Evolution of metallogeny of granitic pegmatites associated with orogens throughout geological time

A. V. TKACHEV

Vernadsky State Geological Museum, Russian Academy of Sciences, Moscow, Russia
(e-mail: tkachev@sgm.ru)

Abstract: Since c. 3.1 Ga, pegmatite mineral deposits in orogenic areas have been formed throughout geological time in pulses, alternating with total absence of generating activity. The higher activity peaks of 2.65–2.60, 1.90–1.85, 1.00–0.95, and 0.30–0.25 Ga suggest a quasi-regular periodicity of 0.8 ± 0.1 Ga. This series is dominated by pegmatites of Laurasian blocks. The lower peaks at 2.85–2.80, 2.10–2.05, 1.20–1.15, and the higher one at 0.55–0.50 make up another series represented by pegmatites in Gondwanan blocks only. Each pegmatite class is characterized by a life cycle of its own, from inception to peak through to decline and eventual extinction. The longest cycle is recorded for the rare-metal class deposits, which first appeared in the Mesoarchaean and persisted through all the later eras, deteriorating gradually after the Early Precambrian. Muscovite pegmatites first appeared in the Palaeoproterozoic and reached the end of their life cycle at the Palaeozoic–Mesozoic boundary. The miarolitic class of pegmatite deposits in orogenic setting first came into being in the terminal Mesoproterozoic and dominated the pegmatite metallogeny of many Phanerozoic belts. The evolution of the pegmatite classes was controlled by the general cooling of the Earth and by associated changes in the tectonics of the lithosphere.

Supplementary material: Geochronological data used is available at http://www.geolsoc.org.uk/SUP18435.

Fersman’s fundamental works (1931, 1940) and Landes’ extensive paper (1935) were the first to present global-scale reviews of the distribution of all types of granitic pegmatites on continents and, most interestingly (for the purposes of this paper), throughout geological time. Schneiderhöhn’s book (1961) added little new to geochronological aspects of the topic. The paper by Ovchinnikov et al. (1975) was the first to demonstrate entirely new approaches to geochronological synthesis of this sort. Ovchinnikov et al. (1975)’s study assessed the intensity of pegmatite generation through the Earth’s history not merely in approximate terms (such as ‘many’, ‘few’, ‘very extensive’), but also presenting rigorous quantifications based on the geochronological data amassed by that time. Shortly after, Ginzburg et al. (1979) published one of the most important works analysing global pegmatite metallogeny. In addition to many other aspects, the authors carried out a seamless integration of geochronology and geology for pegmatite provinces worldwide, revealing evolutionary trends of the most important pegmatite classes (‘pegmatite formations’, to use the authors’ terms). Although the majority of the empirically established evolutionary trends remained unexplained, this was great progress in the field of pegmatite metallogeny. However, over the years, as new information was accumulated, it became clear that not all data used in the book were correct, in terms of present-day geochronological standards.

Since then, there have been no publications of detailed work in the field of the global evolution of granitic pegmatite metallogeny. Some studies addressed global aspects of selected pegmatite classes (Solodov 1985; Makrygina et al. 1990; Černý 1991b; Zagorsky et al. 1997, 1999; Shmakin et al. 2007; London 2008). These works analyse all issues pertaining to the intensity of granitic pegmatite generation in the geological record without offering numerical calculations. No previous attempt has been made to unravel the driving forces of the established evolutionary trends. The main purpose of this study is to bridge this gap and propose a new evolutionary paradigm.

Data sources

At present, the magmatic origin of granitic pegmatites is a matter of near total consensus. Granitic pegmatites are formed mainly in orogens as a result of crystallization of melts that are produced and variously evolved in thickened continental crust as a result of powerful heat generation (due to mechanical and radiogenic decay processes) and also to slow heat dissipation. In each particular orogen, synkinematic pegmatites are the earliest
type. They are common constituents of migmatitic fields. These pegmatites crystallize from poorly evolved melts. This is why they do not contain any specific minerals that might be used to distinguish them from ‘normal’ granites. The amount and especially the quality of economic minerals (such as K-feldspar, quartz, and muscovite) in these pegmatites, are not economically attractive. All of these pegmatites should be attributed to a separate class (non-specialized or non-mineralized pegmatites) and present no interest for pegmatite metallogeny studies. They are therefore not discussed in the analysis of pegmatite evolution below.

Crustal granitic melts keep on generating during the post-culmination (extension- or relaxation related) phase of orogen evolution, lasting up to 60 million years (Thompson 1999). These granitoids, not in all cases but quite commonly, are pegmatite-bearing. Some of the pegmatite fields are not accompanied by any reliably identified fertile granitic intrusions. Such relations are most common for deposits of the muscovite and abyssal feldspar–rare-element pegmatite classes (Ginzburg et al. 1979; Černý 1991a). Miarolitic pegmatites always show clear connection with their related granitoid massifs.

Many of these pegmatites related to the post-culmination orogenic phase are being mined or are of potential interest in the extraction of numerous rare elements, industrial minerals, gems, and specimens for collections. This genetic type of mineral deposits is of particular economic importance as a source of Ta, Li, Rb, Cs, various ceramic and optical raw materials, sheet muscovite (the only natural source), and crushed muscovite. In this paper, all pegmatites that show even the slightest potential for the extraction of these commodities are referred to as ‘mineralized pegmatites.’

Mineralization features exhibited in a pegmatite field depend on a number of factors. Amongst others, the crucial factors are fertile magma sources, P–T conditions and duration of melt evolution and crystallization, as well as host rock composition (Ginzburg et al. 1979; Kratz 1984; Černý 1991a, b, c). It is the mineralized pegmatites in late-orogenic to post-orogenic settings that are the focus of this study; these are here jointly referred to as ‘orogenic pegmatites.’

Pegmatites located in intraplate anorogenic granites (rapakivi, alkaline granites, syenites) may be of economic interest as sources of rare elements, feldspar raw materials, gems, and minerals for collections. However, the number of these deposits are small when compared to orogenic deposits; it was impossible to collect enough representative geochronological data to establish their generation scenario through geological time.

As with the crystallization of any granite, in general the formation of a pegmatite is quite a high-temperature process. Hence, the most reliable results are obtained from the study of U– (Th) –Pb isotope system on zircon, monazite-xenotime, tanta-
loniobates, and cassiterite, because of the high closure-temperatures and low susceptibility to external thermal and chemical influences (Faure & Mensing 2005). These features of the system are especially important, because pegmatites of some deeply generated fields may have remained in high-temperature conditions for time periods as long as tens of millions of years. Experience shows that Re-Os molybdenite dating results are reliable enough for the purposes of this study.

The K–Ar, Ar–Ar, Rb–Sr or Sm–Nd isotope systems are less resistant to external influences and have lower closure-temperatures. Hence, the results obtained by these methods do not always correctly reflect the time of pegmatite formation or crystallization of other magmas. This inconsistency was statistically confirmed by, for example, Balashov & Glaznev (2006). Nonetheless, some dating results obtained by these methods were used. However, this was only done in the absence of conflicting information and with at least partial support from independent geological and geochronological studies within the same pegmatite field. In most cases, these data are related to Phanerozoic pegmatites.

