to peralkaline melts may be relatively high, they have not generally been mined because of the high cost of uranium extraction from highly refractory minerals. The Kvanefjeld deposit in the Ilímaussaq peralkaline complex (Sørensen *et al.* 1974) represents one of the best examples of such a uranium deposit, where U is mainly hosted by

steenstrupine, a complex silico-phosphate of U, Th, and REEs.

Another consequence of the strongly incompatible behaviour of U together with Zr, REEs, Th, Nb, and Ta in peralkaline magmas is the association of uranium mineralization with the most fractionated part of peralkaline complexes. These rocks are also generally located in the more apical part of the complex or at its margin, where low-viscosity residual melts and exsolved fluids are emplaced. Although the fluid/melt partition coefficients are extremely low in peralkaline systems, in some occurrences, U, Th, and the REEs may be transported a few hundreds of meters (e.g. Bokan Mountain) or over several kilometers (e.g. Ilímaussaq and Th veins of the Front Range) from the intrusion. None of these veins has sufficient grade or tonnage to be mined.

From a geodynamic point of view, peralkaline intrusions are emplaced in post-orogenic to anorogenic environments. They tend to form circular intrusions, generally concentric and associated with ring dikes, because of their emplacement by cauldron subsidence in a consolidated basement lacking shear tectonics. They form very large to small plutonic complexes. They may represent a wide range of intrusions from gabbros to highly-fractionated granites, are always emplaced at a very high structural level and are commonly associated with contemporaneous volcanism.

The major uranium deposits related to fractional crystallization are the following:

- *Kvanefjeld (Greenland)*: 108,850 t U @ 240 ppm U + Zn, Be, F, Nb, Th
- *Ghurayyah (Saudi Arabia)*: 45,700 t U @ 120 ppm U + Ta, Nb, Y, Zr
- *Poços de Caldas (Brazil)*: 22,700 t U @ 720 ppm U + U, Mo, Zr, Th
- *Bokan Mountain (Alaska)*: production 1000 t U @ 1% U

Ilímaussaq (South Greenland)

Regional geology. The Ilímaussaq intrusion is a peralkaline, silica-undersaturated plutonic complex situated in the eastern part of the 1.1–1.3 Ga Gardar intracratonic rifting province of South Greenland (Fig. 5-1). Magmatism in the Gardar Province is related to two main rifting events at 1280 Ma and 1180 to 1140 Ma. A series of 10 major plutonic complexes of gabbroic and nepheline-bearing to quartz-saturated granitoid

rocks intrude a basement consisting of Paleoproterozoic granites and gneisses, which were unconformably overlain by Mesoproterozoic Gardar basalts and sandstones of the Eriksfjord Formation (Fig. 5-1). Two of the intrusive complexes contain carbonatites, and many alkali gabbroic dikes throughout the Gardar province contain anorthosite xenoliths (Bridgwater & Harry 1968). The association of alkali gabbro intrusions, anorthosite xenoliths, and carbonatites with granitoid intrusions has been interpreted to reflect large-scale melting processes of asthenospheric mantle. The alkali basaltic melts were ponded at the crust-mantle boundary and would have given rise, by fractionation, to massive-type anorthosites and later to alkaline or peralkaline melts in the roof region of these magma chambers (Larsen & Sørensen 1987). A U-Pb age of 1160 ± 5 Ma on baddeleyite from the first magma batch (Krumrei *et al.* 2006) indicates that the intrusion belongs to the second period of rifting and is the youngest of the province. The igneous complex covers about 156 km^2 , and comprises an extraordinary diversity of granitic and syenitic rock types, emplaced as three successive melt batches, which intruded and fractionated at a depth of 3 to 4 km (Larsen & Sørensen 1987).

Stage I corresponds to a mildly alkaline augite syenite which occurs both as a few-hundred-metres-thick shell along the W, S, and SE margins of the intrusion, and as a 150 m-thick layer on top of the intrusion (Fig. 5-1). The contacts between augite syenite and the nepheline-sodalite syenite (stage III) may locally appear gradational, but the sharp cross-cutting relationships occurring at other places demonstrate that the augite syenite is older.

Stage II corresponds to a small unit of quartz syenite and alkali granite, which represents a break in the magmatic evolution of the complex. This unit occurs at the top of the intrusion.

Stage III corresponds to the intrusion of nepheline- and sodalite-bearing syenite, presumed to be derived from the same deep magma chamber as the stage I intrusion. Stage III starts with Si-undersaturated, slightly peralkaline rocks, evolving toward nepheline syenites. Within these syenites, minor volumes of pulaskite, foyaite and sodalite have crystallized in situ from the roof downwards. The sodalite-rich syenite represents a 600 m-thick flotation cumulate of sodalite (Ferguson 1964). Chemically and texturally

distinct varieties of nepheline syenites formed below these roof rocks. Small bodies of medium- to coarse-grained lujavrite (hereafter indicated as m-c lujavrite), the most U mineralized rocks, intrude arfvedsonite lujavrites, but are thought to represent a more primitive stage of the lujavrite magmas (Andersen *et al.* 1981). The m-c lujavrites occur in the uppermost structural level of the peralkaline complex, just below the Gardar supracrustals (Fig. 5-1).

The mineral assemblages of these rocks are described by Ferguson (1964, 1970), and Sørensen (2001). The early magmatic and intercumulus assemblage may partially be overprinted by later magmatic minerals and, in most places, by a post-magmatic or hydrothermal assemblage. The mineralized lujavrites typically contain nepheline, eudialyte, sodalite, and clinopyroxene crystals enclosed in a mixture of albite and microcline, whereas aegirine and amphibole appear texturally late. Black lujavrite is rich in arfvedsonite and green lujavrite is rich in aegirine. Minor constituents include Ussingite $\text{NaAlSi}_3\text{O}_8 \cdot \text{NaOH}$, naujakasite $\text{Na}_6(\text{Fe}, \text{Mn})\text{Al}_4\text{Si}_8\text{O}_{26}$, steenstrupine $\text{Na}_{14}\text{Ce}_6\text{Mn}_2\text{Fe}_2(\text{Zr}, \text{Th}, \text{U})(\text{Si}_6\text{O}_{18})_2(\text{PO}_4)_7 \cdot 3\text{H}_2\text{O}$ and villiaumite (NaF).

Geochemistry. The lujavrites represent the most fractionated melts and therefore are the most enriched in incompatible elements. Most lujavrites, are cumulates and thus their geochemistry is not always representative of melt compositions. In a Q-P diagram (Fig. 5-2), the first unit (augite syenite) is close to quartz saturation, the rocks from the second unit (quartz syenite and granite) are quartz-saturated and the richest in potassium, and the third unit of various types of nepheline syenite is strongly quartz-undersaturated and variable in composition. The most sodic members tend to be the most silica-undersaturated, the naujaite being the most extreme because of its strong enrichment in sodalite. In an A-B diagram (Fig. 5-3), the first unit is moderately meta-aluminous and is the only one to have an agpaitic ratio lower than one (Table 5-2).

The second unit is peralkaline $(\text{Na}+\text{K})/\text{Al} = 1.1$ to 1.2). The third unit is strongly peralkaline $(\text{Na}+\text{K})/\text{Al}$ up to 2.0). The abundance of Fe-Ti-minerals tends to decrease with increasing peralkalinity, especially in layered lujavrites.

The K/Rb ratios gradually decrease during fractionation from 500 in the augite syenite to 35 in the most evolved lujavrites, whereas Zr/Hf ratios continually increase, from 45 to 97. Total REE contents and LREE/HREE ratios increase during fractionation (Bailey *et al.* 2001) whereas elements such as Zr, Cl, F, Zn, and W show irregularities or maximums in some rock types that are related to cumulus processes (Bailey *et al.* 2001). Thorium and U contents increase, from the augite syenites to the m-c lujavrites, with a Th/U

ratio close to 3 (Fig. 5-4). However, the other types of lujavrites have the lowest Th/U ratios at less than 1.

Larsen and Sørensen (1987), propose fractionation at depth, possibly at the crust–mantle boundary, of the melts parental to the Ilímaussaq rocks. Successive ‘leaking’ of this deep-level magma chamber is responsible for the three magma batches rising to the high-level chamber, where further low-pressure, closed-system crystal fractionation continues.

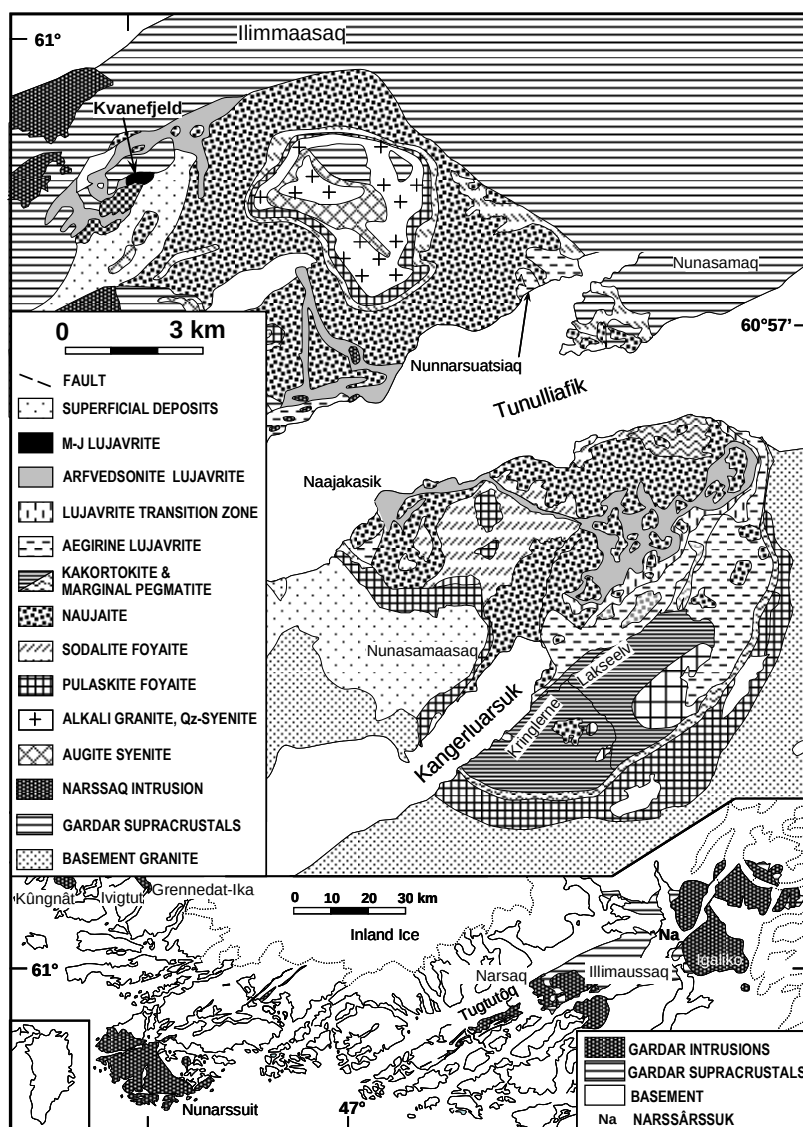


FIG. 5-1: Geological map of the Ilímaussaq peralkaline complex (South Greenland), showing the location of the Kvanefjeld deposit (modified from Sørensen and Larsen 2001). Lower insert is the simplified geological map of the Gardar alkaline province with the location of the Ilímaussaq intrusion.

TABLE 5-2: AVERAGE CHEMICAL COMPOSITION OF THE MAJOR ROCK TYPES IN THE ILIMAUSSAQ COMPLEX

Sample	1	2	3	4	5	6	7	8	9	10	11	12	13	14
n	6	5	5	8	11	5	9	13	11	5	11	10	8	3
SiO ₂ wt%	55.82	56.14	63.59	72.01	60.34	58.5	49.98	46.32	48.6	52.22	52.57	54.09	52.63	53.68
TiO ₂	2.29	1.64	0.32	0.28	0.19	0.32	0.30	0.40	0.44	0.35	0.20	0.18	0.20	0.37
Al ₂ O ₃	14.94	16.15	14.53	10.52	16.53	16.16	17.96	20.20	8.47	10.93	15.76	15.93	13.71	13.22
Fe ₂ O ₃	2.03	1.47	3.03	2.65	3.38	3.07	4.77	3.39	8.26	5.61	5.82	8.29	4.55	3.48
FeO	8.82	7.64	4.07	3.18	2.42	4.38	4.57	4.11	14.84	5.72	4.36	0.92	7.36	8.23
MnO	0.24	0.22	0.18	0.14	0.18	0.21	0.25	0.19	0.47	0.40	0.26	0.22	0.55	0.61
MgO	1.01	1.35	0.15	0.14	0.14	0.16	0.13	0.13	0.48	0.28	0.21	0.11	0.12	0.16
CaO	3.47	3.88	0.96	0.40	1.53	2.02	1.83	1.89	2.48	3.58	1.92	0.69	0.34	0.25
Na ₂ O	5.53	5.16	5.41	4.88	8.19	7.56	11.55	14.63	8.67	10.43	9.55	11.52	10.18	7.91
K ₂ O	4.25	4.83	6.60	4.52	4.94	5.31	3.76	3.48	2.51	3.02	4.69	2.95	3.47	5.40
P ₂ O ₅	0.74	0.52	0.02	0.02	0.05	0.05	0.05	0.03	0.02	0.04	0.03	0.03	0.44	0.47
I,L,	0.67	0.62	0.50	0.45	1.26	1.47	3.06	1.73	2.47	2.70	2.47	2.88	4.29	3.48
Total	99.81	99.62	99.36	99.19	99.15	99.21	98.21	96.50	97.71	95.28	97.84	97.81	97.84	97.26
P	-150	-133	-51	-68	-186	-167	-325	-432	-271	-336	-243	-321	-261	-145
B	0	-4	27	142	-52	-56	-197	-311	-93	-153	-139	-142	-114	-75
Q	202	179	102	84	82	107	130	108	327	161	141	121	165	167
A	-100	-91	-64	-61	-100	-112	-166	-217	-255	-314	-167	-146	-145	-120
Na+K/Al	0.92	0.85	1.10	1.23	1.14	1.13	1.28	1.38	2.00	1.87	1.32	1.39	1.50	1.43
Rb (ppm)	69	97	438	481	323	301	395	332	193	314	525	662	685	1280
La	71	74	252	325	223	236	540	409	393	1330	555	1410	3300	2830
Th	4.1	5.7	53	79	38	26	54	33	15	46	43	55	89	466
U	1.3	1.7	16	27	9.4	7.5	15	9.6	6	21	17	58	178	490
Zr	288	431	1950	2035	2230	1870	3200	3610	5830	20100	7610	7250	4130	724
Nb	55	92	312	291	327	250	546	580	495	1890	779	648	445	484
Be	3.5	3.2	28	22	16	14	31	21	15	28	26	32	46	104
F	990	1340	3310	3850	2270	2170	3090	4740	10200	6160	7070	502	1310	1140
Th/U	3.15	3.35	3.31	2.93	4.04	3.47	3.60	3.43	2.50	2.19	2.53	0.95	0.50	0.95

From Bailey (2006). Unit I: 1) fine-grained augite syenite 2) coarse-grained augite syenite; Unit II: 3) quartz syenite , 4) alkali granite; Unit III: 5) pulaskite, 6) foyaite, 7) sodalite foyaite, 8) naujaite, 9) black kakortokite, 10) red kakortokite, 11) white kakortokite, 12) aegyrine lujavrite, 13) arfvedsonite lujavrite, 14) medium- to coarse-grained lujavrite.

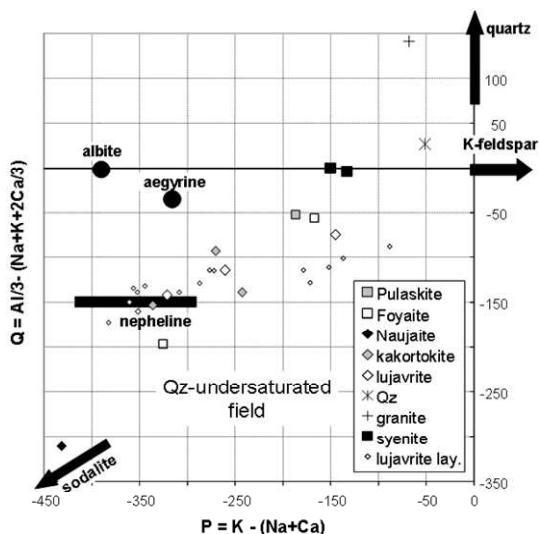


FIG. 5-2: Variation of P versus Q parameters for units from the Ilímaussaq peralkaline complex. Q-P diagram from Debon & Lefort (1983), parameters in thousands of cations. The values for the main rock-forming minerals are also indicated (large symbols are average data from Sørensen *et al.* 1974; intermediate size symbols are from Bailey 2006; smallest symbols are layered lujavrite from Bailey *et al.* 2006).

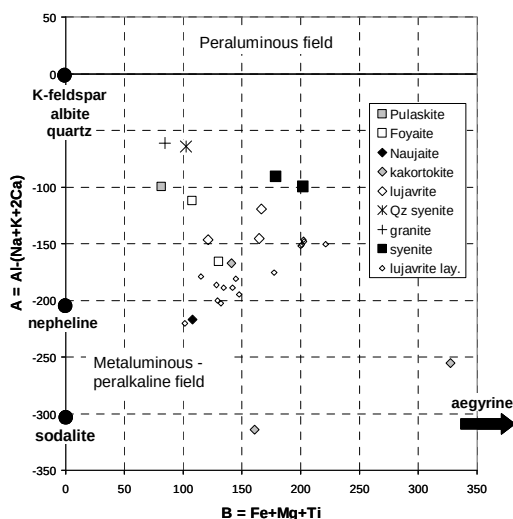


FIG. 5-3: Variation in the peraluminous index A with fractionation parameter B for the Ilímaussaq peralkaline complex igneous units. A-B mineralogical-chemical diagram from Debon & Lefort (1983), parameters in thousands of cations (same symbols as Figure 5-2).

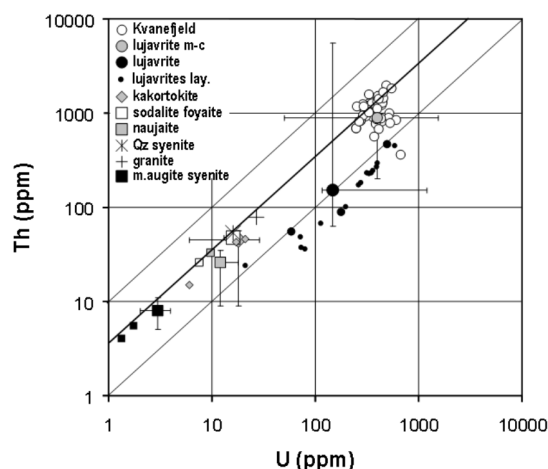


FIG. 5-4: Th-U variations in plutonic units of the Ilímaussaq intrusion (same symbols as Fig. 5-2), and in the Kvanefjeld deposit (Clausen 1982).

Fluid inclusion studies have shown that water and methane are important fluid phases in equilibrium with the peralkaline melts (Konnerup-Madsen *et al.* 1985, 1988). Higher hydrocarbons also occur in small percentages, whereas highly saline fluid inclusions are abundant in the alkali granite. Pressures of 1 kbar have been obtained from fluid inclusion data for the alkali granite (Konnerup-Madsen & Rose-Hansen 1984) and from the estimated 3 to 4 km thickness of the overlying Eriksfjord Formation. Primary fluid inclusions in sodalite crystals from naujaite give temperatures between 700 and 900°C and pressures between 1.5 and 3.5 kbar, reflecting their crystallization during the ascent of the magma. Temperatures of crystallization from 450–500 to 1000°C have been reported for nephelines and pyroxenes (Markl *et al.* 2001).

Mineral compositions (Markl *et al.* 2001) show that the earliest augite syenite melt at Ilímaussaq was reduced, with oxygen fugacity (f_{O_2}) almost two orders of magnitude below the FMQ buffer, methane was stable, water activity was very low (<0.2), relatively low silica activities (0.8), and relatively high alkali/Al ratios. Subsequently, high-level fractionation of augite syenitic melts led to a drastic reduction of silica activities to 0.4 and a strong decrease of f_{O_2} , down to four orders of magnitude less than the FMQ buffer. Further deep-level fractionation also leads to decreasing silica activities and increasing (Na+K)/Al ratios, thus generating the peralkaline rocks of the syenitic magma batch.

The anorthosite xenoliths found throughout the Gardar province may reflect these deep-seated processes. Subsequently, when the melt (Na+K)/Al ratio became greater than 1, crystallization of alkali feldspar and nepheline, with a (Na+K)/Al ratio of 1, further increased the ratio of the melt. At this stage, fluid inclusions show that methane is still the stable fluid phase, but fO_2 increases significantly, to two orders of magnitude above FMQ, whereas silica activity drops to 0.3 in the late stages of the last peralkaline magma batch. When these rocks are intruded into the red Mesoproterozoic sandstones, the latter turn green as a result of hematite reduction.

In lujavrites, fO_2 is no longer controlled by a solid–solid buffer reaction and fO_2 can vary. The transition from a methane-dominated to a very low-salinity aqueous fluid-dominated system in the late lujavrite crystallization stages appears to occur at fO_2 close to the magnetite–hematite buffer and silica activity drops to values lower than 0.2. This explains why most Fe in the final lujavrite is trivalent, and U reaches its highest concentration. Karup-Møller (1978) used ore minerals from the Ilímaussaq lujavrites and pegmatites to suggest that fO_2 was between 10^{-21} and 10^{-16} at 500°C. These fO_2 are 1 to 4 orders of magnitude above the magnetite–hematite buffer. By changing solid-phase buffers from FMQ to arfvedsonite–aegirine, alkaline and peralkaline melts are intensely reduced in the early stages of crystallization, but highly oxidized in the late stages. These calculations also support a model of successive magma batches derived from a single continuously fractionating magma at depth.

Uranium Mineralization. The Kvanefjeld U deposit was discovered in 1956, one year after the initiation of a regional uranium exploration program. It is situated at the northwestern edge of the intrusion (Fig. 5-1). The deposit consists of magmatic mineralization formed during the latest emplacement stages of the intrusion. The uranium mineralization is associated with the youngest nepheline syenite, and with the medium-coarse lujavrite, which is generally dark, shows a pronounced magmatic layering, and is enriched both in U and Th, as well as in Nb, Zr, Be, Li, F, and REEs. In the Kvanefjeld area, under a roof of gabbro and basic volcanic rocks, a mega-breccia of volcanic and sedimentary rocks, together with rocks from the early phases of the intrusion are

cemented by lujavrite. Three main types of lujavrite occur: the first was a fine-grained arfvedsonite lujavrite, followed by a naujakasite lujavrite and by the medium-coarse lujavrite. The U content of the arfvedsonite lujavrite and the medium-coarse lujavrite is about 150–200 ppm. In contrast, the naujakasite lujavrite contains several zones with 300–500 ppm U. The highest U and Th values are generally observed in the upper part and at the contact of the medium-coarse lujavrite with enclosing altered volcanics intruded by fine-grained lujavrite. The thickness of the mineralized zones vary from a few metres up to about 50 m. Analcime–steenstrupine veins also intersect altered volcanics. High-grade veins and pegmatites are not of economic importance because of their small volume, but their U and Th contents vary widely (117 – 1200 ppm U and 63 – 5500 ppm Th).

Two U enrichment trends can be observed in the Kvanefjeld mineralization (Fig 5-5):

(i) a magmatic trend with an average Th/U ratio of about 3.5, similar to the main fractionation trend defined by the less fractionated units of the complex and

(ii) a fluid fractionation trend with a U enrichment at constant Th content, and a Th/U ratio decreasing to less than 2.

These two trends and the large variations in U and Th contents observed in the different types of lujavrites reflect the effects of magmatic and late- to post-magmatic processes related to exsolution of a fluid phase.

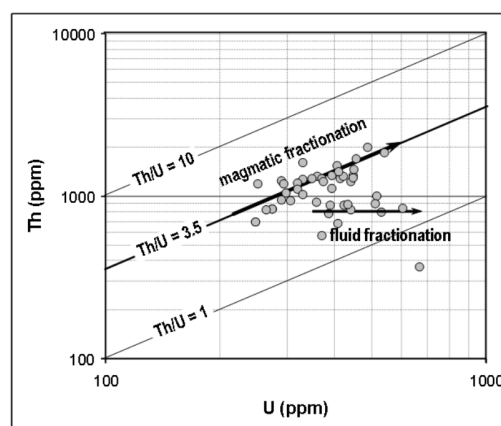


FIG. 5-5: Relations between Th and U in the Kvanefjeld mineralization (data from Clausen 1982) indicating two types of fractionation.

TABLE 5-3: U AND TH CONTENTS OF THE MAJOR ROCK TYPES IN THE ILIMAUSSAQ PERALKALINE COMPLEX AND MAIN U BEARING MINERALS

Rock type	U minerals	U (ppm)	Th (ppm)	Th/U
augite		3	8	2.7
syenite		(2-4)	(5-11)	
foyaite	eudyalite	16	48	3.0
		(13-19)	(43-53)	
naujaite	eudyalite	12	26	2.5
		(6-27)	(9-35)	
kakortokite	eudyalite	18	45	2.5
	rinkite	(6-29)	(9-57)	
lujavrite	eudyalite	146	154	1
	monazite	(117-1200)	(63-5500)	
	lovozerite			
	steenstrupine			
lujavrite m-c	steenstrupine	400	890	2.2
	monazite	(50-1550)	(200-1100)	

From Sørensen *et al.*, (1974)

Numerous mineralized veins, up to several km long, striking NE-SW, perpendicular to the main axis of the Ilimaussaq peralkaline complex, and predominantly vertical, transect all rock types to the east of the Ilimaussaq intrusion (Hansen 1968). Three main types of veins are mineralized:

(i) *brown albitic veins*, composed of albite, hematite, quartz, and small amounts of chlorite, calcite, apatite, monazite, eudyalite, and thorite;

(ii) *green veins*, composed of albite and aegirine, occasionally with microcline, hematite, and some accessories, such as mica, apatite, chlorite, quartz, sphalerite, monazite, bastnaesite $\text{CeCO}_3(\text{OH}, \text{F})$, neptunite $\text{KNa}_2\text{Li}(\text{Fe}, \text{Mn})_2\text{Ti}_2\text{Si}_8\text{O}_{24}$, and thorite;

(iii) *composite veins*, with the components from the two other types, with accessory eudyalite $(\text{Na}_{12}\text{Ca}_6\text{Fe}_3\text{Zr}_3[(\text{Si}_3\text{O}_9)_2(\text{OH}, \text{Cl})_2])$ and thorite.

Nearly all veins have elevated and highly variable Th/U ratios (1 to 57), as well as Th (60 to 4500 ppm) and U (17 to 1500 ppm) contents, indicating that Th has been mobilized by a fluid and deposited in the veins over several kilometres. However, U has migrated further, as reflected by low Th/U ratios of the veins, because of the oxidizing conditions recorded by the presence of hematite in the veins. The source of these veins is certainly the Ilimaussaq peralkaline

complex, as they have been traced to the contact with the complex and similar veins have been observed within the complex.

Steenstrupine, is an important rock-forming mineral in the lujavrites and the main U mineral in the medium-coarse lujavrites of the Kvanefjeld uranium deposit. Steenstrupine occurs instead of eudyalite, the main U- and Th-bearing mineral in the other rocks of the complex (Table 5-3 and 5-4). Steenstrupine is also common in the sodic pegmatites, in the hydrothermal mineralization, and in the fenitised volcanics of the roof of the Kvanefjeld plateau. In the large Lovozero complex of the Kola Peninsula, Russia, steenstrupine is also common in the zeolite-rich cores of the pegmatites and in hydrothermal veinlets (Khomyakov 1995, Semenov 1997). Uranium and thorium contents of steenstrupine crystals vary from 0.1 to 1.4 wt% U and 0.2 wt% to 7.4 wt% Th (Hansen 1968, Makovicky *et al.* 1980) (Table 5-4). The central part of these crystals is metamict because they are richer in U, Th, and Y than the non-metamict crystal margins, whereas ZrO_2 , MnO , Na_2O , and P_2O_5 are higher in the margins (Table 5-5). The decrease of U and Th contents in the marginal parts reflects the progressive fractionation of U and Th into the exsolving fluid phase to produce the uranothoriferous mineralization in the fenitised volcanic rocks of the roof of the complex (Sørensen *et al.* 1974).

The preferential loss of uranium during fluid oversaturation of the lujavrite melts may also explain the occurrence of small pitchblende veins in the Proterozoic granite close to Narssaq, Narssarsuaq, and Julianehåb in South Greenland (Armour-Brown *et al.* 1982). The 1180 Ma age obtained for the pitchblende compared to 1160 Ma obtained for the Ilimaussaq intrusion suggests that this uranium mineralization is derived from hydrothermal activity associated with the emplacement of the Ilimaussaq peralkaline complex.

The high $f\text{O}_2$ developed in the latest intrusion should cause a preferential loss of uranium as $(\text{UO}_2)^{2+}$, as indicated by the high Th/U ratios observed in the upper part of the drill hole 42 through the Kvanefjeld deposit (Fig. 5-6). Deposition of uranium elsewhere is indicated by the fluid fractionation trend, with uranium enrichment at nearly constant thorium content in the whole-rock geochemical data of the Kvanefjeld tunnel (Fig. 5-5). At the scale of

steenstrupine crystals, the reverse evolution is observed (Table 5-5), but the higher Th/U ratio in the centre of the crystal relative to its rim results from the central part of the mineral, which is the richest in U and Th, being metamict and therefore having lost a large part of its initial U.

TABLE 5-4: U AND TH CONTENTS OF THE MAIN RADIOACTIVE MINERALS OF THE ILIMAUSSAQ PERALKALINE COMPLEX

	U (ppm)	Th (ppm)
Steenstrupine	2,000-15,000	2,000-74,000
Monazite	138-12,800	300-57,300
Thorite	31,000	405,000
Britholite		6,900
Eudyalite	55-600	30-340
Pyrochlore	2,000	7,000
Rinkite	3,000-12,000	

From Sørensen *et al.*, (1974)

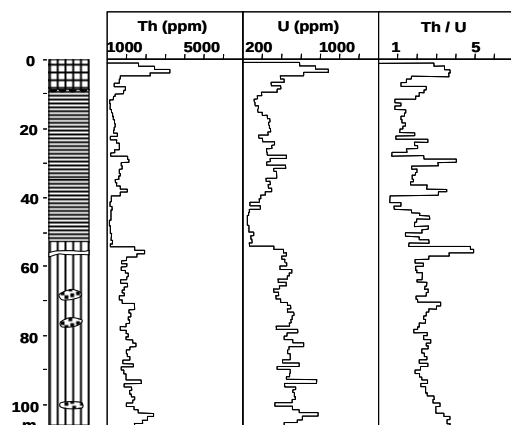


FIG. 5-6: Distribution of U and Th and Th/U ratios along drill hole 42 through the Kvanefjeld deposit (from Sørensen *et al.* 1974). Vertical stripes correspond to the fine-grained lujavrite with naujaite enclaves; horizontal stripes correspond to altered volcanics; gridded box corresponds to altered volcanics with m-c lujavrite injections.

Resource estimations. In the 1970s, detailed exploration at Kvanefjeld, with 11852 m of core from 76 drill holes, led to a total estimated resource of 43,000 t U at 340g/t U for a 250g/t U cut-off, with 27,000 t U of reasonably assured resources and 16,000 t U of estimated additional resources. Further exploration from 1979 to 1986 resulted in the identification of 60,000 t U of speculative resources. According to the International Nuclear Energy Agency, untested

mineralized host rocks could add significantly to these estimates. In addition to U, other mineralization includes Be (180,000 t at 0.1% Be) and Li (235,000 t at 600-1900 ppm). Estimates of U potential using an average U concentration of 100 to 200 ppm over 12 km² and a thickness of 600 m, would correspond to 3 Mt U (Bondam & Sørensen 1959). The new estimations provided by Greenland Mineral and Energy Ltd (<http://www.ggg.gl/projects/kvanefjeld-rees-uranium-zinc/>), including newly-discovered mineralized fields, give a Rare Earth Oxide (REO) resource of 10.33 Mt at 1.08 % REO with 221,000 t U, and additional resources of zinc.

The recovery cost of U by sulfatizing roast was \$40 to \$70 /kg of U, depending of the type of ore (Gamborg-Hansen 1977). For the Carbonate Pressure Leaching process, the recovery cost was \$70 to \$90 per kg of U. About 5000 tons of ore were extracted from the Kvanefjeld tunnel driven through the ore body in 1980-81.

Bokan Mountain (USA)

Regional geology. The Bokan Mountain U-Th deposit at Kendrick Bay, southeastern Alaska, is associated with a Late Jurassic peralkaline granite. The pluton is post-tectonic and intrudes Paleozoic metasediments and metavolcanics, as well as quartz diorites and quartz monzonites, at the western side of the Chatham Strait fault (Fig. 5-7). The area belongs to the Alexander Terrane, located at the Eastern margin of the Canadian Cordillera (Churkin & Eberlein 1977). A U-Pb age on zircon of 171 ± 5 Ma has been obtained for the Bokan Mountain granite (De Saint André *et al.* 1983), corresponding to a major episode of thrust tectonics in the North American Cordillera. However, as peralkaline magmatism is typically emplaced in stable continental crust in anorogenic settings, the Alexander Terrane was probably accreted to the North American continent more recently (Collot 1981).

Geology of the pluton. The pluton is circular, zoned, and forms a topographic high. It outcrops over 3.2 km², but a larger complex is presumed at depth. The ring structure is defined by a core of aegirine granite forming the main part of the pluton, surrounded by an aegirine-only granite and a border zone pegmatite (Fig. 5-8).

TABLE 5-5: MICROPROBE ANALYSES OF STEENSTRUPINE CRYSTALS FROM A VEIN WITH 350 PPM U AND 2,000 PPM TH SAMPLED IN DRILL CORE 50 FROM THE KVANEFJELD DEPOSIT (FROM KHOMYAKOV & SØRENSEN 2001). OH AND F HAVE NOT BEEN ANALYZED.

Sample wt %	50-149.7 centre (5)	50-149.7 rim (6)	50-150.6 centre (2)	50-150.6 rim (1)	199104 centre (4)	199104 rim (1)
SiO ₂	26.94	27.75	27.54	25.98	27.40	27.06
TiO ₂	0.16	0.07	0.28	0.18	0.25	0.20
ZrO ₂	n.a.	n.a.	n.a.	n.a.	1.41	2.33
ThO ₂	3.95	0.42	4.13	1.40	5.04	1.83
Al ₂ O ₃	n.a.	n.a.	n.a.	n.a.	0.28	0.04
La ₂ O ₃	8.78	12.85	8.94	14.22	11.49	11.79
Ce ₂ O ₃	14.57	15.53	13.80	13.76	15.74	16.49
Pr ₂ O ₃	1.29	1.04	1.28	1.54	2.39	2.08
Nd ₂ O ₃	3.72	2.80	3.50	2.78	4.01	4.22
Sm ₂ O ₃	0.31	0.19	0.19	0.15	n.a.	n.a.
Y ₂ O ₃	1.47	1.05	1.56	0.31	1.05	0.10
FeO	4.34	2.50	3.96	1.85	3.87	3.42
MnO	3.61	6.22	3.42	5.69	4.42	5.00
CaO	2.18	1.58	2.00	1.65	1.88	1.95
SrO	0.10	0.14	0.07	0.11	n.a.	n.a.
Na ₂ O	8.41	12.98	7.88	13.65	2.75	6.59
K ₂ O	0.05	0.02	0.20	0.31	n.a.	n.a.
P ₂ O ₅	9.36	11.26	8.50	9.35	11.31	12.30
Nb ₂ O ₅	0.10	0.24	0.13	0.67	n.a.	n.a.
U ₃ O ₈	n.a.	n.a.	n.a.	n.a.	0.70	0.35
Total	89.34	96.64	87.38	93.60	93.89	95.75
ΣRE ₂ O ₃	28.67	32.41	27.71	32.45	33.63	34.58
La ₂ O ₃ /Nd ₂ O ₃	2.36	4.59	2.55	5.12	2.86	2.79
Th/U	-	-	-	-	7.46	5.42

n.a. = not analyzed.

Fine-grained aegirine granite xenoliths, preserved in the central upper part of the pluton, probably represent the roof of the pluton, as attested by the occurrence of enclaves of host metamorphic rocks. However, no magmatic contacts have been observed between the various units and the petrologic variations seem to result from an internal evolution related to differences in crystallization conditions. This interpretation is confirmed by the lack of geochemical differences between various units (see below). Felsic dikes in the host rocks are rich in Ta, Nb, U, Th, REE, and Zr, and therefore probably derive from the Bokan Mountain intrusion.