Unfortunately, there is no representative body of sufficiently accurate dating obtained by modern methods with a special focus on pegmatites. For this reason, in order to create a larger statistical sample, this study relies on the close genetic and temporal relationships between pegmatite fields and granitoid complexes, which differ from area to area. Where possible, age data was collected for those granites that occur within pegmatite fields and for those that are considered to be sources of the pegmatites that are under consideration. For a few provinces, geochronological data alone has been found for granites located outside pegmatite fields. These dates were used in case the researchers of the region were definite in the comagmatic origin of the granites with fertile granites from pegmatite fields. Only zircon and monazite age data were accepted in these cases.

Even the above ‘wide span’ approach to pegmatite geochronology did not enable the author to collect data precise enough to allow all the pegmatite fields but even some well-known pegmatite belts to be placed within any age range with an accuracy of 25 million years. In some cases, dating measurements from these areas have been made by obsolete methods, and in others, no reliable links between dated granitic intrusions and undated pegmatites from the same area have been
revealed. This is the case, in particular, with the Ukrainian Shield, most fields in the Palaeozoides of Central Asia and China, and the Mesozoides of Indo-China and adjacent areas.

In each particular pegmatitic province, only some veins and fertile granitic massifs have been dated, and the number of pegmatitic fields (and hence, the intensity of pegmatite-generating processes) differs essentially from province to province. For this reason, within each particular region (entire tectonic province or part thereof) the known dates have been extrapolated to all pegmatite fields that, according to alternative geological information, may be related to the same stage of generation.

All geochronological data and their sources, the main references to geological information on pegmatite deposits, and extrapolation results are presented in the supplementary material. Depending on the purpose of use, for this data it is distributed along the geochronological scale with steps ranging from 25 to 100 million years.

**Data verification: comparison with an independent database on crustal magmatism**

The only way to verify the reliability and representativity of the collected database on pegmatite geochronology for the purpose of global interpretation, is to compare it to an independent dataset that to some extent covers the subject of study. For example, this could be a database on the crustal magmatism on the Earth. Conveniently, a recently published analysis of a database on the terrestrial magmatism (Balashov & Glaznev 2006) contains processing results of crustal magmatic dates based on 9808 measurements, mostly U–(Th) –Pb ones. Figure 1 shows comparison between these results and processed dates from the supplementary material. It demonstrates a coincidence for all the main and most minor maxima and minima in both of the independently graphed diagrams. This result shows that the data collected and extrapolated are reliable and representative and can be used in global analysis and synthesis.

**Comparison to previous studies**

Apart from the above-mentioned publications (Ovchinnikov et al. 1975; Ginzburg et al. 1979), as far as the author is aware, there is no known published literature worldwide to propose a quantitative analysis of granitic pegmatite, in terms of the intensity of their development or evolutionary changes.

**Integrated results in Ovchinnikov et al. (1975)**

are based on a synthesis of 809 dates, including 340 ones from pegmatites of the USSR and the rest from elsewhere. Approximately half of the dates were acquired by the K–Ar method, about a third by the U–(Th) –Pb method (mostly on uraninite), and the rest by the Rb–Sr. Neither this paper nor the extended variant (Ovchinnikov et al. 1976) contain a table of measured ages. Only some of the data used in this study were obtained by the authors in their own laboratories. Some of the ages may have been obtained from non-mineralized pegmatites. At present it is impossible to clarify this issue because no list of references for these dates has been published. Unfortunately, the reference list for the dates used in the diagrams illustrating the generation intensity and evolution of granitic pegmatite classes (Ginzburg et al. 1979) is incomplete. At the same time, these materials show clearly that the ages used in this work were acquired only from the mineralized pegmatites with a very small number of pegmatites in anorogenic environs.

In both of these works, data was generalized with intervals as large as 100–200 Ma or even more (for some of the Early Precambrian periods). For a better comparison, our data in this case was also generalized with a step of 100 Ma. Then, all three diagrams were plotted in a single chart for comparison (Fig. 2).

Our comparison shows the presence of both similarities and significant differences between geochronological reconstructions. One principal similarity is the distinct pulsation in the intensity of pegmatite formation displayed in all cases through geological time. Another one is the non-ideal, albeit quite apparent, coincidence of peaks in the Neoarchaean (‘Kenoran’) and Palaeoproterozoic (‘Svecofennian’), at the Mesoproterozoic–Neoproterozoic boundary (‘Grenvillian’), at the end of the Palaeozoic (‘ Hercynian’), and in the latest Mesozoic–Cenozoic (‘Kimmerian–Alpine’). However, there are quite a number of discrepancies at some principal points, the most important of these are described below.

Geological time ‘infilling’ in Figure 2a is more complete than in the parts b and c of the chart. The diagram in Figure 2b is the most ‘rarefied’ among all the diagrams; this may result from the fact that Ginzburg et al. (1979) used only the data from significant (in the authors’ judgment) pegmatite provinces worldwide.

According to our data, the oldest mineralized (rare-metal) pegmatites originated in the Mesoarchaean c. 3.1 Ga ago (Fig. 2c). Judging by the results in Ovchinnikov et al. (1975), this event took place c. 0.3 Ga earlier (Fig. 2a), whereas, according to Ginzburg et al. (1979), it was 0.3 Ga younger.
In Černý’s well-known review (1991b) the first rare-metal pegmatite generation is attributed to the initial phase of the Kenoran orogeny, that is, c. 2.75 Ga.

Figure 2a has a notable peak at c. 2.3 Ga, which is missing in parts b and c of the figure. The «Pan-Brazilian» pulse (0.5–0.6 Ga) in Figure 2c is stronger than in the other counterparts.

The data in Figure 2c clearly fall into clusters. These clusters have certain signatures of quasi-regular periodicity. This feature is detailed in a special section below. Here it should be pointed out that neither Ovchinnikov et al. (1975) nor Ginzburg et al. (1979) discuss the issue of cyclicity or periodicity. This might be due to the fact that Figure 2(a, b) gives very little basis for such discussion.
The discrepancies described and some other minor differences, which an attentive reader can see in Figure 2, are primarily related to the accuracy of the data used. The works cited for comparison are based on research results from the period of geochronological studies when the precision of both tools and basic physical constants was far from the current high level. Besides, in obtaining magmatic ages, data from all methods (K–Ar, Rb–Sr, U–Pb) were used indiscriminately, without paying special attention to closure temperatures of isotopic systems or to likely disturbance and restarting of isotopic clocks. All these could lead to the widening of pulses on the time scale and to false peaks, which are most evident in Ovchinnikov et al. (1975). Besides, it is clear that some incorrect data was
used. Thus, the age of the Bernic Lake field in Ginzburg et al. (1979) is taken to be 2.0 Ga, while currently it is known to be 2.64 Ga (see supplementary material). In the same work, discrepancies for some other provinces are not so large, although still essential. Note that these discrepancies are non-systematic; that is, ages from Ginzburg et al. (1979) may be either younger or older than current datings: Wodgina, 2.7/2.84 Ga; Greenbushes, 2.7/2.53 Ga; Mama–Chuya belt, (0.7–0.4)/(0.38–0.33) Ga; and so on.