The main rock-forming minerals are euhedral quartz, microcline with abundant perthite, nearly pure albite, aegirine, and arfvedsonite. Albite occurs as individual crystals, perthites, along the growth zones of quartz,

coronas around microcline crystals, and along microfractures in microcline. There is a progressive transition between purely magmatic crystallization of albite and late- to post-magmatic growth from a fluid phase. Arfvedsonite and aegirine crystals are poikilitic, enclosing quartz, albite and microcline crystals. Zircon is ubiquitous, abundant, and generally xenomorphic, suggesting a late crystallization stage. Fluorite is very common. Astrophyllite, a typical mica ((K,Na)₃(Fe²⁺,Mn)₇Ti₂Si₈O₂₄(O,OH)₇) of the peralkaline intrusions, only occurs in the fine-grained unit. Magnetite and hematite are related to alteration.

An early stage of microfracturing, occurring at a sub-magmatic stage ("protoclasis" of Collot 1981) typically affects the granite and results from an overpressure produced during the exsolution of the vapor phase from the magma.

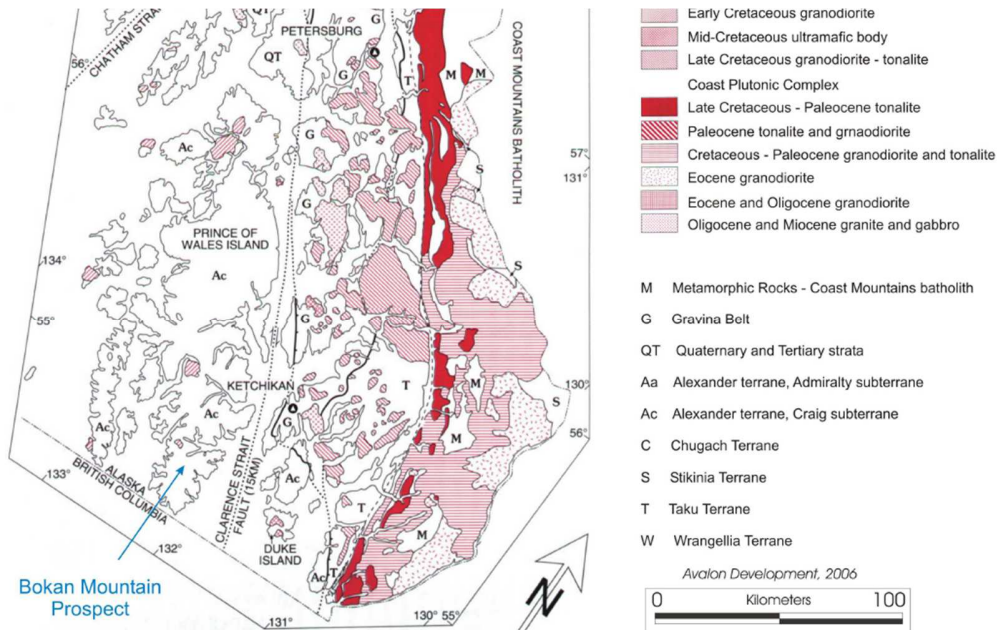


FIG. 5-7: Geologic map of SE Alaska with the location of the Bokan Mountain granite and of the major terranes (from Churkin & Eberlein 1977).

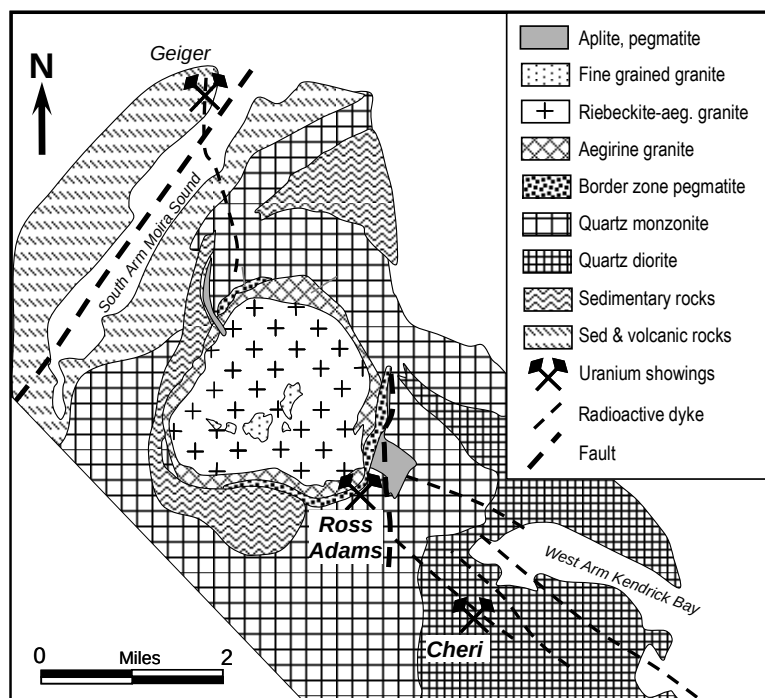


FIG. 5-8: Geology of the Bokan Mountain peralkaline complex area and location of the main uranium showings (modified from MacKevett 1963).

TABLE 5-6. SELECTED GEOCHEMICAL ANALYSES OF THE BOKAN MOUNTAIN GRANITE AND MINERALIZATION

wt%	293	430	494	275	290	549	276	411	464	473	541	550	554	578	524	534	538	567
SiO ₂	73.7	73.6	74.6	73.5	74.9	72.4	72.6	73.2	74.5	73.8	74.3	73	72.9	70.4	66.4	61.9	67.3	65.4
TiO ₂	0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.01	0.03	0.01	0.02	0.02	0.03
Al ₂ O ₃	11	10.7	11	11.3	10.3	12.2	12.6	12.3	11.6	11	11.6	11.6	11.3	13	14	12.4	11.3	14.4
Fe ₂ O ₃	5	5.6	4.6	4.7	4.8	4.1	3.4	3.5	3.7	5.2	4.4	4.7	4.4	4.9	2.7	6	6.1	5.8
MnO	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.05	0.04	0.03	0.02
MgO	0.017	0.015	0.028	0.017	0.017	0.03	0.033	0.033	0.017	0.015	0.015	0.025	0.03	0.032	0.139	0.221	0.086	0.028
CaO	0.035	0.105	0.133	0.07	0.056	0.364	0.07	0.056	0.024	0.032	0.02	0.136	0.137	0.155	2.2	0.532	0.161	0.057
Na ₂ O	4.6	4.9	5.7	5.3	4.5	5	4.8	5.2	5.1	6.5	8.8	8.4	9.4	7.4	8.8	9.5	7.2	11.1
K ₂ O	4.4	4	3.1	4	4.2	4.7	5.3	4.2	4.4	2.4	0.1	0.1	0.1	0.2	0.5	0.4	0.1	0.2
P ₂ O ₅	0.05	0.02	0.04	0.02	0.01	0.05	0.02	0.03	0.01	0	0.02	0.04	0.01	0.15	0.15	0.18	0.15	0.04
Total	98.82	98.96	99.22	98.94	98.80	98.86	98.84	98.54	99.37	98.97	99.28	98.03	98.30	96.28	94.95	91.19	92.45	97.08
P	-56	-77	-123	-88	-58	-74	-45	-80	-72	-160	-282	-274	-306	-240	-352	-317	-236	-356
Q	167	164	163	151	181	136	135	149	155	149	126	131	98	146	48	22	137	0
B	63	70	58	59	60	52	43	45	47	65	55	59	56	62	37	81	78	73
A	-27	-37	-39	-37	-34	-35	-23	-18	-31	-46	-59	-50	-89	7	-98	-91	-18	-82
F (ppm)	320	820	1020	1320	980	3300	810	1140	1130	1140	140	1140	1050	1040	10200	1800	530	420
U	5,6	2,2	17	14	6	17,5	7,6	26	14,5	10,5	24	81	91	1300	6900	17400	6100	820
Th	9	15	40	37	11	56	13	34	27	18	32	4590	3710	2150	3240	19400	27000	1180
Zr	1086	1370	1125	578	2221	642	722	1185	946	1076	134	772	1300	5129	762	2714	2072	8466

(Collot 1981, unpub. data). Analyses 293, 430, 494: albite – aegirine granite; 275 290, 549: albite – aegirine – arfvedsonite granite; 276, 411, 464: fine-grained albite – aegirine – arfvedsonite granite; 473, 54: non- mineralized albitized albite – aegirine granite; 550, 554, 578; mineralized albitized albite – aegirine granite; 524, 534, 538, 567; mineralized albitized and dequartzified albite – aegirine granite. P, Q, A, B: chemical – mineralogical parameters (in thousands of cations) used and defined in the diagrams.

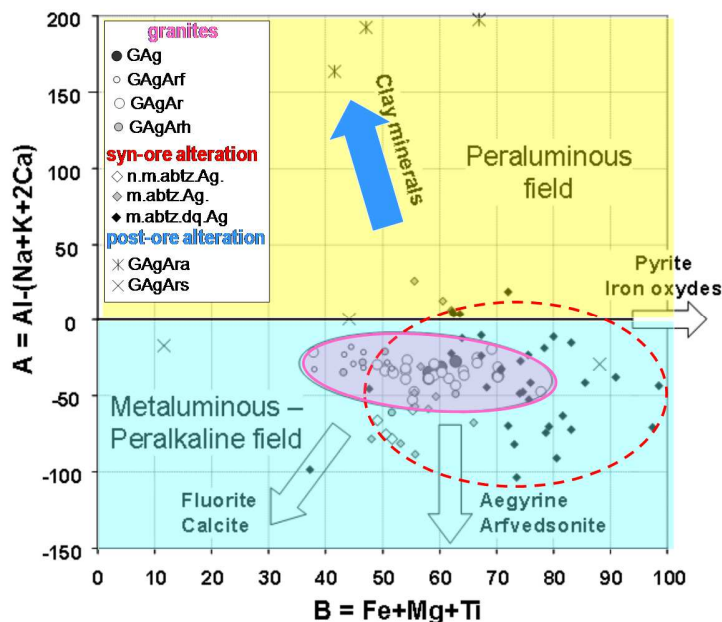


FIG. 5-9: Variation of the peraluminous index (A) with fractionation (B) for the Bokan Mountain igneous units and altered zones. The field of unaltered granites is circled. Diagram from Debon and Lefort (1983), parameters in thousands of cations. Granite types: aegirine (GAg), aegirine-arfvedsonite (GAgAr), fine-grained aegirine-arfvedsonite (GAgArf), heterogeneous (GAgArh), silicified (GAgArs), argilized (GAgAra). Altered and mineralized bodies: non-mineralized albitized aegirine (n.m.abtz.Ag); mineralized albitized aegirine (m.abtz.Ag); non-mineralized albitized and dequartzified aegirine (m.abtz.dq.Ag).

Geochemistry. The Bokan Mountain granitic pluton is typically peralkaline with a negative peraluminous index (Table 5-6, Fig. 5-9). The granites are also rich in iron (4.6 wt % Fe_2O_3 on average), poor in calcium (0.09 wt % CaO on average, essentially hosted in fluorite), and extremely poor in magnesium (0.01 wt% on average). The main chemical variations correspond to the relative abundance of albite and K-feldspar, with a tendency for larger amounts of quartz in the most potassic samples (Fig. 5-10). Mafic mineral content is relatively high and variable, especially in the aegirine–arfvedsonite albitic granite. No variation of chemical composition is evident from the major element contents that can be attributed to magmatic fractionation trends.

The granites are moderately rich in Th and U (24 and 12 ppm, respectively). Thorium is well correlated with U, with a slight decrease of Th/U ratios from about 4 to 2 (Fig. 5-11) with increasing Th and U concentrations, but without any correlation to petrographic units. The Zr contents are elevated, averaging 1040 ppm (150 to 2200 ppm), but Zr contents in unaltered granite are not correlated with Th (Fig. 5-12) nor with F, which are elevated and extremely variable (200 – 3200 ppm).

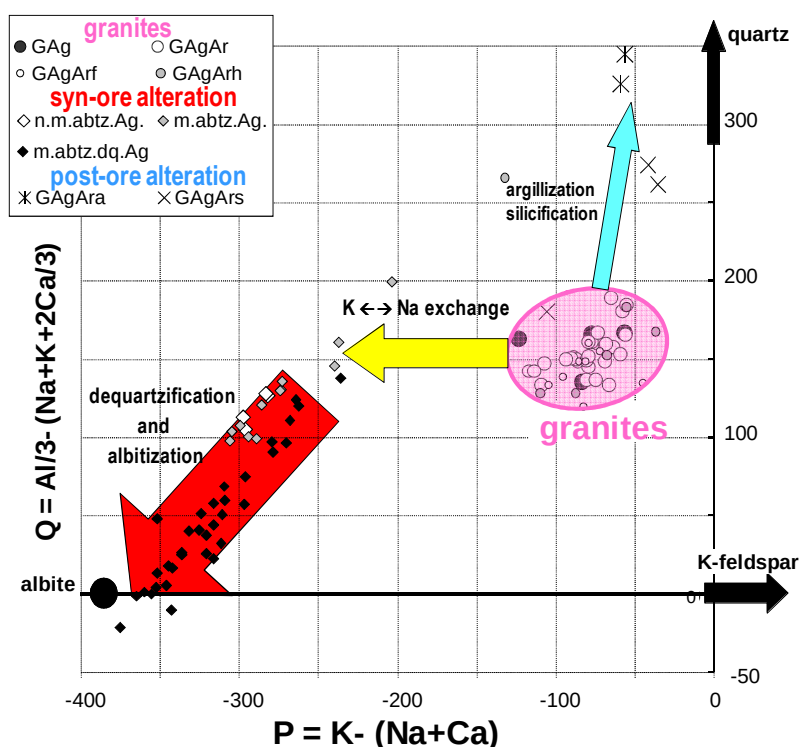


FIG. 5-10: Variation of the K-feldspar/albite (P) parameter versus quartz contents (Q) for the Bokan Mountain igneous units and altered zones. The field of unaltered granites is circled. Q-P mineralogical-chemical diagram from Debon & Lefort (1983), parameters in thousands of cations. Same legend for the symbols as in Fig. 5-9.

Uranium mineralization. The Ross-Adams U deposit was discovered by D. Ross in 1956. It was mined from 1957 to 1971, first from open pit and later as underground working by a series of companies (Climax Molybdenum, Bay West Inc., Standard Metals, and Newmont Exploration Ltd) during the 1950s and the 1960s. Standard Metals and R. Dotson (a prospector) discovered a NW-trending U vein and shear system SE of the Ross-Adams deposit – the Dotson Shear – extending over 2.5 km to the shoreline. A U vein and shear system, named the I and L deposit, was also discovered adjacent to the Ross-Adams area in the early 1970s, and represents the fault-displaced NW extension of the Dotson shear-vein system. Although significant U grades were defined at the I and L deposit, it remains unmined. High contents of REEs, Nb, and Zr are associated with some of the U zones (MacKevett 1963; Thompson 1988; Staatz *et al.* 1979).

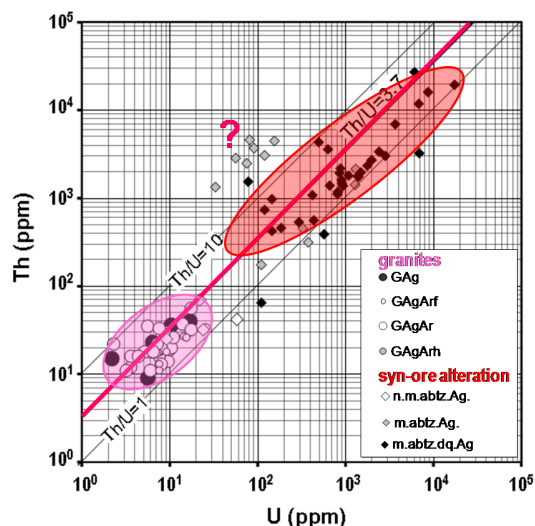


FIG. 5-11: Relation between Th and U in different plutonic units of the Bokan Mountain peralkaline complex. Same symbols as in Figure 5-9.

Several types of U mineralization have been recognized by MacKevett (1963) in the area: (i) the main type of U-Th mineralization, the Ross-Adams deposit, is associated with a desilicified and albitized part of the pluton forming plunging pipe-like bodies along the contact with the aegirine granite or occurring as pods in en echelon NW-striking shear zones. The ore minerals are uranothorite and uraninite, with

generally less than 2% of sulfides disseminated in nearly pure albite;

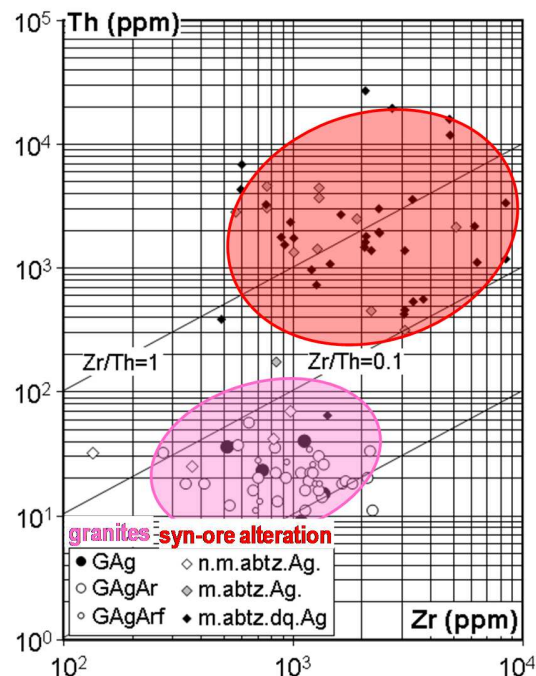


FIG. 5-12: Relation between Th and Zr in the different plutonic units of the Bokan Mountain peralkaline complex. Same symbols as in Figure 5-9.

(ii) some disseminated U mineralization also occurs in NW-oriented, steeply-dipping albite-quartz dike systems and associated shear zones peripheral to the complex;

(iii) mineralized pegmatite veins also occur within a marginal phase of the complex;

(iv) uranium vein type mineralization;

(v) matrix replacement phases in hornfelsed and hydrothermally-altered sandstones adjacent to, or overlying, the intrusion.

Nearly all U occurrences are in or near the granite stock. A total of 89,000 t of ore have been mined at a grade of about 0.85 % U and 2.5 % Th with an average Th/U close to 3, mainly from the Ross-Adams deposit.

The Ross-Adams ore body is situated at the SE margin of the aegirine granite. It is up to 24 m wide and was mined along strike for over 300 m. The pod-like ore zones have dense microfracturing produced during late magmatic devolatilization of the magma chamber.

Geochemical data indicate that alteration of the host aegirine granite proceeds in two main steps (Fig. 5-10). First, albitization of K-feldspar

produced typical chess-board structure in the crystals, at nearly constant quartz content. The rocks affected by this type of alteration are only weakly, or not, mineralized. This was followed by progressive dissolution of quartz with simultaneous precipitation of albite, resulting in albitites. During alteration, the contents of Al- and Fe-minerals (B parameter) do not change significantly in the rocks (Fig. 5-9). Variation in the peraluminous index is due to the combined effect of incipient clay alteration in some samples, increasing the peraluminosity of the rock because of Na or Ca leaching, and to the formation of minor amounts of Ca-bearing, Al-free minerals such as fluorite and carbonates, which decreases the peraluminous index value. The mafic mineral content of the syenites tends to be higher, because the volume loss from quartz dissolution was not totally compensated for by albite formation and there was formation of secondary pyrite and iron oxides. The U-Th mineralization is essentially hosted in these desilicified and albitized sections of the granite, predominantly as uranothorite, forming 30- to 500 μm -long crystals disseminated in chessboard albite. Zirconium is not significantly enriched during the mineralization process, as its concentration only doubled. The fluorine content of the mineralized rocks is slightly higher than that of the granites, but a large part may have been lost with fluids.

Veins represent the second most important uranium mineralization at the Ross-Adams deposit. The largest ones extend to more than 2.5 km away from the Bokan Mountain pluton and strike WNW. The most important veins are 25 to 250 m long and form a cluster up to 130 m thick in the vicinity of the pluton where aplites are abundant. Pegmatite dikes of aegirine granite occur within these veins. Compared to the intragranitic Ross-Adams mineralization, the exogranitic mineralization is characterized by the association of large amounts of REEs and Nb together with Th, U, and Zr. The main REE-bearing mineral is bastnaesite ($[\text{LREE}]\text{FCO}_3$). In the veins, the quartz-aegirine ore lenses are up to 3 m thick and 30 m long. U grades defined by drilling vary from 0.14 to 1.92 % U. Wall rock alteration within, and adjacent to, the ore bodies consists of pervasive albitization with lesser amounts of chlorite, fluorite, calcite, quartz, sericite, and tourmaline. Hematite is present as an envelope around the ore zones. Chlorite is heavily developed within higher-grade ($> 0.43\%$ U)

zones, while pervasive hematite alteration is more commonly associated with areas of lower grade ($<0.43\%$ U) (Thompson 1988).

Genetic model. By comparison with other world occurrences of peralkaline plutonic complexes, the Bokan Mountain granite represents the most fractionated, and most apical part of a larger magmatic complex existing at a relatively shallow depth. Similar albite-aegirine-arfvedsonite granites are known in many part of the world, with the best-known being those from Nigeria (Bowden & Turner 1974). Uranium mineralization associated with this type of granite is most likely derived directly from the mantle. Although no Sr-Nd-O isotopic data are available on the Bokan Mountain granites, peralkaline magmas are considered to reflect low degrees of partial melting of the mantle with subsequent fractionation during ascent of the melts. They have $^{87}\text{Sr}/^{86}\text{Sr}$ ratios lower than 0.705 and $^{18}\text{O}/^{16}\text{O}$ ratios lower than 7 to 8 (Javoy & Weiss 1987). The initial enrichment of the mantle in incompatible elements, and especially in U and Th, through the subduction of an oceanic slab with variable amounts of sediments derived from alteration of the continental crust, may considerably enhance the final U and Th concentrations in the melts that are generated. It is not possible to generate peralkaline melts from crustal material, which is generally peraluminous, except if evaporites can be incorporated. Contrary to most disseminated uranium mineralization associated with pegmatoids in anatectic domains, the Th/U ratios remain close to the chondritic ratio of 3.5 during magmatic fractionation, even in the mineralized albitites.

The peralkaline nature of the melts and their high temperatures have allowed continuous and simultaneous fractionation of U and Th during magmatic fractionation. However, the unaltered granites only have moderate U (12 ppm in average) and Th (24 ppm in average) enrichment, despite their association with late-magmatic U mineralization. Such an inconsistency can be resolved if the outer part of the pluton crystallizes rapidly and magmatic fractionation has to proceed inwards. Therefore, the most fractionated melts, richer in U, Th, and fluids, occur in the deep central part of the pluton.

At a certain stage of fractional crystallization, fluid saturation of the melts occurred, leading to the deposition of a major part

of the U and Th mineralization. Expulsion of fluids was focused in a zone of weakness, principally at the southeastern margin of the pluton, producing intense microfracturing and alteration of the granite. The alteration started with albitization of K-feldspar and, with increased fluid percolation, simultaneous quartz dissolution and new albite growth occurred, leading to the formation of albitic episyenites. At a pressure lower than 1 kbar, dissolution of quartz may occur when a fluid cools from magmatic temperatures of 700-600°C down to about 350°C, because the quartz solubility curve passes through a maximum at about 350°C under low pressure conditions. According to the inward fractionation hypothesis proposed above, the fluids are exsolved from a hotter and deeper-seated part of the pluton and migrated towards the marginal and cooler part of the pluton, with the cooling of the magmatic fluid allowing the dissolution of quartz. Similar mineralization may have also existed at the level of the roof of the pluton, but was eroded away.

Part of the fluids enriched in F, U, Th, Nb, and REEs escaped into the host rocks along veins, some of them extending several km from the pluton. The filling of these veins started at the magmatic stage as indicated by the occurrence of aplitic-pegmatitic veins of aegirine granites within them.

Thompson (1988) reported only aqueous fluids in the mineralized zones with homogenization temperatures between 320 and 331°C, corresponding to temperatures of at least 420°C when pressure corrections are taken into account. These inclusions probably represent relatively late fluids. Unpublished observations of the first author indicate three-phase inclusions containing a solid and variable proportions of water and carbon dioxide (5 to 70 vol%) in the residual quartz of mineralized albitites. The variable proportions of the latter two phases may be an indication of a fluid unmixing. Some of the fluid inclusions in fluorite also contain several birefringent solids. In contrast to Ilimaussaq (Konnerup-Madsen *et al.* 1985), the carbon dioxide phase is devoid of methane or higher hydrocarbons. Unmixing of water and carbon dioxide fluids may have enhanced the overpressurization of the fluids responsible for the microfracturing of the granite in the ore zone.

Oxygen ($\delta^{18}\text{O}_{\text{H}_2\text{O}}$ of 6.8 to 8.1‰) and carbon ($\delta^{13}\text{C}$ of -4.3 to -7.0‰) isotope data support a magmatic origin for the calcite associated with the U-Th deposits (Thompson 1988). Sulfur isotope analyses on pyrite, galena, sphalerite, and pyrrhotite indicate disequilibrium, but $\delta^{34}\text{S}_{\text{H}_2\text{S}}$ is estimated at 7.6 ‰ (Thompson 1988), also consistent with addition of a marine sedimentary component for the S in the source of the melt.

Other occurrences of albitic episyenites.

Numerous occurrences of mineralized or barren albitic episyenites have been reported all over the world, mainly in highly potassic calcalkaline granites (A2 type granites) and more rarely in peraluminous granites (S-type granites), where the episyenites generally correspond to simple quartz dissolution. Unlike at Bokan Mountain, most of these occurrences result from hydrothermal fluid circulation disconnected from magmatic crystallization of the granite. Examples include the U mineralized episyenites of the high-K calcalkaline Kab Amiri granite, Central Eastern Desert, Egypt (Abdel-Monem *et al.* 1994), the barren episyenites from amphibole-bearing and biotite high-K calcalkaline granodiorites and monzogranites from the Central Iberian Massif, Spain (Recio *et al.* 1997), the Sn mineralized episyenites from the peraluminous topaz, highly potassic calcalkaline granite of Agua Boa, Amazonas, northern Brazil, (Costi *et al.* 2002), the U-Zr mineralized episyenites related to the highly potassic calcalkaline Kurupung granite of Guyana (Cinelu & Cuney 2006), the barren episyenites from the high-K calcalkaline Bohus granite of Sweden (Gerasimovsky *et al.* 1997), the slightly U-Th-REE mineralized episyenites of Les Bombes, south-east French Massif Central, in the high-K calcalkaline granite of La Borne (Cathelineau 1987), and the barren episyenites of the Salvezines peraluminous granite, French Pyrénées, (Boulvais *et al.* 2007).

Other world occurrences of mineralized peralkaline granites.

Albite-riebeckite granites of Kaffo Valley,

Nigeria, have many similarities with the Bokan Mountain (Orajaka 1986). They are post-tectonic intrusions of peralkaline character with vein and disseminated uranium deposits.

The Ghurayyah deposit is located in Saudi Arabia, 55 km from the Dhuba harbour along the Red Sea. Mineral resources at Ghurayyah were estimated in 2006 at 45,700 t U at an average grade of 120 ppm associated with important Ta, Nb, ZrO₂, Y, REE resources. Average Th content is 399 ppm. Rare metal mineralization is associated with radioactive peralkaline microgranite intrusions in the complex. The peralkaline subcircular granite complex to which these microgranites belong was emplaced during an extension event which lasted from 620 to 530 m.y. (Stoeser 1986). The Ghurayyah Granite is characterized by typical subsolvus feldspars, arfvedsonite and aegirine (Drysdall *et al.* 1984). The mineralized microgranite presents a wide range of accessory minerals disseminated in its groundmass, including zircon, pyrochlore,

samarskite, aeschinite, columbite-tantalite, cassiterite, xenotime, monazite, uraninite, thorite, ilmenorutile, galena, sphalerite and other sulphide minerals (Lalande 1977).

Poços de Caldas U-Mo-Zr mineralization (Brazil). Poços de Caldas (Sao Paulo and Minas Gerais states) is one of the largest peralkaline complexes of the world, extending over 800 km² (Schorscher & Shea 1992). Similar to Ilimaussaq, it is composed of a silica-undersaturated igneous suite. The igneous complex forms a nearly perfect circular ring structure (caldera), about 28 km in diameter (Fig. 5-13). It is emplaced in the Guaxupé block, between the Paraná basin and São Francisco craton. Widespread alkaline-carbonatite and alkaline magmatism with both potassic-ultrapotassic and sodic affinities occurred in the

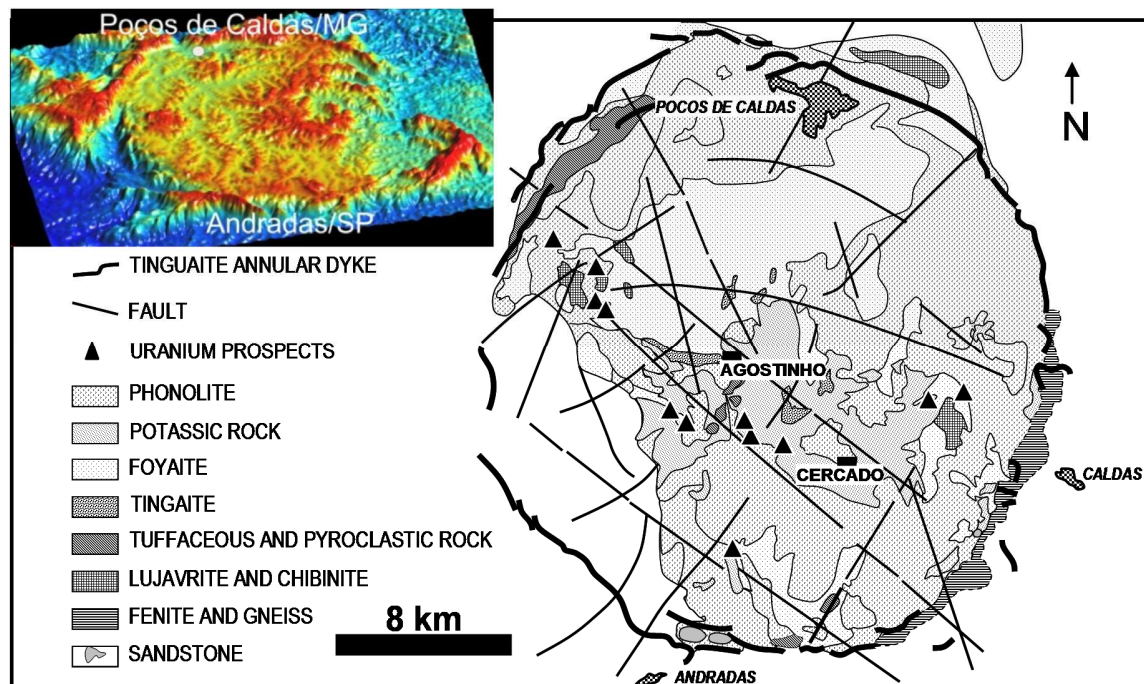


FIG. 5-13: Geologic map of the Poços de Caldas peralkaline complex with the location of the main uranium ore deposits and showings (modified from Fraenkel *et al.* 1985). Inserted the radiometric map of the complex.

area from 90 to 54 Ma. Ulbrich *et al.* (2002) proposed a date of 79 Ma as the age for the emplacement of the main nepheline syenites and phonolites of the Poços de Caldas complex, with a duration of about 1-2 Ma for the entire felsic magmatism. Two ⁴⁰Ar/³⁹Ar ages (75.7 ± 0.3 and

76.2 ± 0.8 Ma) for phlogopite from a fresh lamprophyric dike cross-cutting hydrothermally-altered nepheline syenites and phonolites from the Osamu Utsumi open pit uranium mine (Shea 1992), show that the duration for the entire

magmatism may be well in excess of 5 Ma (Vlach *et al.* 2002).

The Poços de Caldas massif was formed by multiple magmatic intrusions with phonolite-nepheline-syenite compositions, and by basanite-tephrite-ankaramite volcanic rocks successively intruded by phonolitic and lamprophyric dikes (Lustrino *et al.* 2003). The rocks are mainly phonolites (~80% of outcrops) and nepheline syenites (~14%), generally very potassic. Sodic breccias, agglomerates, tuffites made of tephrite with subordinate ankaramite, nephelinite, and phonotephrite are all sodic and occur at the western margin of the complex. Dikes of tephrite, trachybasalt, and basaltic trachyandesite occur to the NW. Mafic-ultramafic dikes of lamprophyric to carbonatitic affinity occur within the Osamu Utsumi pit. Ultramafic lamprophyric with alnöitic affinity and siliceous carbonatite dikes intrude the Neoproterozoic quartz mangerites near the northern borders of the complex.

Phonolites are fine-grained and composed of K-feldspar, nepheline, clinopyroxene with diopside cores to hedenbergite and aegirine-augite rims, titanite, and rare Ti-magnetite. Phonolites are characterized by astrophyllite $(\text{K,Na})_3(\text{Fe,Mn})_7\text{Ti}_2(\text{SiO}_3)_8(\text{O,OH})_7$, eudyalite - hainite $\text{Na}_4\text{Ca}_8(\text{Ti,Zr,Mn,Fe})_3(\text{SiO}_2)_8\text{F}_8$, arfvedsonite, aenigmatite $(\text{Na}_2\text{Fe}^{2+}_5\text{TiSi}_6\text{O}_{20})$, sodalite, and fluorite. Nepheline syenites have approximately the same mineralogy, but contain less than 20 vol% of mafic minerals, with occasional sodalite, biotite, and arfvedsonite. The occurrence of eudyalite-rich variants (lujavrite) is rare. Zircon is ubiquitous.

Most of the rocks are very poor in silica and Mg and rich in Fe, as typical peralkaline rocks. But these rocks are also characterized by very low Na content and strong enrichment in K, which results at least partially from late stage K-metasomatism (Table 5-7). High-K compositions with strong U, Y, and REE enrichments also characterize the mafic-ultramafic lamprophyres. Several late carbonatitic veins cross-cut all the above rocks. Some of them are made up of almost pure calcite, while others contain calcite + fluorite, calcite + phlogopite + fluorite, calcite + phlogopite + magnetite + apatite, or calcite + phlogopite + fluorite ± albite. Occasionally they contain thorite, REE-fluorocarbonates, and accessory minerals.

Six stages have been recognized in the complex formation: (i) volcanic activity; (ii)

caldera formation; (iii) alkaline magmatic activity; (iv) annular dike formation; (v) foidite, chibinite, and lujavrite intrusions, (vi) ultramafic dike emplacement. Two large fault systems with predominant directions of N60W and N40E occur in the alkaline complex. The former is correlated with the regional tectonics and the latter with the caldera formation process (Fraenkel *et al.* 1985).

In the Poços de Caldas peralkaline complex, the 48 radioactive showings can be grouped into three associations: U-Zr, Th-REE, and U-Mo (Oliveira 1974, Fraenkel *et al.* 1985, Waber *et al.* 1992). The U-Zr association is the most common one in alluvial and eluvial type deposits. The Th-REE association is the second-most common one, especially in the Morro do Ferro deposit. The U-Mo association occurs in lodes and lenticular bodies and forms the mined deposits. Almeida and Paradella (1976) propose the existence of 7 secondary circular structures, based on the analysis of satellite images, and show that the radioactive anomalies are all located along the border of these structures.

The Osamu Utsumi uranium mine (also named Cercado), which comprises the ore bodies A, B, and E, covers an area of about 2.5 km² at the southern margin of a major secondary circular structure called Planalto (Fig. 5-13). The Agosthino deposit is located at the northeastern margin of the Planalto structure (Fig. 5-14). Ore bodies A and B are hosted by tinguaites and phonolites of the internal part of the Planalto structure, whereas the ore body E is hosted by pyroclastics, including various types of phonolites, tuffs, and breccias cross-cut by dikes of ultrabasic rocks and overlying a foyaite along the circular structure. Several breccia pipes cut these formations.