The onset in the Mesoarchaeon: why at that time?

The earliest occurrences of mineralized pegmatites appeared in the Barberton greenstone belt and the adjacent Ancient Gneiss Complex of Swaziland c. 3.1 Ga ago (Harris et al. 1995; Trumbull 1995).

All of them have typical features of the rare-metal pegmatite class, although their mining perspectives have never been highly valued. However, some of the deposits were sources of cassiterite placers in the Tin belt of Swaziland; these placers have been mined occasionally since the end of the 19th century (Maphalala & Trumbull 1998).

In order to understand why pegmatites were formed at this location, precisely at that period close to c. 3.1 Ga, one should compare the geology of the region with that of pegmatite provinces that developed prior to and during the same period and that contain no rare-metal-enriched pegmatites. Table 1 presents such comparison for the Isukasia ‘barren’ block and the Swaziland ‘productive’ block. These units are similar as both of them contain grey gneiss complexes (TTG: tonalite–trondhjemite–granodiorite complex) and supracrustal rocks (Isua and Barberton greenstone belts, respectively), as well as late- and post-orogenic

| Compared features | Isukasia block with the Isua GB (1–5) | Swaziland block with the Barberton GB (6–13) |
|-------------------|--------------------------------------|-------------------------------------------|
| Tectonic development | Discontinuous active inner tectonics c. 3.85–3.60 Ga; A few episodes of external stresses on a stabilized block during 3.6–2.55 Ga | Discontinuous active inner tectonics c. 3.66–3.08 Ga Anorogenic intraplate magmatism during 2.87–2.69 Ga |
| Main structural complexes involved and produced in orogenesis | TTG; Volcano-sedimentary supracrustals c. 3.8–3.7 Ga; Late to post tectonic granitoids and pegmatites c. 3.6 and c. 2.95 Ga | TTG; Volcano-sedimentary supracrustals c. 3.55–3.20 Ga; Late to post tectonic granitoids and pegmatites c. 3.10–3.07 |
| Greenstone belts | c. 35 km long and up to 2 km wide; Supracrustals up to 0.5 km thick: Volcanogenic/chemogenic/terrigenous ≈10/10/1 | c. 140 km long and up to 50 km wide; Supracrustals up to 12 km thick: Volcanogenic/chemogenic/terrigenous ≈10/1/7 |
| Meta-terrigenous rocks | Few dozens m thick c. 3.7 Ga; Volcanic rocks as an evident provenance only; High-ferruginous low-mature | Up to 6 km thick c. 3.26–3.20 Ga Essential up to main role of TTG in provenance One half is represented by mature low-ferruginous and low-calcium sediments with a big share of argillaceous varieties |
| Presence of highly evolved granitoids and mineralized pegmatites | No | Yes: c. 3.1–3.07 Ga |

Abbreviations: GB, greenstone belt, TTG, tonalite–trondhjemite–granodiorite complex (‘grey’ gneisses). Chemogenic rocks mostly include chert, BIF and carbonate ones. 1–5: Nutman et al. (1984, 2000, 2002); Hannmer et al. (2002); Friend & Nutman (2005). 6–13: Maphalala et al. (1989); Trumbull (1993); de Ronde & de Wit (1994); Harris et al. (1995); Trumbull (1995); Hofmann (2005); Hessler & Lowe (2006); Schoene et al. (2008).
granites. However, some differences are clearly apparent. The most important difference is that the Swaziland block contains large reservoirs of terrigenous rocks with a notable amount of high-maturity metasediments. Compared to their source rocks, these metasediments are enriched in K-feldspar and light-coloured mica and depleted in plagioclase and dark-coloured minerals (Hessler & Lowe 2006). The large proportion of these rocks points to deep chemical decay of the provenance rocks caused by weathering of voluminous continental masses. In the Tin belt of Swaziland, the fertile granite of the Sinceni field is the best studied geochemically, so far (Trumbull 1993; Trumbull 1995; Trumbull & Chaussson 1998). It displays all features of highly evolved granites melted from a source resembling these metasediments.

In southwestern Greenland (not only in the Isukasia block but in the whole Itsaq Gneiss Complex), there is no reservoir of terrigenous rocks comparable with that in the Barberton belt. This area is totally devoid of mineralized pegmatites, although it shows voluminous late-phase pegmatites related to the large (50 km × 18 km) post-orogenic c. 2.54 Ga Qørqut Granite Complex (Brown et al. 1981). It was formed by anatexis of the Itsaq gneisses and evolution of resultant melts (Moorbath et al. 1981). In the entire Archaean craton of Greenland, it is only c. 2.96 Ga late-tectonic granites in the Ivisârtq greenstone belt of the Kapîsîlik block (Friend & Nutman 2005) that are accompanied by two small groups of pegmatite dykes with sparse beryl crystals that are of mineralogical interest (Seacher et al. 2008). The Ivisârtq belt incorporates the largest Mesoarchaean supracrustal complex in the craton, but its metasedimentary constituents are not voluminous, which makes it similar to the Isua belt and different from the Barberton belt. The rare occurrence of beryl in pegmatites of the Ivisârtq belt and total lack of any rare-element minerals in pegmatites of the Isukasia block might result from compositional differences between metasedimentary rocks due to their different origins. However, no comparison has so far been made.

Besides the Itsaq Gneiss Complex, several smaller blocks older than 3.6 Ga are known worldwide. All of them are composed of broadly similar rock complexes (Nutman et al. 2001). No mineralized pegmatites are mentioned in the geological literature on these blocks, which provides a good reason for claiming their absence.

Therefore, the generation in the Earth’s crust of pegmatites with distinct features of the rare-metal class is restricted to those time intervals and areas in which the first large-scale terrigenous sediment accumulations occurred, that along with other supracrustal and infracrustal rocks, could have been affected by anatectic processes. Even though economically attractive granitic pegmatite deposits in orogenic belts may be located in quite different non-metaterrigenous rocks (amphibolites, anorthosites, marbles, etc.), closer inspection of each particular pegmatite-bearing province reveals considerable masses of metapelitic to metapsammatic rocks. This does not mean that metaterrigenous rocks are the only contributors to the production of fertile melts, but their input of fluid and ore-forming components into anatectic melts must be critical for the completion of the ore-forming process in a pegmatite chamber.