Ore body A has measures 900 m by 500 m. It comprises of subvertical breccia veins in which uranium has been leached out from the upper part, with redeposition of U at a subhorizontal redox front at a depth of -100 m (Lichtner & Waber 1992). Ore body E is slightly larger, at 1100 m by 500 m, and its mineralization occurs at a depth of -150 m. The mineralization here is completely epigenetic and associated exclusively with a subhorizontal redox front. Ore body B (1400 m by 400 m) mineralization corresponds to hydrothermal deposition of uranium in lenticular bodies developed in the porous pyroclastics and

TABLE 5-7: SELECTED COMPOSITIONS OF UNMINERALIZED AND MINERALIZED ROCKS OF THE POÇOS DE CALDAS PERALKALINE COMPLEX, BRAZIL.

wt%	84-56	84-57	84-45	84-46	84-58	84-44	84-47
SiO ₂	56.81	55.06	58.65	59.93	54.98	55.09	56.6
Al ₂ O ₃	22.89	23.78	24.14	22.49	21.43	20.92	21.29
Fe ₂ O _{3t}	3.11	3.29	1.46	1.23	3.88	3.73	4.2
MnO	0.24	0.25	0.02	0.01	0.58	0.52	0.02
MgO	bdl	bdl	0.03	0.2	bdl	0.27	0.15
CaO	bdl	bdl	bdl	bdl	0.91	1.22	bdl
Na ₂ O	0.09	0.22	0.19	0.06	0.3	1.79	0.13
K ₂ O	12.95	12.63	13.26	13.77	12.75	10	13.42
TiO ₂	0.53	0.65	0.51	0.37	0.63	0.45	0.4
P ₂ O ₅	0.12	bdl	0.15	0.17	bdl	0.11	0.16
IL	3.63	3.74	3.25	2.01	3.33	5.61	3.62
Total	100.4	99.62	99.86	100.24	98.79	99.71	99.99
CO ₂ %	0.27	0.27	0.05	0.05	1	1.61	0.07
P	273	262	276	291	245	133	281
Q	37	30	37	38	13	21	25
B	46	49	25	25	56	59	61
A	171	191	185	146	107	96	128
F ppm	1100	1100	900	600	700	900	700
Ba	634	1845	409	459	1088	1142	473
Co	<10	<10	<10	<10	<10	<10	<10
Cr	56	<10	<10	<10	<10	11	<10
Cu	<10	<10	<10	<10	<10	<10	<10
Ni	<10	<10	<10	<10	<10	<10	<10
Sr	468	235	132	255	331	335	561
V	312	177	398	179	128	186	238
Rb	346	532	463	432	501	520	500
U	15.6	14.5	62	78.5	29	345	963
Th	51	41	30	91	38	35	47
Th/U	3.27	2.84	0.48	1.15	1.28	0.10	0.05

84-56: massive foyaite, CDSC262-596m; 84-57: foyaite, CDSC262-622 m; 84-45: sulfide foyaite, open pit; 84-46: ocellar foyaite, open pit; 84-58: tingaite, CDSC262-602 m; 84-44: tingaite breccia, open pit; 84-47: mineralized tingaite contact ore E – B; bdl: below detection limit (M. Cuneý, unpub. data).

intersected by mineralized breccia pipes rich in fluorite. Secondary uranium mineralization is also developed at the redox interface. In the Agosthino deposit, the U mineralization is restricted to a subvertical breccia body, which seems to result from the combination of magmatic and late magmatic-hydrothermal fractionation, as in all other world occurrences, but part of the economic mineralization is attributed to an overprinting by supergene alteration (lateritization). The U-Zr-REE mineralization is concentrated in the matrix of the breccias. Unlike at Ilimaussaq and Bokan Mountain, Th is not significantly enriched during the mineralization process at Poços de Caldas, despite the occurrence of fluorine in the system (Table 5-7).

Hydrothermal alteration following formation of the breccias has resulted in potassic alteration and pyritization of the hosting phonolites and syenites. Small fluorite and pyrite

veins are also common. Feldspars were completely altered into pure potassic feldspars, nepheline into illite and kaolinite, and clinopyroxenes into a mixture of TiO₂-rich minerals, clay minerals, and pyrite. In this alteration zone K, S, U, Pb, Rb, Ba, and Mo are strongly enriched, whereas Ca, Na, Mg, and Sr are depleted. Fluid inclusion data indicate relatively low temperatures for the alteration process, around 250°C, and aqueous fluid compositions with approximately 7 wt% KCl. In the breccia pipes where Zr, Hf, F, and minor REEs are enriched, fluid inclusion data indicate boiling of a brine with a temperature of 210°C and 40–45 wt% KCl-NaCl, together with FeSO₄ and KF. Higher temperature fluids of magmatic origin should have been involved in these processes to explain the formation of the breccia pipes, the Zr, Hf, F, and REE enrichment, the occurrence of uraninite, and the $\delta^{34}\text{S}$ values of the

hydrothermal pyrites near 0‰, but such fluids have not been observed.

The main U minerals are uraninite with 2 wt% Th, coffinite, and brannerite. The major REE mineral is monazite and the main Zr minerals are zircon and baddeleyite associated with jordisite (MoS_2), pyrite, galena, sphalerite, and fluorite in the high-grade mineralization of the breccias. Low-grade pitchblende mineralization occurs in the altered host rocks. The ultramafic dikes of the open pit are dated at 76 Ma, putting a lower age limit to this early mineralization event.

Subsequently, lateritic weathering led to the oxidation of the sulfides to iron hydroxides-oxides and alunite-jarosite to depths of 80 to 140 m below surface, and to greater depths along fractures. The resulting redox front is typically marked by a change from reddish-brown oxidized to bluish-grey reduced rocks. A secondary mineralization of pitchblende, partly occurring as nodules associated with secondary pyrite, has been developed at this redox front. This secondary pyrite has low $\delta^{34}\text{S}$ (–13‰) attributed to microbial activity.

In the immediate vicinity of the redox front, K-feldspar is altered to kaolinite. Gibbsite is present at the saprolite-laterite contact. Part of the uranium is adsorbed on poorly crystalline phases. Only a slight loss of REEs, mainly the LREEs, is observed between oxidized and reduced rock. REEs are mainly trapped as clay-sized crandallite group minerals. Selective fractionation of Ce, typical of lateritic weathering, is observed. Zirconium occurs in caldasite, a U- and Th-bearing mixture of fibrous baddeleyite (ZrO_2) and in metamict zircon (Tolbert 1966). Caldasite forms a significant part of the ore.

Weathering of Poços de Caldas has required up to 2 m.y. in the absence of pyrite, and up to 0.5 m.y. when pyrite is present at the Osamu Utsumi uranium mine. A large proportion of the uranium mined at Poços comes from these weathering redox fronts. The resources estimated in 1985 were 22,780 t U 25,000 t Mo_2O_3 and 172,400 t ZrO_2 (Fraenkel *et al.* 1985). The cumulative uranium production from Poços de Caldas between 1985 and 1995 was 1,030 t U.

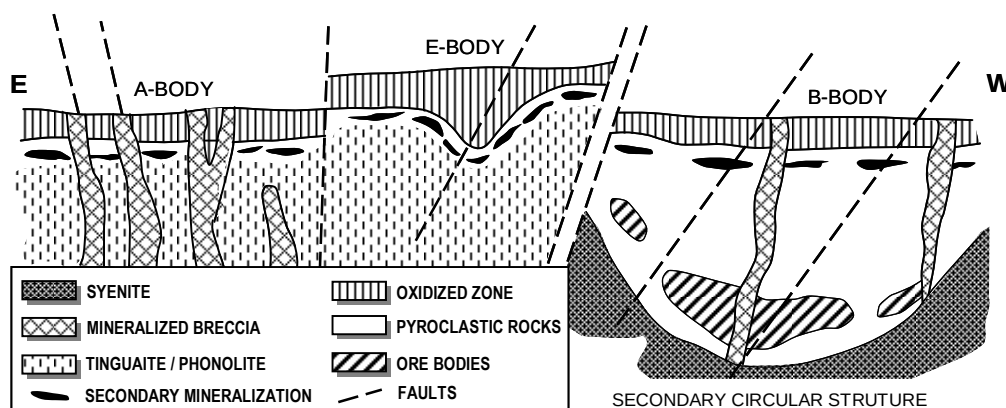


FIG. 5-14: Schematic cross section through the Cercado deposit of the Poços de Caldas peralkaline complex (Fraenkel *et al.* 1985).

Thor Lake (Canada). The Thor Lake deposits in the Great Slave Lake district, are related to a subcircular (16 by 33 km), high level peralkaline complex, the Blatchford Lake Intrusive Suite. They consist mainly of Be, Y, REE, Nb, Ta, Zr, and Ga deposits (Davidson 1982), but Th and U mineralization is also known in the area. The Thor Lake area has peralkaline intrusions in association with less fractionated plutonic units. The Blachford Lake complex consists of an earlier suite of gabbro, quartz syenite and granite, and later intrusion of the Grace Lake granite,

intruded itself by the Thor Lake syenite in its core. The two members of the late suite are peralkaline and classically show a noticeable enrichment in Fe, Mn, Na, Nb, REE, F, and partially in Be and Li. The granite and syenite are characterized by moderately high K_2O (4.6 % in average), average U (3 ppm) and Th (18 ppm) contents and a Th/U ratio of 6. Principal radioactive minerals in these rocks are zircon and bastnaesite with minor REE-rich apatite and monazite. Felsic dikes within the Grace Lake granite and Thor Lake syenite are one order of

magnitude richer in U and Th than host rocks, and are also enriched in elements occurring in the Thor Lake deposits.

The suite intrudes Archean metasediments and calc-alkaline granitoid intrusions of the Slave Province to the north and is proximal to Paleoproterozoic metasediments, metavolcanics, and intrusions of the Athapascow aulacogen of Great Slave Lake on the south. The Hearne Channel granite and the Thor Lake syenite have been dated by U-Pb on zircon at 2175 ± 10 Ma and 2094 ± 10 Ma, respectively (Bowring *et al.* 1984). The Grace Lake granite is reddish-pink to buff to pale gray, coarse-grained and equigranular. It is a hypersolvus granite with mesoperthitic alkali feldspar, quartz and riebeckite. Accessory minerals include fluorite, magnetite, zircon, carbonates and phenacite. The Thor Lake syenite is medium- to coarse-grained, hypersolvus with perthitic alkali feldspar, riebeckite, and minor quartz. Accessory minerals include magnetite, zircon and ferromagnesian mica. A pyroxene-olivine-bearing syenite forms a thin rim along the margins of the syenite body and subhorizontal pods and sheets of syenite pegmatite occur within it. Contact between granite and syenite is transitional in the western part of the complex. In the north and east, the granite dips under the syenite and the contact is sharp.

Most mineralized zones occur within the western part of the Thor Lake syenite. The deposits are related to albitite or greisen. For example, the core is occupied by breccias composed of syenite pegmatite clasts cemented by a fine- to coarse-grained matrix of albite and ferromagnesian mica. Accessory minerals include zircon, uranothorite, allanite, magnetite, hematite, columbite, pyrochlore, carbonates, bastnaesite, and lanthanite. Quest Uranium Corporation have reported average assays of 535 ppm U from 196 channeled samples in the Thor Lake area. Uranium occurs in an east-west trending pegmatite dike complex that can be traced over a strike length of 2.5 km and widths of up to 400 m.

Palabora carbonatite (South Africa). The peralkaline Palabora intrusive complex was emplaced at about 2030 ± 18 Ma in Archean formations in north-eastern Transvaal (Eriksson 1984). The successive intrusions comprise a kidney-shaped pyroxenite followed by a syenite and ultrabasic pegmatoids. The apatite-

phlogopite-rich pyroxenite extends over 6 by 2.5 km and forms a pipe at depth. The ultrabasic pegmatoids intrude the pyroxenite in three localities. In the central one, a phoscorite (a magnetite-olivine bearing rock) and carbonatites form the Loolekop pipe (1.4 by 0.8 km). This pipe is intruded by a dike-like banded carbonatite. In the late carbonatite, at Loolekop, repeated fracturing was associated with fluid percolation responsible for the deposition of chalcopyrite and minor cubanite in vertical veins. Sets of parallel veins may occur over a width of 10 m, but individual veinlets are less than 1 cm thick and less than 1 m long. The ore grade is about 1 % Cu. In the phoscorite, magnetite with up to 4 wt% TiO_2 may reach 25 to 50 vol%. Apatite reaches economic grade in the phoscorite. Baddelyite occurs in both the phoscorite and the carbonatite.

Sulfur isotopes and fluid inclusions show that the fluids depositing copper were derived from carbonatitic magma (Eriksson 1989). The copper ore body is hosted in a carbonatite pipe with grades highest at the core (1.0% Cu). Drilling has shown that the ore body extends beyond a depth of 1000 m. Ore reserves are in the order of 300 Mt Cu at 0.69% Cu.

The main U- and Th-bearing mineral is uranothorite, $(\text{U,Th})\text{O}_2$, a quite exceptional mineral in uranium deposits. Its crystallization necessitates the absence of quartz in the host rock because the presence of silica leads to the formation of uranothorite. It is recovered as a heavy mineral together with baddelyite and processed for uranium production. Thucholite, a variety of Th-U rich pyrobitumen has been observed in fractures in the center of the transgressive carbonatite. The average uranium content of the ore body is 25 ppm U (Verwoerd 1986).

Palabora Mining Company started U production in 1994 as a by-product of the Palabora copper mine. The mine yielded a cumulative production of 640 t U and the uranium treatment circuit was stopped in 2002. Thorium may become a significant product in the future, with the development of a thorium reactor. The uranothorite contains more than twice the Th resources, compared to U (55.82 wt% ThO_2 26.65 wt% UO_2 , 13.40 wt% PbO , 4.12 wt% REE_2O_3). With an average grade of 0.01 % Th, the carbonatite should contain about 10,500 t Th for every 100 m of depth (Verwoerd 1986).

Another carbonatite with potential U and Th resources is the Eldor in the Labrador Trough, Québec. Commerce Resources Corp reports an average U content of 200 ppm on 60 rock samples (85 to 1400 ppm U). However, the most common U- and Th-bearing phase in this carbonatite is pyrochlore, which cannot be easily leached for production of U.

PARTIAL MELTING

Introduction

Deposits related to partial melting are those corresponding to uraninite enrichment disseminated in pegmatitic dikes injected into high-grade migmatitic gneisses. These are generally called pegmatoids because they tend to lack the zonation that is commonly observed in pegmatites derived from fractionation of large plutonic bodies. They are also called alaskites in reference to the terminology used at the Rössing deposit in Namibia.

The major deposits related to partial melting processes in the world are the following in order of the size of their global resources:

- Rössing (Namibia) 246,500 t U at 300 ppm
- - Husab (Namibia), 188,000 t U at 410 ppm
- - Anomaly A (Namibia), 48,740 t U at 175 ppm
- Ongolo (Namibia), 46,430 t U at 115 ppm
- Valencia (Namibia), 32,400 t U at 108 ppm
- Charlebois Lake (Canada), 17,500 t U at 600 ppm
- N. Shore Double S (Canada), 8,925 t U at 110 ppm
- Oshively (Namibia), 7,970 t U at 166 ppm
- Kalinovskoye (Ukraine), 7,550 t U at 500 ppm

The Rössing uranium deposit in Namibia is the most significant of these deposits (Berning *et al.* 1976, Cune 1980 1982). They represent the most extreme temperature and pressure conditions known for economic U mineralization, although Rössing is one of the lowest-grade U deposit ever mined.

The Rössing Deposit (Namibia)

General Geology. Uranium mineralization associated with leucogranite dikes has been known since 1928. These granites have been termed pegmatite by Smith (1965), potash granite by Nash (1971) and alaskite by the geologists at

Rio Tinto Exploration. Rio Tinto began the first serious evaluation of alaskite-related deposits in 1966 and mining at Rössing started in 1976. The open pit presently extends over 3.5 km in area, and down to 350 m. Total production by the end of 2007 was slightly over 90,000 t U. In December 2005, the operational life of the mine was extended to 2016, with potential to further extend to 2021. The deposit is situated approximately 55 km northeast of Swakopmund (Fig. 5-15), in the Namibian desert. Several other deposits of the same type have been discovered in a similar geological setting, such as the SH and SK deposits, immediately east and west of the Rössing deposit, respectively, and the Goanikontes, Ida, and Valencia deposits, some tens of kilometres to the SW and NE.

Granitoid dikes hosting the uranium mineralization at Rössing appear to be derived from the partial melting of uranium-rich sediments. Uranium mineralization is further enriched during emplacement of the alaskites by percolation of magmatic fluids through reducing metamorphic host rocks and by recent supergene weathering, mainly in the upper part of the deposit.

Uraniferous alaskites are located in strongly metamorphosed and partially migmatized rocks of the central and deeper portion of the Damara Panafrican belt between the Omaruru and Okavango Lineaments (Fig. 5-15). The Damara Orogen belt belongs to the Neoproterozoic-Cambrian Pan-African mobile belt system. This belt is composed of two branches, one parallel to the Atlantic coast (the Kaoko belt), separated from the Ribeira belt along the eastern coast of Brazil by the opening of the Atlantic Ocean, and a second one, transecting the African continent, oriented to the northeast. The orogen results from a triple continental collision between the Kalahari craton to the north, the Congo craton to the south, and Sao Francisco craton to the East. The “Matchless Amphibolite” to the south (Fig. 5-15) comprises tholeiitic metabasalts with N-MORB to E-MORB compositions (Schmidt & Wedepohl 1983) and represents the sliced upper part of an oceanic crust. This formation therefore corresponds to the suture formed by the northerly dipping subduction of the Damara Ocean between 750 and 520 Ma ago. Early crustal thinning took place around 700 Ma ago and may have been accompanied by limited seafloor spreading.

The Central Zone of the northeast branch comprises metamorphic polyphased deformed rocks of the Damara sequence, resting unconformably on the Mesoproterozoic gneissic and volcanosedimentary Abbabis Complex, which has a minimum age of 1038 Ma, obtained from the Kibaran granitoids that intrude them (Kröner *et al.* 1991). The metamorphic grade increases westward, reaching high-grade conditions with local partial melting in the coastal

area. The sedimentary package (Table 5-8 and Fig. 5-16) was deposited in an epicontinental platform in a passive margin environment. In the western part of the 400 km-wide north-eastern branch of the Damara belt, sedimentation began at 900-800 Ma on the Mesoproterozoic Abbabis granite-gneiss, in three independent intracontinental rifts oriented nearly east-west. The 40 km wide grabens are mainly filled with high-energy siliciclastic sediments corresponding

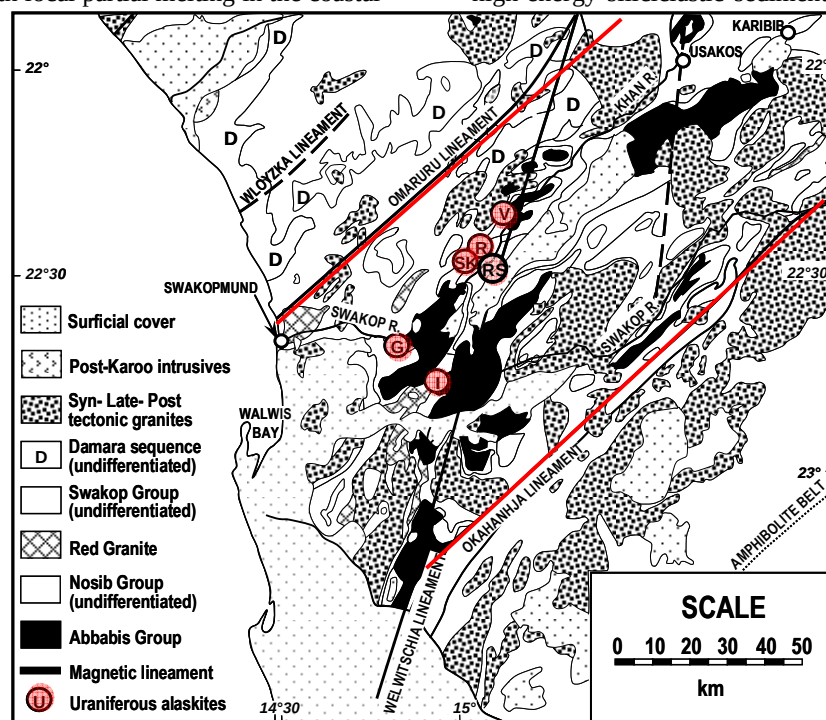


FIG. 5-15: Geological map of the southern part of the Damara Belt with the location of the main alaskitic uranium deposits: (V) Valencia, (R) Rössing, (SK) SK anomaly, (RS) Rössing South or Husab, (G) Goanikontes, (I) Ida (after Jacob *et al.* 1986). Between the red lines: central part high grade metamorphic zone of the Damara Belt.

to the Nosib Group. Rhyolitic volcanics have been recognized in the less metamorphosed northern and southern grabens, and evaporitic sediments occur in the southern one (Porada & Behr, 1988). Evidence of former evaporites in the Khan Formation is given by the presence of marialitic scapolites (Cuney, 1981) and by the common occurrence of anhydrite, accounting for up to 20 vol% of some of the gneisses associated with carbonates (Nash, 1972).

The original lithologies of the central graben have been obscured by intense metamorphism. The total thickness of the Nosib Group may locally reach up to 6000 m. Low

energy epicontinental sedimentation was subsequently deposited to form the Swakop Group, which extends over the width of the Damara basin.

The overlying Khan Formation is 100 to 250 m thick and comprises the Lower Banded Gneisses, the Upper Banded Gneisses, and the amphibole schist units. These are mainly composed of hornblende $\pm$ clinopyroxene, quartz-plagioclase-K-feldspar gneiss enclosing disrupted amphibolite layers. Garnet or calcite may appear in some of these gneisses. Zircon, monazite, and titanomagnetite are more common than allanite, titanite, and anatase. The foliation is frequently

enhanced by coarser-grained quartz–plagioclase ± K-feldspar ± biotite ± magnetite segregations.

Numerous felsic dikes are generally emplaced parallel to the metamorphic foliation, but

TABLE 5-8: LITHOSTRATIGRAPHY OF THE CENTRAL PART OF THE DAMARA BELT.

Formation		Lithology	Accessories
Swakop Group	Khomas	Tinkas (Kuseb)	Pelitic schist (biotite, muscovite, cordierite, sillimanite)
		Karibib (Husab)	Dolomitic and calcitic marble
		Chuoss (<250m)	Glaciomarine diamictite
	Rössing (200m to absent)	Metaquartzite, calcsilicates Upper metapelitic gneiss Upper marble Lower metapelitic gneiss Lower marble	graphite, sulfides, zircon
~~~~~ Minor unconformity ~~~~~			
Nosib Group	Khan (500m)	Amphibole biotite schist Upper gneiss (hornblende, biotite, quartz, plagioclase ± K-feldspar ± pyroxene ± calcite) Pyroxene-garnet-amphibole gneiss Lower gneiss	titanomagnetite, zircon, allanite, titanite, monazite
	Etusis (up to 6000m)	Congl, quartzite, migmatized feldspath. gneisses Rhyolitic volcanics (?)	titanomagnetite, monazite, zircon, apatite, tourmaline
~~~~~ Major unconformity ~~~~~			
Abbabis		Migmatized granite gneisses/basement	monazite, zircon, apatite

Modified from Nash (1971); Smith (1965); Martin & Porada (1977).

discordant dikes also exist. Scapolite with 51% meionite, locally occurs in fractures perpendicular to the foliation layers and cross bedding structures are very well preserved, reflecting continental or near-shore deposition. Biotite abundance varies within the Etusis lithologies. Muscovite may occur in small amounts and is always secondary.

The initial structure of the feldspar-rich quartzites was affected by partial melting, in which quartz and feldspar became coarser, and tend to form layered-to-diffuse segregations or pegmatite veins crosscutting the original stratigraphy defined by the heavy mineral layers. Frequently tourmaline is concentrated in recrystallized graphic segregations with quartz, or in the pegmatite veins. The majority of the Rössing dome, formerly described as the Abbabis granite-gneiss, consists of varying degrees of partially melted Etusis Formation rocks (Nex *et al.* 2001). The Etusis Formation is therefore much thicker than the 800 m previously estimated by Berning *et al.* (1976).

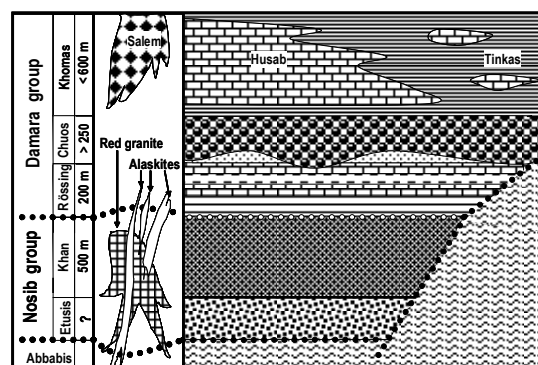


FIG. 5-16: Lithostratigraphic succession of the Damara sediments and level of emplacement of different types of granitoids in the Rössing area.

Marine sediments subsequently filled the entire Damara Basin, forming the Damara Group, which is composed of four formations: the Rössing, Chuoss, Karibib (Husab) and Kuseb

(Tinkas). The Rössing Formation comprises six units including a basal, impure forsterite, phlogopite and graphite marble (Lower Marble) overlain by the lower biotite-cordierite gneiss, a conglomerate, the upper marble, the upper biotite-cordierite gneiss, and a feldspathic quartzite layer. The local occurrence of veinlets of marialite-rich scapolite (up to 49 % marialite) in the upper Khan gneisses and in a quartzite of the Rössing Formation may reflect the former presence of evaporites in the sedimentary succession.

The tillite of the Chuos Formation corresponds to an important glaciation event and covers the entire Damara Basin. The tillite has a grey, massive-to-schistose matrix containing unsorted, elongated, and angular erratic blocks, measuring up to one meter in diameter. The superimposed Karibib Formation comprises a succession of white to bluish-grey, well-bedded marble units containing thin layers of calc-silicate. The Kuiseb Formation consists of biotite-cordierite-sillimanite schists intruded by numerous pegmatite dikes.

The collision between the Congo and Kalahari cratons resulted in thrusting, associated folding, and foliation development at high structural levels, still preserved in the Usakos-Karibib area to the NE of the Rössing area. At deeper structural levels, in the southern-most metamorphosed Central Zone, regional migmatization is associated with voluminous synkinematic granitoid emplacement. Crustal shortening has produced pure, shear-dominated deformation and an orogen-parallel ductile flow of the rocks with constrictional-type strains and a SW sense of displacement (Kisters *et al.* 2004).

Four distinct structural domains are associated with the Damara orogen: a weakly metamorphosed northern platform, a 40 km wide transition zone with metamorphic grade increasing southwards, a 160 km wide highly metamorphosed central zone where the Rössing deposit is located, and a 100 km wide southern zone with a southward-decreasing metamorphic grade. In the central zone, the metamorphism is estimated at medium-pressure, high-temperature granulite facies conditions, with temperatures of ca. 750°C, pressures of 5–6 kbar (Masberg *et al.* 1992, Jung & Mezger 2003) and partial melting. This area is characterized by the development of kilometer-scale, NE-trending, elongated domes. According to Poli and Oliver (2001), these domes are overturned and verging southwest. Many of

them present a core of basement gneisses or Etusis Formation. The overlying Swakop sequence is draped around the domes and folded into tight and geometrically complex synclines, striking northeast-southwest. The main structural orientation in this area is northeast and results from an F3 deformation event. The early phases of folding produced the overturned and recumbent structures that were accompanied by intense shearing and limited thrusting. Several less intense phases of folding occurred after F3 and produced NE- and NW-trending folds.

Several generations of granitoids have been recognized in the Central Zone including the Red and Grey granites, the syn- to post-tectonic Salem granites, post-kinematic Donkerhuk and Bloedkoppie leucogranites, and the late- to post-tectonic layered leucogranites, also called the G4 granites or alaskites. The Salem granitoids have batholithic dimensions and are the most abundant and represent different generations of granodiorites, granites, and adamellites, dated at 601 ± 79 Ma.

Deformation and plutonism related to D2 is considered to have occurred before ca. 540 Ma, because at that time, D3 produced not only the dome structures, but also the intrusion of voluminous granitic bodies. All monazite fractions from the metasedimentary rocks and migmatites from the Central Damara orogen are nearly concordant and define three different events: (i) at 540 Ma, attributed to the prograde metamorphic event M1 associated with thrusting and D3 dome formation, corresponding to the main collisional phase between the Kalahari and Congo cratons (Kisters *et al.* 2004); (ii) at ca. 520–510 Ma, related to the peak regional metamorphism M2, during which partial melting of the metasediments occurred; and (iii) at ca. 490–470 Ma (Jung and Mezger 2003), corresponding to a late metamorphic event M3, during which voluminous intrusions of post-collisional A-type and S-type granites occurred.

The equigranular, homogeneous, fine- to medium-grained Red or Grey Granites form irregular bodies within the Etusis Formation. They were dated by U-Pb in zircon from the Goanikontes area at 525 ± 5 Ma, and were thus formed at the beginning of the M2 phase. The 508 ± 2 Ma age obtained on zircons and monazites and the 509 ± 1 Ma age obtained on uraninite from an alaskite from Goanikontes (Briqueu *et al.* 1980) confirm emplacement of the alaskite during

the final stage of the M2 event, associated with partial melting.

All uranium-bearing occurrences related to alaskites discovered in Namibia (Rössing, Goanikontes, Ida, Valencia, etc.; Fig. 5-15) are situated in the southern Central Zone. The alaskite

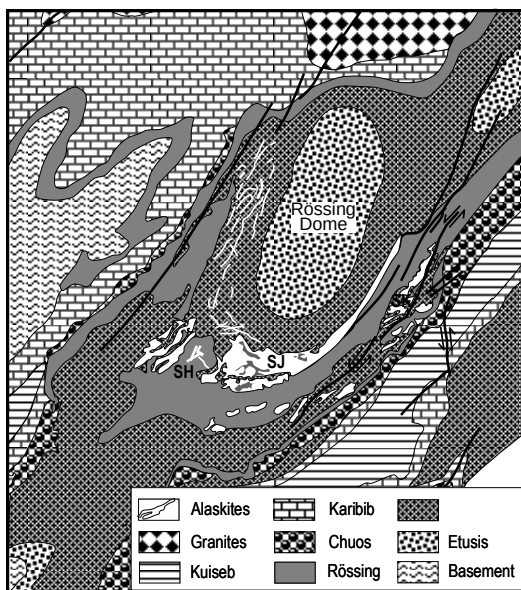


FIG. 5-17: Geological map of the Rössing Dome area with the main uranium deposits: SJ, SH, and SK (modified after Basson & Greenway 2004).

dikes are widely distributed in all the formations of the Damara metasedimentary package, but are not always uraniferous. They were emplaced post-F3 and are mainly oriented NNE. The alaskites of the Rössing uranium deposit occur in a migmatitic zone to the southwest of a dome structure comprising an intensely migmatitic core (Fig. 5-17). In the open pit, granitic rocks vary widely in texture, color, grain size, thickness, and emplacement habit. They range from centimeter-thick quartzo-feldspathic anatectic lenses to several tens of meters-thick dikes, to a few hundreds of meters-thick stocks. Uraniferous alaskites tend to be parallel to the major foliation of the hosting metamorphic rocks, but they can be locally discordant when emplaced along the axial planes of F3 folds, transgressing the concentric envelopes consisting of various lithologies. Their contacts with enclosing rock is commonly sharp as observed in the Rössing open pit, but progressive transitions between quartz-rich meta-arkoses and pegmatoids can be observed when

extraction of the melts has been limited. In the western part of the open pit, the more massive alaskite bodies contain large host rock xenoliths more than 100 m in size. Their texture and grain size are highly variable: they are commonly pegmatitic, but also granitic, and more rarely aplitic, orbicular and graphic. Such textures may alternate in the same dike and define layering. Biotitic schlierens commonly occur at the margin of the dikes. The rarity of garnet and the absence of cordierite in these granites suggest that the water needed for partial melting, typically provided by biotite incongruent dehydration-melting reactions, was absent.

Nex and Kinnaird (1995) have shown that the layered leucogranites of Goanikontes can be divided into six types, which have been also recognized at Rössing by Basson and Greenway (2004). Types A, B, and C are deformed by D3 and D4, while types D, E and F are post-D3. Type A and B granites are pale pink to white, irregular to boudinaged, notably folded by D3, but with a weak S3 foliation. Type C granites are white to pale pink and abundant in relatively undeformed parts of the cover sequence, occasionally boudinaged, and emplaced in F3 'flexures'. Type D granites are white, 0.2 to 3 meters wide, and can be host to U mineralization. They contain smoky quartz, occur as irregular to anastomosing shapes, and are typically confined to high-strain zones between the Khan and Rössing Formations. Type E granites dominate in high-strain, post-D3 zones and form 30 m wide massive bodies parallel to the metamorphic foliation, with highly variable grain size and color. Type F granites are distinctively red, post-kinematic, having parallel sides and cross-cutting all pre-existing structures (Nex & Kinnaird 1995; Nex *et al.* 2001).

The alaskites are preferentially emplaced into the pyroxene-hornblende gneiss and biotite-amphibole schist units of the Khan Formation in the northern part of the pit, and into biotite-amphibole schist, lower marble, and lower biotite-cordierite gneiss of the Rössing Formation in the central part of the pit. Skarn bodies from a few centimeters to several meters wide occur at the contact between the alaskites and the marbles of the Rössing Formation. The majority of skarns are composed of green clinopyroxene, brown calcic garnet, and varying amounts of scapolite. Pyroxene, garnet, phlogopite and titanite up to several tens of centimeters in size may occur locally.

General Characteristics of the U Mineralization.

Primary U mineralization is observed only in the felsic rocks. Some alaskites can be entirely barren or only slightly mineralized, whereas only a few restricted sections are economic. The mineralized felsic rocks are mainly located in the upper part of the Khan Formation and in the lower part of the Rössing Formation, with uranium enrichments in the vicinity of the marbles of the Rössing Formation and the amphibolite from the Khan Formation. Mineralized alaskites contain smoky quartz, tend to be richer in biotite, are generally pink to reddish, and commonly correspond to the latest intrusions. Barren alaskites are white to pale pink, although white mineralized alaskites do exist. Within mineralized alaskites, the distribution of U can be extremely variable and enrichments are commonly observed at the margin of the dikes and along biotite selvages. Pinkish K-feldspar is predominant, but plagioclase-rich alaskites also exist (Fig. 5-18). Biotite generally represents less than a few percent of the volume of the rock, and has up to 1.8% F and 800 ppm Cl (Cuney, unpub. data). Biotite clusters are most frequent near the contact with the host rocks and muscovite occurs in very limited amounts as alteration of plagioclase and K-feldspar, and associated with chloritized biotite frequently accompanied by calcite. These rocks have been called “alaskites” at Rössing because of their predominantly potassic character and their low biotite content in accordance with the definition of Spurr (1900).

Euhedral Th-bearing uraninite (3.3 to 8.0 wt% ThO₂) is the predominant ore mineral, with Elevated Y and REE contents (1.6 to 7.14 wt% Y₂O₃ and 2.3 wt% ΣREE), as commonly observed in uraninites formed at high temperature. The Ca contents are however unusually high and variable (0.3 to 9.5 wt% CaO) for uraninite formed from a leucocratic granite, which generally contains low Ca. This high Ca content is interpreted as reflecting the interaction of the alaskite melts with the enclosing calcium-rich lithologies of the Khan and Rössing Formations, as observed in pitchblende from vein-type deposits hosted by calcium-rich lithologies, formed at lower temperature. Uraninite occurs as finely disseminated grains ranging in size from a few microns to 0.3 mm, with the majority in the 0.05 to 0.1 mm fraction. Uraninite grains are included in quartz, feldspar, and biotite, but also appear

interstitially to these minerals. Uraninite can be also included in zircon and apatite crystals, indicating that its crystallization started very early. Uraninite enrichments are also common in the biotite selvages, in which the uraninite crystals tend to be distributed as trails. This type of uraninite has crystallized from a late magmatic fluid.

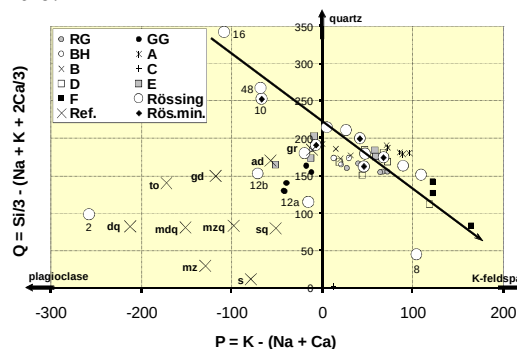


FIG. 5-18: Q-P chemical-mineralogical diagram of Debon and Lefort (1983) representing variations of the proportions of quartz relative to plagioclase and K-feldspar (RG = Red Granite, GG = Grey Granite) and alaskites (BH = basement hosted alaskites and type A to F) from Goanikontes (from Nex *et al.* 2001) and alaskites (min. = mineralized alaskites = black diamond) from Rössing (data from Cuney 1981). Rössing samples are numbered. Ref. = average composition of main plutonic rock types.

Other high- and low-temperature U-bearing minerals can also occur. Nb-Ta-Ti oxides are the most common high-temperature U-bearing minerals, commonly with about 27 wt% UO₂, occurring as 1 to 3 mm octahedral to dodecahedral crystals of betafite, isolated or in clusters. Hosting only a few percent of the total uranium content of the alaskites in the Rössing deposit, betafite becomes the predominant U-bearing mineral in the SH mineralization (Fig. 5-16). No specific mineralogical or textural features seem to control the occurrence of betafite in the alaskites.