**Cyclicity in the metallogeny of granitic pegmatites**

The matter of cyclicity in the intensity of generation of mineralized pegmatites in the Earth’s crust has to the author’s knowledge not been discussed before. However, this kind of cyclicity has been established in the course of this study. The author has identified at least two cyclic trends. The cyclicity is more evident when the data collected is distributed with a step of 50 Ma (Fig. 3). The peaks in pegmatite generation intensity at 2.65–2.60, 1.90–1.85, 1.00–0.95, 0.55–0.50, and 0.30–0.25 Ga are the highest. If the 0.55–0.50 Ga peak is excluded, the rest of the peaks form a quasi-regular cyclic trend with a periodicity of 0.8 ± 0.1 Ga (Series 1). On the other hand, the 0.50 Ga peak together with the lower second-order peaks form another series with nearly the same periodicity: 0.55–0.50, 1.20–1.15, 2.10–2.05, and 2.85–2.80 Ga (Series 2). It is of special interest that peaks of Series 2 correspond to pegmatite fields of Gondwanan continental blocks only. The maxima of Series 1 are more varied, but the input of Laurasian continental masses is the most important. Hence, there is a certain lack of synchronism between these two large groups of continental blocks with regards to the position of pegmatite production peaks on the geological timescale.

If one compares this conclusion with the existing concepts of continental crust growth and supercontinental cycles (Condie 1998, 2002; Kerrich et al. 2005), the most active formation of pegmatite deposits occurred during the stages of the most intense growth of the supercontinents. Besides, the peaks at 2.65–2.60 Ga (Kenorland supercontinent) and 1.90–1.85 Ga (Columbia supercontinent) coincide with final phases of the most powerful pulses of growth of juvenile continental crust in the Earth’s history. Studies in younger epochs show no pulses of crust growth of the same extent, as the process was wavelike, with smoothed shape of the curve (Condie 2001). This means that the
formation of the younger supercontinents was not accompanied by intense growth of juvenile continental crust. However, the processes generating mineralized pegmatites were characterized by even stronger pulses in the Neoproterozoic and Phanerozoic. These pulses coincided with the formation of the Rodinia, Gondwana, and Pangaea supercontinents. As a corollary, ancient continental crust and its erosion products must have been in even greater predominance in fertile sources of granitic melts in post-Early Precambrian orogens, as compared with Archaean and Palaeoproterozoic ones.

It should be specially noted that ‘empty’ time gaps between pegmatite generation pulses became shorter and shorter over the course of time. Ultimately, since 0.6 Ga, such gaps are not observed at all at a data-generalization step of 50 Ma (Fig. 3). With a step of 25 Ma, two such gaps appear in the Phanerozoic interval, while the diagram for the earlier period becomes much more ‘sparse’ (Fig. 1). This frequency pattern suggests that the total continental crust area had reached a critical value by 0.6 Ga. Then, the interaction of continental blocks in collision belts created orogens at almost any period divisible by 25 Ma, with the resultant formation of pegmatite-hosted mineral deposits.

**Mineral deposits affiliated with the main pegmatite classes throughout geological time**

The modern classifications of granitic pegmatites (Zagorsky et al. 2003; Černý & Erict 2005) take into account geological settings favorable for pegmatite generating processes, as well as mineralogical and geochemical signatures. They are multi-branched and contain a number of hierarchic ranks such as classes (‘formations’ in Russian classifications), subclasses, types, and subtypes. Geochronological and geological data on pegmatite deposits amassed to date has allowed the author to trace evolutionary trends for the classes only. In general, the authors analysed the pegmatite classes identical to those from Černý & Erict (2005). However, some of the classes have changed names: in line with the wording used customarily in the Russian pegmatitic classifications (Ginzburg et al. 1979; Zagorsky et al. 1999, 2003), the modifier ‘rare-element’ has been changed for ‘rare-metal’.
Besides, the authors have been addressing rare-metal–miarolitic pegmatites, which are important for the analysis carried out in this study. They are not specified in the classification by Černý & Ercit (2005) within either the rare-element class or the miarolitic one. Zagorsky et al. (2003) pick out miarolitic varieties in almost all the pegmatite classes and consider these varieties to be in a special, additional classification, as parts of corresponding classes. In this classification the rare-metal–miarolitic pegmatites are placed in the rare-metal class (formation) of the basic classification. The author has followed this definition of the term.

Rare-metal pegmatites first appeared in the Mesoarchaean inside and south of the Barberton greenstone belt c. 3.1 Ga ago (see above) and continued to form in later eras (see supplementary material). Only at the very end of the Mesoarchaean in granite-greenstone belts of the Pilbara craton, were the first pegmatites formed, with accumulations of rare metals, reaching the values that were attractive to start hard ore mining for Ta, Sn, and minor Be. The most notable deposit of this kind is Wodgina–Cassiterite Mountain. According to the statistics on the USGS website, the mine has produced up to 25% of the world’s primary tantalum over the last five years. Albite–spodumene and albite-type pegmatites prevail among economically attractive bodies in the Pilbara craton (Sweetapple & Collins 2002). Complex-type pegmatites are also known, but they play no essential role in rare-metal reserves of the region.

All types of rare-metal pegmatites have been established for the Neoarchaean as well. Unlike the previous era, this one is earmarked by complex-type deposits, with Tanco (Li, Ta, Cs, Be), Bikita (Li, Cs, Be, Ta), and Greenbushes (Li, Ta, Sn) being the brightest examples. These deposits show extremely high degrees of differentiation of their inner structure (Martin 1964; Partington et al. 1995; Černý 2005) and display the world’s highest ore grades of Li, Ta, and Cs in the whole exploration history of pegmatite deposits. The Palaeozoic types of rare-metal pegmatites do not differ from the Neoarchaean ones, but no deposits with equally high-grade mineralization have been found within them.

Some rare-metal pegmatite bodies are of mining interest not only for rare elements, but also for gems and high-priced mineral collections specimens from residual miarolitic cavities. In the prevailing number of such pegmatite bodies the latter represent a greater economic interest than the former. These pegmatites appeared for the first time in the terminal Precambrian in the fields of the Eastern Brazilian pegmatite province (Morteani et al. 2000; Pedrosa-Soares et al. 2011). Thus, at that time, a new intermediate rare-metal–miarolitic type of pegmatites appeared. Vugs in the Archaean and Proterozoic rare-metal pegmatites are not abundant and are only of scientific interest. These vugs are not residual cavities: they are small, are not attached to core zones, and were produced by the leaching activity of relatively low-temperature fluids. For instance, Stillling et al. (2006) mention rare vugs with only low-temperature mineral lining in the Tanco pegmatite, which are concentrated in the pegmatite’s upper and central intermediate zones. The author has not succeeded in finding any published descriptions of high-temperature mineral associations in the vugs of such ancient pegmatites from orogenic settings.

Since the Neoproterozoic–Palaeozoic boundary, rare-metal–miarolitic pegmatites have been routinely formed, and with time they came to dominate in the Phanerozoic belts over the rare-metal pegmatites (supplementary material). Besides, global analysis of the rare-metal pegmatite class shows concurrent gradual degradation of the pegmatites’ inner zoning. This is most clearly manifested in the general decrease of the average amount of minerals in pegmatites and in the increasing prevalence of very primitively zoned albite–spodumene type bodies over the better zoned types in all pegmatite provinces (Solodov 1985). The appearance of an exotic variety as aphanitic pegmatite dykes in the Hindu Kush belt (Rossovskyi et al. 1976) may be viewed as a climax of the trend. Solodov’s conclusion (1985) about the total extinction of the complex-type pegmatites in orogens by the Cenozoic is evidence of general degradation of pegmatite-forming melts in terms of their geochemical evolution. So, in the course of geological time, there is a distinct general change for the worse, in the chances of the correct conditions to evolve, especially to crystallize pegmatite-forming melts.

Note that the oldest rare-metal granites in orogenic belts (Abu Dabbab and their counterparts in the Eastern Desert of Egypt, Taourirt in the Hoggar Mts, etc.) were formed at the waning stages of some orogens in the Early Palaeozoic (Abdalla et al. 1998; Kesraoui & Nedjari 2002). These granites are ore-bearing for Ta, Sn ± Li, Be. Besides, they are quasi-synchronous to rare-metal pegmatites in other parts of the same provinces. Rare-metal granites and pegmatites are very similar in terms of petrology, mineralogy, and geochemistry (Beskin & Marin 2003). The number of rare-metal granites increased manifold in the Hercynides and Mesozooids. The large Alakha (Li, Ta) deposit generated in the Altai orogen at the Triassic–Jurassic boundary is represented by a spodumene granite unknown in any earlier epoch (Kudrin et al. 1994). In deposits predating this boundary, spodumene is known only from pegmatites. This granite differs from
some of the albite-spodumene pegmatites only by a plug-shaped morphology, smaller-sized minerals (0.1–10 mm, mainly 2–3 mm), and a stronger primitive zoning revealed only from sampling results (Kudrin et al. 1994).

**Muscovite pegmatite** deposits are the main suppliers of sheet muscovite and practically the only source of high-quality (low-defect) large sheets of muscovite. They are usually barren of rare-metal minerals, except for some known cases of scarce accessories. The colour of the sheet muscovite is light brown, reddish brown, and sometimes light green. **Muscovite–rare-metal pegmatite** deposits, which are also mined for the same purpose with a non-systematic co-production of some rare-metal minerals (usually beryl), mostly contain books of greenish or whitish muscovite of a lower quality.

The first muscovite pegmatite deposits were formed in the Belomorian and East Sayan belts in the Palaeoproterozoic c. 1.87 Ga ago (see supplementary material). They are located in metamorphic formations with a great share of aluminous (two-mica ± garnet ± kyanite/sillimanite) middle-amphibolite-facies paragneisses and schists of kyanite–sillimanite (Barrovian) series in elongate fold belts (Ginzburg & Rodionov 1960; Sal’yé & Glebovitsky 1976). Occurrences of this kind of metamorphic rocks are known in the Archaean structures but are not numerous, and they are relatively small in area (Percival 1979 and references therein). They appear to be related to partial convective overturn adjacent to Archaean rising granitic domes rather than to collision belts (Collins & Van Kranendonk 1999). The size of the Barrovian-type metamorphosed blocks created due to this tectonic scenario and the duration of favorable conditions were probably not sufficient for the muscovite pegmatite deposits to be generated there.

The average quality of the sheet muscovite in the Belomorian belt deposits is the highest in the world (Tkachev et al. 1998). The Bihar Mica belt, famous for its deposits, essentially exceeds the Belomorian one and any other of the Palaeoproterozoic provinces in terms of resource abundance, although it is slightly inferior to its older counterparts listed above in terms of the average quality of muscovite sheets: birth defects in mica crystals (colour zoning, ‘A’-structure, staining, microintegrowths with other minerals, etc.) are more abundant here. Muscovite pegmatite deposits keep this bias tendency: in general the share of high-grade crystals (in terms of sheet mica quality) in the best Palaeozoic deposits of this class is notably lower than in the Proterozoic ones (Tkachev & Gershenkop 1997; Tkachev et al. 1998). Probably the youngest (Permian) deposits of the muscovite pegmatite class are located in the Urals belt (supplementary material). Historic archival exploration data from the deposits show that only few bodies in them contain vanishingly small amounts of high-grade sheet muscovite. The rest of pegmatites contain only low-grade muscovite. No Mesozoic or Cenozoic muscovite pegmatite deposits have been found worldwide so far. Merely small-scale accumulations of light-coloured sheet mica that have been found in their host muscovite–rare-metal pegmatites, requiring a slightly lower lithostatic pressure to be generated as compared to muscovite pegmatites proper. Although Barroavian-type metamorphic complexes in orogens continued to appear even in the Late Cenozoic, and mineralized granitic pegmatites are widespread in them. However, these orogens are not fertile for muscovite pegmatites proper. Hence, there are many reasons for supposing cessation of this pegmatite class deposits in the crust after the Palaeozoic.

For example, numerous pegmatites are known in the Miocene-age Muzkol Metamorphic Complex of the Barrovian type in the Pamirs (Dufour et al. 1970), but none of them contain sheet mica zones! The pegmatites with mialoritic mineralization are the most noted in the region (Zagorsky et al. 1999). Besides, some bodies contain uneconomic rare-metal mineralization. Small deposits of sheet mica (Zagorsky et al. 1999) are hosted by a rock complex of the same type and of similar age in the Neelum River valley, High Himalaya (Fontan et al. 2000). These deposits belong to the muscovite–rare-metal class, and, in addition, a number of them contain mialoritic vugs with gems (Zagorsky et al. 1999).

The most ancient (c. 1.73 Ga) mialoritic pegmatites proper, that is, those that are attractive for mining exclusively because they contain crystals lining cavities (mainly residual ones), are related to anorogenic rapakivi granites of the Ukrainian shield (Lazarenko et al. 1973). Some other occurrences of ancient mialoritic pegmatites are known in rapakivi granites of the Fennoscandian shield (1.67–1.47 Ga) with most notable ones in the Wyborg batholith (Haapala 1995). It is of specific interest that in post-orogenic settings this class of pegmatites appeared only at the end of the Mesoproterozoic. According to the data collected (supplementary material and references therein), the most ancient pegmatite fields of this category are hosted by the c. 1.07 Ga old Katemcy and Streeter granites in the western part of the Llano Uplift in Texas. The post-orogenic nature of the granites is reliably established (Mosher et al. 2008). These deposits were mined to extract black quartz (morion) and jewel topaz crystals (Broughton 1973). Of the rare-metal minerals, only cassiterite was recorded. Note that rare-element pegmatites in the NE of the uplift (Ehlmann et al. 1964; Landes 1932) are related to the Lone Grove
granite with major-element chemistry very similar to the Katemcy pluton but its age is c. 20 Ma older (Rougvie et al. 1999). Here, pegmatites are located in the granites as well as in the country gneiss of the upper amphibolite facies. Topaz is unknown, and vugs are very scarce, although one of them has a notable size (Landes 1932). Hence, the conditions of melt differentiation and crystallization over a period of c. 20 Ma changed in such a manner that petrologically similar granites gave birth to metallogenically different-class pegmatites.