Monazite is a common mineral in most occurrences at Rössing, but is generally not abundant because of the low solubility of monazite in low-temperature, peraluminous melts (Montel 1993). However, monazite may be enriched in some alaskitic granites such as those from Southern Finland (Lauri *et al.* 2007). Such monazite accumulations can tentatively be explained as restitic to cumulative enrichments

after partial melting of paleoplacers in meta-arkosic conglomerates. Zircon is systematically present in the alaskites, but its abundance is generally very low, also because of its low solubility in low-temperature, peraluminous melts (Watson & Harrison 1983). However, it can be very abundant, as in the late orogenic alaskitic granites from Southern Finland (Lauri *et al.* 2007). As with other U-Th-rich silicates, these zircons are strongly metamict. Allanite and titanite are rare at Rössing and occur in the vicinity of Ca-rich rocks (i.e. marbles and calcsilicates). Dark purple fluorite is frequently present in the vicinity of chloritized biotite and as coatings on fracture surfaces. Apatite, xenotime, pyrite, chalcopyrite, bornite, molybdenite, arsenopyrite, anatase, magnetite, hematite, and ilmenite are encountered locally.

Hexavalent uranium minerals are common in the Rössing ore body and contribute to its uranium enrichment, particularly in the upper part of the deposit. Oxidizing meteoric fluids have dissolved uraninite in the eroded part of the deposit and redeposited uranium in small fractures, mainly as uranophane, but also as metatorbernite, metahaweeite, and carnotite.

Alaskite Geochemistry. Major and trace element compositions of mineralized and barren alaskites from the Rössing area are presented in Table 5-9. Silica and alkali contents range widely and the alaskites can be divided into quartz–plagioclase-rich compositions to quartz-poor–K-feldspar-rich compositions, through granitic compositions with about the same proportion of quartz, K-feldspar, and plagioclase (Fig. 5-18). Mineralized alaskites can belong to any group. Most of these granites have a low peraluminosity (low A parameter in Fig. 5-19 and low normative corundum in Table 5-9), and very low amounts of mafic minerals as shown by the low B parameter in the A versus B chemical mineralogical diagram (Fig. 5-19).

The low peraluminosity of these granites, despite their derivation from the partial melting of metasediments, results from the preferential melting of weakly peraluminous quartzofeldspathic metasediments or acidic metavolcanics, materials with nearly eutectic compositions. Another specific characteristic of this type of melt is the highly variable trace and major element composition of the pegmatoids that reflects the diversity of the material being melted

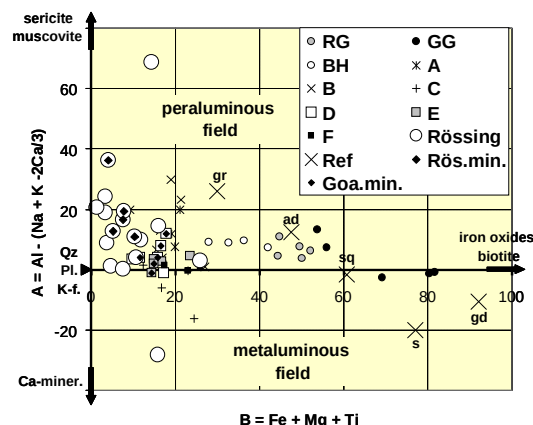


FIG. 5-19: Chemical-mineralogical diagram of representing the variations of the peraluminous index (A) versus differentiation index (B) for the granites of the Rössing area. Same legend as FIG. 6-18.

and the complex interplay between restite unmixing, mineral-melt fractionation, and fluid-rock equilibration between the injected sheets of granite and the host rocks of highly variable chemistry. Restite unmixing is the most viable mechanism from the common occurrence of biotite-rich schlierens and anorthite-rich plagioclase in the plagioclase-rich pegmatoids.

Uranium and thorium contents are not correlated (Fig. 5-20) and do not correlate with other incompatible elements.

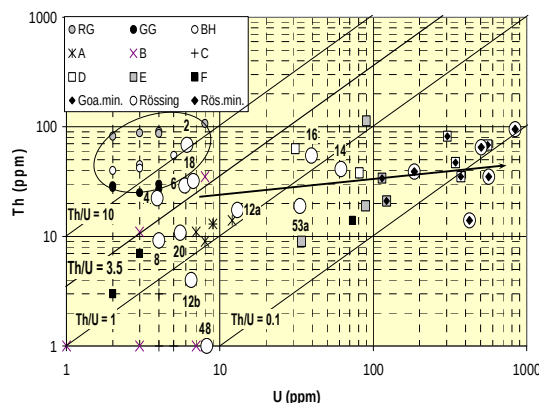


FIG. 5-20: Relations between U and Th for the granites from the Rössing area. Legend as Fig. 5-18. Average crustal Th/U = 3.5, arrow indicates U enrichment trend and circle is field of red and grey granites. Increase of Th limited by the solubility of Th-rich minerals in low temperature crustal melts.

TABLE 5-9: SELECTED ANALYSES AND NORMS OF ALASKITES FROM THE RÖSSING OPEN PIT AND GORGE DOME (CUNNEY UNPUB. DATA).

	2	8	10	14	16	47a	50	51a	52k
SiO ₂ (wt %)	69.02	65.16	80	77.76	84.61	76.89	74.09	73.35	76.45
Al ₂ O ₃	19.11	16.96	10.96	12.03	8.82	13.13	13.34	13.7	12.32
Fe ₂ O ₃ tot	0.36	0.39	0.34	0.16	0.18	0.13	0.25	0.76	0.28
MnO	nd	nd	0.01	0.01	nd	nd	nd	0.01	0.02
MgO	0.06	0.41	0.11	0.03	0.02	0.12	0.16	0.01	nd
CaO	0.88	1.83	0.3	nd	0.89	0.36	0.21	0.54	nd
Na ₂ O	8.01	2.46	3.86	3.29	3.25	3.6	2.53	2.84	2.84
K ₂ O	0.77	10.16	2.97	5.22	0.57	5.45	7.2	6.91	6.27
TiO ₂	0.67	0.05	0.04	0.04	0.05	0.04	0.04	0.04	0.04
P ₂ O ₅	0.07	1.1	0.02	nd	0.05	0.07	0.04	0.37	0.01
P.F.	0.24	0.62	0.6	0.64	0.75	0.7	0.76	0.31	0.39
TOTAL	99.14	99.14	99.21	99.18	99.19	100.49	98.62	98.84	98.62
FeO	0.04	0.1	0.09	0.04	0.06	0.09	0.14	0.16	0.22
Fe ₂ O ₃	0.32	0.28	0.24	0.11	0.11	0.03	0.09	0.58	0.04
Fe ²⁺ /Fe ³⁺	0.14	0.39	0.41	0.44	0.66	3.48	1.8	0.3	5.67
Fe/Mg	6.3	1.1	3.4	5.6	12.6	1.3	1.7	53.2	
Catanorm									
Q	17.78	10.48	45.46	38.62	61.76	34.33	31.19	30.15	35.78
Or	4.55	60.1	17.57	30.88	3.37	32.24	42.59	40.87	37.09
Ab	67.7	20.79	32.62	27.81	27.47	30.43	21.38	24	24
An	3.91	1.91	1.36	0	4.09	1.33	0.78	0.27	0
Di	0	0	0	0	0	0	0	0	0
Hy	0.15	1.13	0.27	0.07	0.05	0.37	0.52	0.27	0.37
Wo	0	0	0	0	0	0	0	0	0
Ma	0	0	0.21	0.05	0.05	0.04	0.13	0	0
il	0.08	0.09	0.08	0.08	0.09	0.08	0.08	0.08	0.08
He	0.32	0	0.1	0.08	0.08	0	0	0	0
Sp	0	0	0	0	0	0	0	0	0
Ru	0.63	0	0	0	0	0	0	0	0
Cor	3.66	1.19	0.89	0.95	1.35	0.81	1.08	1.43	0.84
Ap	0.15	2.4	0.04	0	0.11	0.15	0.09	0.81	0.02
% An	5.4	8.4	4	0	13	4.2	3.5	1.1	0
U (ppm)	6.1	5.5	185	61.6	39.5	423	504.5	840	565
Th	68.4	10.8	39	41.1	54.5	14	65	95.0	35
Th/U	11.2	20	0.2	0.7	1.4	0.03	0.13	0.11	0.06
Pb	9	24	388	206	16	33	49	60	37
V	27	11	<10	<10	<10	28	16	15	25
Ni	<10	<10	10	<10	22	<10	<10	<10	<10
Cr	17	<10	<10	<10	<10	<10	<10	<10	<10
Cu	<10	<10	<10	<10	<10	<10	<10	<10	<10
S	300	300	200	200	200	200	200	1600	500
Ba	64	314	67	81	72	55	94	205	96
Rb	14	794	200	395	44	319	493	508	439
Sr	34	41	42	30	51	23	20	64	46

nd = not detected.

Parameters controlling the formation of the Rössing deposit. A series of parameters have played a role in the formation of a minable deposit at Rössing:

Structural. At a regional scale, the accumulation of alaskite dikes forming the Rössing deposit is located at the south-western margin of a dome structure (Fig. 5-16). According to Anderson and Nash (1997), the emplacement of the Rössing alaskite bodies occurs within an ENE jog along a regional NNE sinistral strike slip structure, the “Welwitschia Trend” (Fig. 5-15), which separates two regions of different structural style, and may thus represent a detachment structure. Other alaskite intrusions are related to late sinistral transtensional ladder veins associated with the NNE sinistral strike slip fault, known as the Welwitschia lineament (Fig. 5-15). More recently, Basson and Greenway (2004) suggested that the alaskite emplacement is also closely linked to the late-kinematic evolution of the Rössing Dome, during a north-south transpressional tectonic episode, causing a rotation of the dome’s long axis, with a concomitant southward dome impingement. A wedge-shaped zone was produced, in which the alaskites were emplaced. From 3D modelling of the uranium content distribution at Rössing, Basson and Greenway (2004) proposed that the uranium mineralization is transgressive with respect to granites and their host lithologies. Such an interpretation is contradictory with the field observations that show that the mineralization is strictly confined to specific alaskite bodies with no mineralization in the hosting metamorphic rocks.

The proposed transgressive distribution of the mineralization results from an interpolation of the uranium contents measured in the exploration drill cores without taking into account the lithological boundaries. Basson and Greenway (2004) also proposed a control for preferential emplacement of the leucogranites in the SJ area by the late-D3 to D4 brittle-ductile deformation that was the site of protracted and elevated fluid flux. However, brittle-ductile fracturing represents a late feature, as it cannot occur in a high-grade metamorphic zone with incipient partial melting. The mineralized granitic bodies have the shape of elongated, centimeter to 100 meters thick, relatively boudinaged, and sub-parallel sets of pegmatoid dikes, the emplacement

of which is controlled mostly by the opening of the foliation of the enclosing metamorphic rocks under highly ductile conditions. The granitic dikes locally merge to form larger granitic bodies a few hundreds of meters thick, as observed in the western part of the Rössing open pit (Figs. 5-21 and 5-22). Such a phenomenon may explain the apparent transgressive character of the largest alaskite accumulations. Economically, they represent the most interesting ore bodies because of the paucity of U-poor host rocks intercalations.

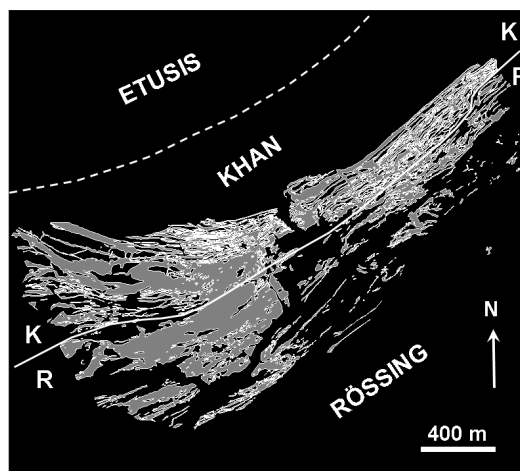


FIG. 5-21: Map of the Rössing deposit emphasizing the location of the alaskites at the interface between the Khan and Rössing Formations and their trend essentially parallel to the schistosity of the enclosing metamorphic rocks (modified from Berning *et al.* 1976).

Two of the largest alaskite bodies are located in the Rössing Formation at the level of the Lower and Upper Marbles (Fig. 5-22). Marbles being extremely ductile at high temperature, the structures controlling the opening of the metamorphic rocks have preferentially developed at the rheological transition between marbles and more competent gneisses. The other major alaskite body occurs in the only level of pyroxene-garnet gneisses of the Khan Formation, between two layers of pyroxene-hornblende gneisses that may also represent a zone of strong rheological contrast. Finally, if a protracted and elevated fluid flux had existed, as proposed by Basson and Greenway (2004), the U mineralization would not have been restricted to the alaskite bodies, but would have also occurred in the metamorphic host rocks.

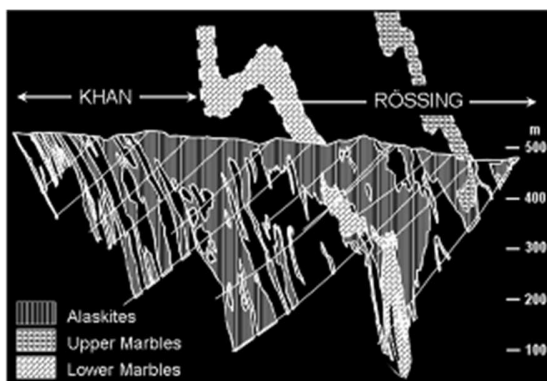
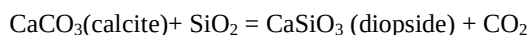


FIG. 5-22: North-south cross section through the Rössing deposit. The two marbles levels of the Rössing Formation are emphasized and their presumed extension is extrapolated above the outcrop surface to underline their role as a barrier for the ascension of the granite melts (modified from Berning *et al.* 1976).

Chemical (CO_2 and Ca activities in the melt). The carbonates have also exerted a strong chemical control on the accumulation of alaskite dikes in their vicinity. Many of the large alaskite bodies are located below the marble levels in the Rössing Formation at the Rössing mine, such as the SH anomaly and most of the G.P. Louw Deposits (see below). They also occur below the Husab marbles where the Rössing Formation does not exist, as at the Ida Dome. Carbonates are also common in the Khan Formation. The preferential accumulation of the alaskites in the vicinity of carbonate-bearing lithologies can be explained as follows. The solidus of a granitic system strongly increases when the molar fraction of CO_2 increases relative to water in the coexisting fluid. The development of skarns at the contact between injected silicate melts and carbonate-bearing lithologies produces decarbonation reactions such as:



The production of carbon dioxide reduces the water partial pressure in the metamorphic fluid in equilibrium with the silicate melts and thus, in the silicate melts themselves. The shift of the solidus of the silicate melts to higher temperature forces them to crystallize, and it is this process that lead to the alaskites observed at the level of the Rössing Formation. The marbles would thus have acted as a chemical barrier

favoring the accumulation of alaskite melts. This interpretation is supported by the decrease of the CH_4/CO_2 ratios in fluid inclusions in the mineralized alaskites of the Rössing open pit, compared to the trend of increasing CH_4/CO_2 ratios of the metamorphic fluid extracted from the metamorphic rocks along a profile from the Etusis to the Rössing Formations (Cuney, unpub. data). The carbon dioxide produced by the skarn reactions has probably migrated along the structural zones controlling the emplacement of the alaskites and thus, had an influence on alaskite crystallization over a much larger volume than the skarn formation zones. The preferential accumulation of alaskite bodies in the Rössing deposit is particularly evident in the vicinity of the Upper Marble Unit (Fig. 5-22). The second large accumulation of alaskite bodies occurs between the Khan and the Rössing Formation below the Lower Marble Unit, which may have extended above the present-day surface, as drawn in the section (Fig. 5-22).

The existence of chemical exchange between the marbles and the alaskite magma is indicated by high Ca contents in the uraninite crystals (up to 9.5 wt % CaO) and local occurrence of allanite, titanite, or amphibole in some of the dikes, some of which have a metaluminous chemistry (Fig. 5-19). In leucocratic granites, Ca contents of uraninite typically vary between 1 and 3 wt% CaO, and neither allanite nor titanite should be stable because of the low Ca contents of the silicate melts from which these granites crystallize (Cuney & Friedrich 1987).

Inasmuch as carbon dioxide-rich fluids have been observed along the entire lithostratigraphic sequence at Rössing and in other mineralized alaskite type occurrences (i.e. Mont Laurier; Cuney 1981; 1982; Bancroft, Baffin Island, Cuney, unpub. data), XCO_2 may have been instrumental in the genesis of mineralized alaskites by limiting the rate of melting of the metamorphic rocks in the source area. The alaskites consist of quartz and feldspars and few mafic minerals, indicating that the degree of partial melting has been low, which facilitated high uranium enrichments in the melts during anatexis.

Physical ($f\text{O}_2$). When a fluid phase exsolves from a silicate melt during the progressive crystallization of U-rich granite, part

of the U fractionates into the fluid phase and may escape from the system if the fO_2 conditions of the environment are oxidizing. The Rössing mineralization is located just below and within graphite-sulfide bearing schists and marbles of the Rössing Formation, the first reducing barrier encountered by the silicate melts during their injection in the lithostratigraphic succession. Sulfide enrichment also occurs in the upper part of the Khan Formation, as evidenced by a copper exploration shaft close to the Rössing and Ida mines. The increasingly reducing conditions from the base to the upper part of the lithostratigraphic succession are also indicated by the progressive increase of the CH_4/CO_2 ratios of the metamorphic fluid extracted from metamorphic rocks sampled along a section from the Etusis to the Rössing Formations (Cuney, unpub. data). The progressive increase of the CH_4/CO_2 ratios along the section shows that the methane-rich fluids tend to diffuse towards the Khan Formation and also probably upwards. Reducing conditions would limit the fractionation of uranium into the magmatic fluids, forcing uraninite to crystallize as disseminations within the alaskites, and inducing the precipitation of uranium from magmatic fluids exsolved from the silicate melts that crystallized in oxidized conditions at lower structural levels. This last type of uranium precipitation probably corresponds to the uraninite trails observed in the biotitic selvages of the alaskites, which has enhanced the grade of the mineralization.

Supergene enrichment. The upper part of the deposit has been enriched by the deposition of hexavalent uranium minerals (uranophane essentially) in the diclases of the alaskites derived from the alteration, during more humid climatic conditions in the past, of a presently eroded section of the alaskites.