Later on, miarolitic pegmatites have been getting progressively widespread in orogenic belts in course of time. It is possible to claim the same about the intermediate rare-metal–miarolitic class (see above in this section). Since the Late Palaeozoic (c. 300 Ma) these classes have been prevailing in mineralized pegmatite fields. At the same time, these pegmatites have been slowly diminishing in size, with their inner zoning becoming progressively less developed, in keeping with the trend mentioned above for rare-metal pegmatites.

### Driving forces of the metallogenic evolution of granitic pegmatites in orogenic belts

Evidence of global changes in the conditions of pegmatite crystallization is provided by the analysis of the following features described in a section above (either considered separately or analysed jointly for increased benefit): (i) gradual deterioration (degradation) of pegmatites of the rare-metal class from the Neoarchaean to the Cenozoic; (ii) the restriction of the highest-grade sheet mica deposits to the Proterozoic and the total absence of muscovite-class pegmatites in post-Palaeozoic orogens; (iii) the first appearance of miarolitic pegmatites in the Late Mesoproterozoic and rare-metal–miarolitic ones in the Late Neoproterozoic; and (iv) progressively increasing proportion of both of these classes in mineralized pegmatite fields from the Cambrian to the Neogene. The first wide-scale appearance of rare-metal granites in the beginning of the Palaeozoic and their increasing abundance in the Phanerozoic orogens on the periphery of pegmatite belts and sometimes instead of them are thought to be part of the same sequence of interconnected events.

The degradation of rare-metal and sheet mica deposits may be tentatively explained by the well-established gradual cooling of the Earth and the decrease in the mean value of the lithospheric heat flow (Taylor & McLennan 1985). On the one hand, the cooling impairs the conditions for crystallization and differentiation of pegmatite-forming melts both during preliminary stages and in the final pegmatite chambers, because the decrease in the global heat flow reduces the possibilities for high-T fields (which make for low heat conductivity in pegmatite-hosting rocks at crustal levels (7–12 km) in orogens, favourable for the rare-metal class formation) to persist over a sufficiently long time. On the other hand, sufficiently lasting existence of extensive areas with Barrovian-type middle amphibolite facies conditions in the middle crust of extending orogens (16–22 km) became possible when at c. 1.9 Ga in the Palaeoproterozoic the mean heat flow values dropped below the Archaean ones. This is the condition for the formation of the muscovite-class pegmatite deposits, because the growth of large (up to 2–3 m²) low-defect mica crystals requires long-lasting high partial pressure of H₂O, which cannot be reached in fertile undersaturated melts without a high enough lithostatic pressure in the country rocks (Tkachev et al. 1998).

However, it is impossible to explain why miarolitic pegmatites appear precisely in the Grenvillian orogens by invoking the lithospheric cooling alone. The formation of residual miarolitic cavities in pegmatites results from H₂O supersaturation of fertile melts (London 2005). If the vugs clustering in the central (core) pegmatite zone are abundant or scarce but sizeable (n-10 n m³), this most likely implies great H₂O supersaturation during subliquidus crystallization, that is, intra-chamber boiling of partially crystallized melt. The higher the supersaturation, the larger the bulk share of miarolitic cavities in a pegmatite. In theory, so ‘sudden’ appearance of miarolitic pegmatites at the end of the Mesoproterozoic and their progressively wider spread in the younger epochs can only result from two factors.

Firstly, it may be supposed that Early Precambrian pegmatite-forming melts were lower in H₂O. Over time, the melts became increasingly enriched in H₂O and reached saturation, and as a result, boiling in the course of intra-chamber crystallization became increasingly common. Secondly, in view of the well-known fact that H₂O solubility in granitic melts decreases with lithostatic pressure (Luth et al. 1964; Luth 1976; Huang & Wyllie 1981; Holtz et al. 1995), the above evolutionary trend is possible to assume that in any given era pegmatite-forming melts had more or less the same H₂O concentrations, but since the end of the Mesoproterozoic these melts crystallized at increasingly shallower depths. If this is so, the pegmatite formation maximum shifted closer to the surface, that is, to a zone with P–T conditions where the boiling limit was lower, while thermal field gradients in the country rocks of crystallizing pegmatites were higher when at greater depths.
The first assumption (variations in H₂O concentration in melts as a function of age) is not very plausible. The second one is more realistic.

The last two decades have seen an increase in thermochronological studies in orogens (e.g., Hodges 2003) and in mathematical modelling of orogens based on realistic physical parameters (e.g., Mareschal 1994). The results of these studies make it possible to calculate exhumation (uplift) rates of extending orogens on a reasonable basis. Unfortunately, for well-known reasons, these studies have focused on Phanerozoic fold belts which describe the number of rate estimates, summarized for different epochs in Table 2. Nevertheless, all the main epochs are (to a varying extent) characterized by uplift rates of orogen roots in the course of the post-culmination extension.

These data provide weighty support to the second assumption: the likely decrease in crystallization depths of pegmatite-forming melts through geological time. Table 2 clearly shows notable differences in exhumation rates of orogens: all Early Precambrian rates are below 0.35 mm × a⁻¹ with a mean value of c. 0.2 mm × a⁻¹ (0.2 km × Ma⁻¹ or 2 km per 10 Ma), whereas, the rates for younger belts are higher by a factor of 3 or more. Some of the exhumation rates in Phanerozoic belts were calculated for relatively short periods of 2–5 Ma. Even when the data is smoothed over longer periods of 10–20 Ma, which is taken to be

| Tectonic structure | Calculated rates*, mm a⁻¹ | References |
|--------------------|---------------------------|------------|
| Cenozoic           |                           |            |
| Himalayan belt, Namche-Barwa syntaxis | 3–10 | Burg et al. (1997) |
| Himalayan belt, Nanga Parbat syntaxis | c. 5 | Shroder & Bishop (2000) |
| Himalayan belt, South Tibet junction | 1–5 | Ruppel & Hodges (1994) |
| Himalayan belt, Zanskar | 1.0–1.1 | Searle et al. (1992) |
| Black Mountains area | 2.3–3.2 | Holm et al. (1992) |
| Ruby Mountains area | 1.33–5.8 | Hacker et al. (1990) |
| Omineca belt, Idaho batholith | 0.4–1.6 | House et al. (2002) |
| Pyrenean belt | c. 2 | Sinclair et al. (2005) |
| Mesozoic           |                           |            |
| Cordilleran belt, Sierra Nevada batholith | 0.35–1.33 | DeGraaff-Surpless et al. (2002) |
| ibid               | 0.5–1.0                   | Vermeesch et al. (2006) |
| Qinling-Dabie-Sulu belt, Dabie zone | 1–8 | Ayers et al. (2002) |
| Late Neoproterozoic – Palaeozoic |            |            |
| Altai belt         | 1.75–1.82                 | Briggs et al. (2007) |
| Appalachian belt, Acadian orogeny | 1–2 | Hames et al. (1989) |
| Ibid               | c. 1.4                    | Armstrong & Tracy (2000) |
| Variscan belt, Iberian crystalline massif | 0.6–1.3 | Martínez et al. (1988) |
| Variscan belt, French Massif Central | 0.3–1.5 | Scaillet et al. (1996) |
| Variscan belt, Bohemian crystalline massif (north-western part) | 1.1–2.5 | Zulauf et al. (2002) |
| Variscan belt, Bohemian crystalline massif (eastern part) | 2.8–4.3 | Kotková et al. (2007) |
| Late Mesoproterozoic – Early Neoproterozoic |            |            |
| Grenvillian belt, western part | 0.33 | Martignole & Reynolds (1997) |
| Grenvillian belt, eastern part | 0.41–1.22 | Cox et al. (2002) |
| Sveconorwegian belt, Bamble zone | 1.5–1.0 | Cosca et al. (1998) |
| Sveconorwegian belt, Idefjorden zone | c. 1 | Söderlund et al. (2008) |
| Palaeoproterozoic |            |            |
| Athabasca round-basin area, Hearn-Rae junction (1.80–1.85 Ga interval) | <0.2 | Flowers et al. (2006) |
| Svecofennian belt | 0.1 | Lindh (2005) |
| Belomorian belt | 0.06 | Alexeev et al. (2003) |
| Limpopo belt | c. 0.3 | Zeh et al. (2004) |
| Neoarchaean |            |            |
| Yellowknife belt | 0.15–0.35 | Bethune et al. (1999) |