Origin of the mineralized alaskites. Two models have been proposed for the origin of the uranium-mineralized leucocratic granite dikes of the Rössing type. They can result either from:

(i) extreme fractionation of a larger granitic body located at depth, such as proposed for Rössing by Bowden *et al.*, (1995), and mineralized pegmatoids from the Grenville belt in Québec (Lentz 1996),

(ii) or from partial melting of U-rich metasediments or metavolcanics, such as proposed for Rössing by Cuney, (1981; 1982) and

for the Mont Laurier alaskites and pegmatoids in Québec (Cuney 1982) or from the Wollaston Belt in Canada (Mercadier *et al.* 2013; McKechnie *et al.* 2013).

The partial melting model for the origin of the mineralized leucocratic granite dikes of the Rössing type is preferred for the following reasons:

(i) the larger granitic bodies that exist in the southern central part of the Damara belt are either older or younger than these dikes;

(ii) this type of granitic dike, globally parallel to the regional metamorphic foliation, always occurs in high-grade migmatitic domains where partial melting remains limited;

(iii) the foliation of the original metamorphic rocks is always well preserved, except where meta-arkoses or meta-acidic volcanics allowed high degrees of melting, as observed in the Etusis Formation of the Rössing area;

(iv) no correlation between uranium enrichment of the granitic dikes and other major and trace elements is evident, but should be if fractionation were important (Fig. 5-23, 5-24 and 5-25);

(v) mineralogy and major element contents define three groups of compositions: a plagioclase-rich group with a generally more restite-rich composition, a granitic group with near eutectic composition, and a K-feldspar-rich group generally poorer in quartz and coarsely pegmatitic;

(vi) there is evidence of restitic material, such as biotite schlierens, but also garnet in some occurrences, such as in the late orogenic granites of southern Finland;

(vii) internal zonation of the pegmatitic granitic dikes is generally absent;

(viii) rather than diverging from a deeper granitic body, the granitic dikes tend to merge to form small (10 to 100 m thick) granitic lenticular bodies;

(ix) the high and variable initial $^{87}Sr/^{86}Sr$ ratios (0.724 to 0.759) obtained on alaskites (only one from Rössing) implies an old crustal source for the magmas (Hawkesworth *et al.* 1981);

(x) their low initial $^{143}Nd/^{144}Nd$ ratios (0.51126 to 0.51117) and high TDM (2.34 - 2.35 Ga) indicate their derivation from Paleoproterozoic crustal sources.

However, as the Etusis sediments are derived from the Paleoproterozoic basement, and because of the diversity of the alaskite types,

more data is needed on the isotopic signature of the alaskites and of the Etusis siliciclastic sediments to better constrain the source of these magmas.

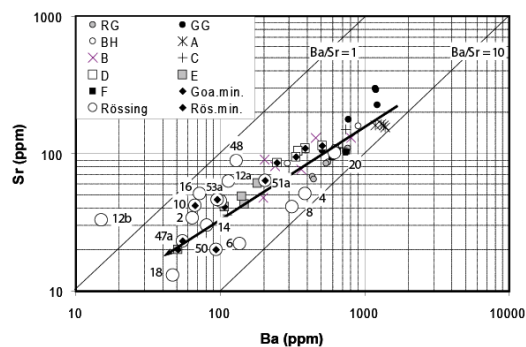


FIG. 5-23. Ba-Sr variations for the Rössing and Goanikontes granites. Same legend as Fig. 5-18.

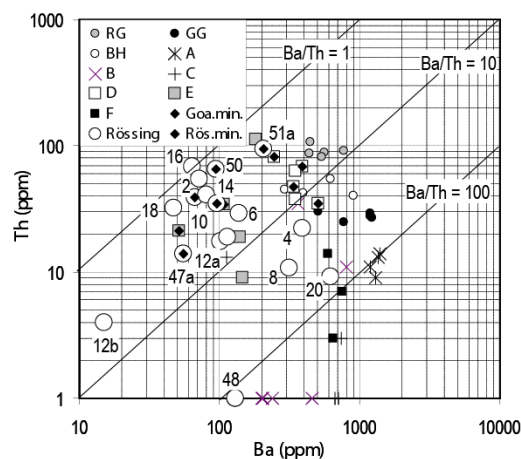


FIG. 5-24. Ba-Th variations for the Rössing and Goanikontes granites. Same legend as Fig. 5-18.

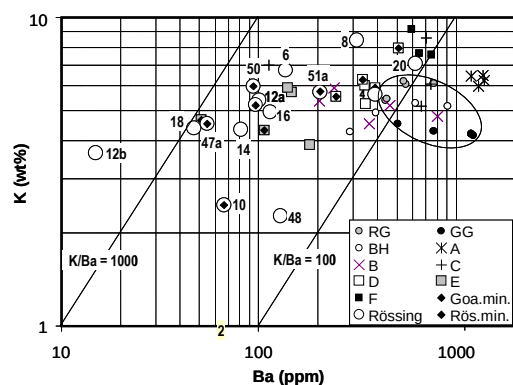


Fig. 5-25: K-Ba variations for the Rössing and Goanikontes granites. Same legend as Fig. 5-18.

Other world occurrences of Rössing-type mineralization

Disseminated uraninite mineralization in synmetamorphic pegmatoids are known worldwide and in most of the major orogens. They typically occur in metasedimentary associations initially corresponding to epicontinental successions comprising arkoses, quartzites, siltstones, carbonaceous shales, marls, and marbles metamorphosed to high-grade amphibolite facies with partial melting, and locally up to granulite facies. Occurrences of syngenetic uranium enrichment are sometimes still recognizable, mostly in quartz-rich meta-arkoses and calcsilicates, which have resisted melting because of their refractory compositions.

Archean Occurrences

Occurrences in the Litsk area of the NE Kola Peninsula, Russia, are probably the only known disseminated U-Th-REE mineralization related to pegmatitic granite with ages of 2.75-2.65 Ga located in a migmatitic domain of Archean age (Savitskii *et al.* 1995). In the literature, they are generally referred to as potassic metasomatites rather than partial melting products. The host rocks are metamorphosed peraluminous sediments (biotite, garnet, and two micas gneisses) and minor amphibolites of the Lower Tundra series. The mineralization is mainly located in the NW striking Voron'ya-Kolmozersk tectonic zone between the Kola and Murmansk cratons. The Dikoe Lake showing represents one of the most important examples, with mineralized pegmatoids occurring as a series of elongated lenses up to a few tens of meters long and a few meters wide. More than 400 uranium-bearing bodies have been discovered in two NW-trending, 6 km long and 1.5 km wide bands. Drilling indicates a discontinuous extension of these bodies down to 400 m. Their average U content is in the order of 100 ppm U and up to 0.2 wt% U, with relatively low Th/U ratios of 0.5 to 0.8. The main ore mineral is uraninite, locally concentrated in biotite selvages. Other U-Th minerals are monazite, uranothorite, and metamict zircon. Other minor minerals present are tourmaline, garnet, pyrite, magnetite, and ilmenite. Yttrium and REE concentrations may reach thousands of ppm.

Hudsonian S.L. Occurrences

The Wollaston and Mudjatik Domains in Saskatchewan, Canada, have metasedimentary associations corresponding to an epicontinental sequence metamorphosed to amphibolite-to-granulite accompanied by partial melting. Many occurrences of synsedimentary uranium enrichments occur in meta-arkoses at Duddridge Lake and calcsilicates at Burbridge Lake and Cup Lake (Parslow & Thomas 1982; Sibbald *et al.* 1976). Partial melting has produced pegmatoids and leucocratic granites rich in monazite and uraninite throughout the belt (Annesley & Madore 1999; Madore *et al.* 2000; Mercadier *et al.* 2013; McKechnie *et al.* 2013). The most common group are A pegmatites, described by Parslow and Thomas (1982), which have low Th/U ratios with the highest and most variable U contents (30 to 4650 ppm). The Charlebois occurrence represents the largest accumulation of pegmatoid dikes in the area, mineralized as disseminated uraninite $\pm$ molybdenite $\pm$ pyrrhotite $\pm$ pyrite $\pm$ magnetite (Mawdsley 1957). A thucolite-type occurrence was reported from one sample, underlining the importance of carbon-bearing sediments as a redox control for uranium mineralization in high-grade metamorphic conditions. Some of the mineralized zones may extend continuously over 1 km with a width of 25 to 35 m ("Row" showing) and average 960 ppm U over 21 m. Assay values up to 0.71 wt % U have been reported. Th/U ratios vary widely, from 0.1 to 4.4, but the average ratio of 0.3, based on 27 samples, is quite low. The mineralized zones are always located between calcic metasediments and the greatest uraninite enrichments are generally associated with biotite schlierens. These occurrences share many of the geological characteristics of the Rössing alaskites.

Steward Lake, Québec, Canada, is located 175 km NE of Schefferville in the Churchill Province. The mineralization is located in coarse-grained pegmatite subcrops and large boulders traced by prospecting over a distance of at least 2.6 kilometers and with a width of up to 700 meters (Freewest Resources Canada Inc).

In northern Québec, Ungava Bay and Baffin Island, uraniferous pegmatoids have been recently recognized over an extremely wide area and are being explored by several companies (Maurice 1979). They occur in the Paleoproterozoic metasedimentary Lake Harbour

Group comprising quartzites, graphitic gneisses, calcitic and dolomitic marbles, and calcsilicates: an association typical of a deposition in an epicontinental environment. Metamorphism during the associated Torngat orogeny at 1.85–1.74 Ga reached high-grade amphibolite facies with partial melting (Neto 2010).

In the Litsk district, Kola Peninsula, Russia, uraniferous pegmatoids, dated between 2.2 and 1.85 Ma, have been reported mainly at Polyarnoe and Litskoe. These are hosted by highly peraluminous sillimanite-biotite migmatitic gneisses. The mineralization occurs mainly as euhedral, Th-bearing uraninite. Remobilization of the uranium is locally observed in veins.

Mineralized pegmatoids in the Wheeler Basin, Colorado (Young & Hauff 1975) are hosted in migmatitic gneisses, and are particularly concentrated in zones intruded by the 1.4-Ga-old Silver Plume granite. Uraninite is mostly concentrated in biotitic schlierens. A major difference with the Rössing deposit is the high Th/U ratios of up to 10 (Phair & Gottfried 1964).

The Orrefjell mineralized pegmatites located 70 km north of Narvik, Norway, form a large (3 by 0.5 km) irregular body alternating with amphibolites and emplaced in mica schist that is rich in graphite, sulfides, and quartzites (Lindahl *et al.* 1985). The main uranium mineral is uraninite associated with molybdenite. The best hole gives intersections of 1m at 3,000 ppm and 2m at 2,300 ppm U.

The late orogenic potassic granites of Southern Finland are strongly enriched in uranium. They were emplaced between 1.85 and 1.79 Ga along a 100–150 km wide zone of Paleoproterozoic basement, extending for ca. 500 km from SW Finland to Russia, and represent one of the most U-Th-K enriched areas in Fennoscandia (Lauri *et al.* 2007). The basement is composed of Paleoproterozoic metavolcanic and metasedimentary rocks belonging to a presumed arc complex formed at ca. 1.90–1.87 Ga and accreted to central Finland during the Svecofennian orogeny at ca. 1.88 Ga. The mineralized granites are leucocratic, generally pegmatitic, and assumed to derive from a low degree of partial melting of supracrustal rocks during a low-P, high-T metamorphic event. They usually form relatively thin, undulating sheets, dikes, and pods, and commonly exhibit gradual transition with the migmatized host rocks, rarely

forming small granite plutons. The composition of the granites varies from plagioclase-rich to K-feldspar-rich, through eutectic granite compositions, as generally observed in migmatitic domains, with minor amounts of biotite (< 5 vol. %) $\pm$ muscovite $\pm$ cordierite $\pm$ garnet. Abundance and paragenetic association of magmatic U-bearing accessory minerals are highly variable from one intrusion to another, and include zircon, apatite $\pm$ monazite $\pm$ uraninite $\pm$ uranothorite $\pm$ allanite, and rare yttracolumbo-tantalite. Most granites are slightly peraluminous ($1.04 < A/CNK < 1.10$) and a few, with garnet and cordierite schlierens, may reach an A/CNK ratio of 1.55. Their variable U, Th, Zr, Y, and total REE contents reflect the extremely variable abundance of accessory minerals. Most samples have low Th/U ratios (≤ 1) indicating strong U enrichment relative to the average crustal Th/U ratio of 3.5. The late orogenic potassic granites of southern Finland share many characteristics with Rössing alaskites, but accumulations large enough to form an economic deposit remain to be discovered. The only significant accumulation of potassic pegmatites is represented by the Palmottu deposit, with about 1,000 t U at 0.1 wt % U (Räisänen 1989).

At Crocker Well, Olary Province, South Australia, uranium resources are estimated at 7,300 t U at 407 ppm U, found in six deposits. Mineralization is primarily thorian-brannerite with some davidites, which occur as disseminations in granites, but also as veins and breccias (Campana & King 1958; Thompson 1965).

At the Six Kangaroos area of Cloncurry-Mt. Isa District, Australia (Brooks 1960; Carter *et al.* 1961).

At Nanambu, Nimbuwah, and Rum Jungle complexes of the Katherine-Darwin area, Australia (Dodson *et al.* 1974; Ayers & Eadington 1975).

Grenvillian Occurrences

In the Grenville orogen an extremely large number of uranium occurrences related to anatectic pegmatoids and pegmatites are known (see Lentz 1996, for a review).

Bancroft, Ontario (Satterly 1957; Robinson 1960) was home to four mines, which produced a total of 5,700 t U between 1956 and 1982. The Faraday mine alone has produced 2,544,716 tons of ore averaging 0.1074% U_3O_8

(about 2320 t U). The principal ore minerals were uraninite, with up to 10 wt % ThO_2 , and uranothorite, whereas allanite, fergusonite, and betafite were more sporadic.

The Mont Laurier area in Québec, constitutes another major occurrence, with a decrease of the Th/U ratios of the pegmatoids from south to north that correspond to a variation of the U-Th mineral paragenesis from uranothorite only, to uranothorite + Th-uraninite, and to Th-uraninite only (Allen 1971; Cuney 1982; Fowler & Doig 1983).

Northern shore of the St Lawrence River. These occurrences include those at Johan Beetz, (Rimsaite 1981), Havre-St-Pierre, Sept-Iles, Port-Cartier, and St-Augustin with up to 2,920 ppm U and 995 ppm Th in pegmatites. Epicontinental platform sediments, as described in the Wollaston belt, typically host most, if not all, of these occurrences. Moreover, there are frequent indications of the former presence of evaporites, such as high-Mg gneiss, marialite-bearing gneiss, occurrence of anhydrite in the gneiss, or in the core of pegmatite, as at Bancroft, and of highly-saline brines in the pegmatites (Cuney, unpubl. data).

Pan-African Occurrences

In addition to Rössing, all other mineralized alaskite occurrences of the Damara belt are found in Namibia. Some are located in the vicinity of the Rössing mine, including the SH deposit, situated 1.5 km southwest of the Rössing ore body, with a size of 500 by 250 m and a high proportion of betafite, and the G.P. Louw deposits, a series of individual alaskite bodies 2 km east of the Rössing open pit, most of them emplaced in marbles of the Rössing or Karibib Formations. Others are located in the Central Damara zone along the same NE trend (Fig. 5-15).

The Valencia deposit, located about 35 km to the NE of Rössing, is much smaller with reserves of 12,000 t U at 0.105 % U, and probable resources of 12,000 t U at 0.1 % U. These estimates take into account mining to a depth of 350 m. The alaskites here tend to form a large body of 500 m by 400 m enclosing xenoliths of the host rocks, from tiny fragments to bodies several tens of metres long. Their strikes and dips do not vary significantly, when compared to those of the country rock. Little tilting and disorientation seems to have taken place during alaskite emplacement.

At *Ida dome*, mineralized alaskite occurs along the contact between the Khan Formation and marbles of the Husab Formation, the Rössing Formation being absent. Patchy radioactivity has been detected over 9 km along strike and over a width of 80 to 90 m. On the eastern side, skarn zones are developed against the marbles and can be also mineralized, when adjacent to mineralized alaskite. The skarns are generally small, although some of them reach several hundred meters in length and more than ten meters in width to the southeast of the *Ida Dome*. Biotite-schist schlierens in the mineralized alaskites are enriched in uranium, compared to the alaskites. The schists with K-feldspar porphyroblasts surrounding the alaskite bodies may be more mineralized than the intrusives over 1-2 meters from the contacts.

At *Goanikontes*, the major mineralized alaskite bodies are restricted to the lower part of the Khan Formation and occur within 400 m of the contact between the Etusis and the Khan Formation (Valois & Walgenwitz 1979; Mouillac *et al.* 1986). The Rössing Formation is very thin here. Surface exposures show that the alaskites do not merge significantly to produce granitic bodies as large as those in the Rössing, SH, or Valencia deposits. However, recent drilling showed mineralized sections of up to 124 m at 300 ppm U. Uraninite is the most abundant primary uranium mineral, but betafite is also present. Indicated resources are 4,960 t U at 200 ppm U and 22,800 t U at 170 ppm U.

At *Holland's Dome*, a weak occurrence, the mineralization has been reported in calcsilicates and skarns with higher grade (1.4 % eU) than in adjacent alaskite. The best drilling intersections are 221m at 430 ppm U and 29m at 1580 ppm U.

Husab(ex Rössing South) corresponds to a recent (2008) and major discovery by a junior company (Extract Resources) just 5 km south of the Rössing open pit under a shallow (50m) alluvial sand cover (Spivey *et al.* 2010). A definitive feasibility study demonstrates the technical and economic viability of two zones. Measured resources are estimated to 32,000 t U at 0.043% U, and indicated resources at 105,500 t U at 0.037% U (JORC and NI 43-101- compliant). Inferred resources are 50,000 t U at 0.029% U. The total proven resources are of 188,000 t U at 0.035% U to June 2011, with 100 ppm cut-off. The deposit is still open along strike and dip. The

higher grade of many zones of the Husab deposit may result from the emplacement of most of the alaskites within the Rössing Formation, which represents the best trap for the alaskite magma and the uranium mineralization as discussed above.

The *Inca* occurrence, located 40 km East of Walvis Bay, is unique in the Damara belt because the U mineralisation is entirely hosted in magnetite skarns. The adjacent alaskites are not significantly mineralized. In the others uranium occurrences from the Damara Belt, the mineralization is confined to the alaskite injections. Inca has 2,800 t U indicated, and 2,400 t U inferred resources at about 0.04% U.

The *Ongolo* deposit, contiguous to *Ida Dome*, is hosted in alaskite. It was discovered in 2010. The resources are 2,600 t U measured, 3,000 t U indicated, and 4,000 t U inferred, at 0.032% U along a strike length of up to 2 km.

At *Currais Novos, Brazil*, one occurrence of possible Rössing-type mineralization is known (Favali 1973; Ramos *et al.* 1974).

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