*All rates are given with a precision shown in the referred works.
a model time-gap between the termination of collision and the start-up of granite magmatism (Thompson 1999 and references therein), they give uplift rates that are still higher than those of the Early Precambrian.

Thus, starting in the Grenvillian epoch, anatectic granitic melts had more opportunities to penetrate from their sites of origin at depths greater than c. 15 km (Brown 2001) into the upper levels by means of passive transport along with country rocks. This process was further facilitated by the wider distribution of brittle deformation at these conditions, providing additional magma conduits to the uppermost crust horizons (Thompson 1999). As discussed above, low-pressure settings are very favorable for large-scale crystallization of miarolitic pegmatites or even rare-metal granites, rather than pegmatites. Since the Mesozoic, the uplift rates became so high as to leave insufficient time for muscovite class pegmatites to originate at a favourable depth in kyanite-sillimanite metamorphic complexes. Instead, the other types of pegmatites were generated, including those with miarolitic cavities. The youngest known pegmatites of the abyssal feldspar–rare-element class are the Ordovician ones (supplementary material); the author believes that the considerations above are also applicable in this case.

It is not easy to unambiguously define the factors responsible for the changes in the post-culmination behaviour of orogens. This issue may conceivably be unravelled by analysing the lithosphere thermodynamic models presented by Poudjom Djomani et al. (2001). These models were developed using the subcontinental lithospheric 4D mapping technique based on mantle xenolith data (O’Reilly & Griffin 1996). According to these models, subcontinental lithospheric mantle (SCLM) of Archaean age (>2.5 Ga) has considerable buoyancy relative to its underlying asthenosphere. The Proterozoic (2.5–1.0) SCLM is slightly thinner and has somewhat lower density parameters than the Archaean one. Nevertheless, it is buoyant relative to the asthenosphere within any reasonably possible thermal fields. As for the Grenvillian–Phanerozoic SCLM (<1 Ga), it is the thinnest, the densest, and has the greatest gradient in terms of the vertical distribution pattern of density. In general, this SCLM is always buoyant only where lithospheric geotherms are elevated, as in the Cenozoic active volcanic provinces. However, as the geothermal gradient relaxes toward a stable conductive profile during orogenic post-culmination extension, SCLM sections thinner than c. 100 km become denser than the asthenosphere or, in other words, negatively buoyant, and as a result the whole lithosphere becomes gravitationally unstable because of a heavy lithospheric root. This could promote delamination of the SCLM in all or some of its parts, upwelling of asthenospheric material, and fast uplifting of the crust. Hence, it does not seem to be mere chance that the generation of the first miarolitic pegmatites started exactly at the Mesoproterozoic–Neoproterozoic boundary, when the continental lithosphere in orogens became unstable because of such a pattern of density distribution. In this connection, it is apropos to point out that a recently developed tectonic reconstruction for the Llano Uplift area in the Grenvillian epoch perfectly fits such a kind of scenario (Mosher et al. 2008).

The changes in SCLM are conditioned by different levels of depletion depending on variations in the volumes and temperatures of mafic–ultramafic melts in different epochs, which is, in turn, a consequence of the Earth’s cooling throughout geological time (Poudjom Djomani et al. 2001). Therefore, this evolutionary trend in pegmatite metallogeny (the appearance of miarolitic pegmatites, extinction of abyssal and muscovite pegmatites) is also related to the same factor that was proposed as the most important cause of the changes (general simplification of zoning, widespread vugs) in the deposits of the rare-metal pegmatite class. However in this case it acts through a more complex chain of processes. No doubt, this chain, amongst other things, also played a role in the evolution of the rare-metal pegmatite class.

Conclusions

The collected data on geology and geochronology of different mineralized pegmatite classes have made it possible to correlate the evolutionary trends of pegmatite metallogeny in orogens with the global evolution of the lithosphere. The metallogeny shows two principal trends: (a) a pulsatory pattern with quasi-regular periodicity; and (b) unidirectional development of pegmatite classes from their inception to their extinction.

The pulses or cyclicity in pegmatite generation are in correlation with the extent of crust magmatism as well as supercontinental cycles. The gradual growth of pegmatite generation is in perfect alignment with the existing concepts of continental crust growth from the Archaean to the Cenozoic. Different factors are responsible for the diversity of pegmatite classes, including compositions of supracrustal country rocks, metamorphic facies series in orogenic belts, and uplift rates during orogenic extension. As these factors changed, the classes of mineralized pegmatites also changed their aspects. These changes controlled life cycles of the classes, from inception to peak through to decline and eventual total extinction. The metallogeny of rare-metal class pegmatites is characterized by the longest life cycle. Generation of these
pegmatites began in the Mesoarchaean and persisted through all the later eras, to wane gradually after the Early Precambrian with the eventual strong degradation of their zoning structure, as observed in Cenozoic deposits. Since the terminal Neoproterozoic, rare-metal–miarolitic pegmatites have been generated more and more frequently inside or instead of rare-metal pegmatite deposits. Mineral deposits of the muscovite pegmatite class appeared for the first time in the second half of the Palaeoproterozoic and came to the end of their life cycle at the Palaeozoic–Mesozoic boundary. Miarolitic pegmatite deposits first appeared in anorogenic settings in the Late Palaeoproterozoic, whereas, in post-orogenic plutons they occurred first in Grenvillian-aged structures and dominated throughout the pegmatite metallogeny of many Phanerozoic belts.

All these changes were ultimately induced by the general cooling of the Earth.

This study was supported by grants from the Russian Academy of Sciences and the Russian Ministry of Education and Science (State Contract # 02.515.12.5010). I am deeply grateful to I. Kravchenko-Berezhnoy and V. N. Endo for providing the data from mineralized pegmatites and fertile granites worldwide would be very much appreciated.

I am deeply grateful to M. Cronwright, T. Oberthür, S. Misra, and N. Kuranova, who helped me to put my ideas into English. I sincerely appreciate the critical reading of the manuscript and many helpful comments and suggestions. Anonymous reviewers are thanked for critical reading of the manuscript and many helpful comments and suggestions.

I would also like to thank M. Cronwright for providing many helpful comments and suggestions. I am deeply grateful to I. Kravchenko-Berezhnoy and V. N. Endo for providing the data from mineralized pegmatites and fertile granites worldwide would be very much appreciated.

References

Abdalla, H. M., Helba, H. A. & Mohamed, F. H. 1998. Chemistry of columbite-tantalite minerals in rare metal granitoids, Eastern Desert, Egypt. Mineralogical Magazine, 62, 821–836.

Alexeev, N. L., Balagansky, V. V. et al. 2003. Rates of Early Proterozoic orogenic processes: a study of U–Pb and Sm–Nd zircon and garnet systems and metamorphic processes in rocks of the Pon’gom-Navolok Island, Central Belomorian Region. In: Kozakov, I. K. & Kotoy, A. B. (eds) Isotope Geochronology for Resolving Problems of Geodynamics and Ore Genesis. Centre for Information Culture Publishers, St. Petersburg, 60–63 [in Russian].

Armstrong, T. R. & Tracy, R. J. 2000. One-dimensional thermal modelling of Acadian metamorphism in southern Vermont, USA. Journal of Metamorphic Geology, 18, 625–638.

Ayers, C. J., Dunkle, S., Gao, S. & Miller, C. E. 2002. Constraints on timing of peak and retrograde metamorphism in the Dabie Shan ultrahigh-pressure metamorphic belt, east-central China, using U–Th–Pb dating of zircon and monazite. Chemical Geology, 186, 315–331.

Balashov, Y. A. & Glaznev, V. N. 2006. Endogenic cycles and the problem of crustal growth. Geochemistry International, 44, 131–140.

Bekskin, S. M. & Marin, Yu. B. 2003. About evolution of the rare-metal granite mineral- and ore-forming process during the geological history. Zapiski Vserossiyskogo Mineralogicheskogo Obschestva (Proceedings of the Russian Mineralogical Society), 132, 1–14 [in Russian].

Bethune, K. M., Villeneuve, M. E. & Bleeker, W. 1999. Laser At/Ar and Ar thermochronology of Archaean rocks in Yellowknife domain, southwestern Slave province: Insights into the cooling history of an Archaean granite–greenstone terrane. Canadian Journal of Earth Sciences, 36, 1189–1206.

Briggs, S. M., Yin, A., Manning, C. E., Chen, Z. L., Wang, X. F. & Grove, M. 2007. Late Paleozoic tectonic history of the Ertix Fault in the Chinese Altai and its implications for the development of the Central Asian Orogenic System. Geological Society of America Bulletin, 119, 944–960.

Broughton, P. L. 1973. Precious topaz deposits of the Llano Uplift area, central Texas. Rocks & Minerals, 48, 147–156.

Brown, M. 2001. Orogeny, magmatites and leucogranites: a review. Proceedings of Indian Academy of Sciences, Earth Planetary Sciences, 110, 313–336.

Brown, M., Friend, C. R. L., McGregor, V. R. & Perkins, W. T. 1981. The late-Archaean Qorqut granite complex of southern western Greenland. Journal of Geophysical Research, 86, 10617–10632.

Burg, J. P., Davy, P. et al. 1997. Exhumation during crustal folding in the Namche-Barwa syntaxis. Terra Nova, 9, 53–56.

Černý, P. 1991a. Rare-element granite pegmatites: Part I. Anatomy and internal evolution of pegmatite deposits. Geoscience Canada, 18, 49–67.

Černý, P. 1991b. Rare-element granite pegmatites: Part II. Regional to global environments and petrogenesis. Geoscience Canada, 18, 68–81.

Černý, P. 1991c. Fertile granites of Precambrian rare-element pegmatite fields: is geochemistry controlled by tectonic setting or source lithologies? Precambrian Research, 51, 429–468.

Černý, P. 2005. The Tanco rare-element pegmatite deposit, Manitoba: regional context, internal anatomy, and global comparisons. In; Linnien, R. L. & Samson, I. M. (eds) Rare-element Geochemistry and Mineral Deposits. Geological Association of Canada, Short Course Notes, 17, 127–158.

Černý, P. & Ercit, T. S. 2005. Classification of granitic pegmatites. Canadian Mineralogist, 43, 2005–2026.

Collins, W. J. & Van Kranendonk, M. J. 1999. Model for the development of kyanite during partial convective overturn of Archaean granite-greenstone terranes: the Pilbara Craton, Australia. Journal of Metamorphic Geology, 17, 145–156.

Condie, K. C. 1998. Episodic continental growth and supercontinents: a mantle avalanche connection? Earth and Planetary Science Letters, 163, 97–108.

Condie, K. C. 2001. Continental growth during formation of Rodinia at 1.35–0.9 Ga. Gondwana Research, 4, 5–16.

Condie, K. C. 2002. The supercontinent cycle: are there two patterns of cyclicity? Journal of African Earth Sciences, 35, 179–183.
EVOLUTION OF GRANITIC PEGMATITES METALLOGENY

Cosca, M. A., Mezger, K. & Essene, E. J. 1998. The Baltica-Laurentia connection: Sveconorwegian (Grenvillian) metamorphism, cooling, and unroofing in the Bambel Sector, Norway. *Journal of Geology*, 106, 539–552.

Cox, R. A., Indares, A. & Dunning, G. R. 2002. Temperature–time paths in the high-P Manicouagan Imbricate zone, eastern Grenville Province: evidence for two metamorphic events. *Precambrian Research*, 117, 225–250.

Degraaff-Surpless, K., Graham, S. A., Wooden, J. L. & McWilliams, M. O. 2002. Detrital zircon provenance analysis of the Great Valley Group, California: Evolution of an arc-forearc system. *Geological Society of America Bulletin*, 114, 1564–1580.

De Ronde, C. E. J. & De Wit, M. J. 1994. Tectonic history of the Barberton Greenstone Belt, South Africa: 490 million years of Archaean evolution. *Tectonics*, 13, 983–1005.

Dufour, M. S., Popova, V. A. & Krivets, T. N. 1970. *Alpine Metamorphic Complex of the Eastern Central Pamirs*. LGU Publishing House, Leningrad [in Russian].

Ehlmann, A. J., Walper, J. L. & Williams, J. 1964. A new, Barringer Hill-type, rare-earth pegmatite from the Central Mineral Region, Texas. *Economic Geology*, 59, 1348–1360.

Faure, G. & Mensing, T. M. 2005. *Isotopes: Principles and Applications*. John Wiley & Sons, New Jersey.

Fersman, A. E. 1931. *Pegmatites: Their Scientific and Practical Importance*. V.I. Granitic Pegmatites. USSR Academy of Sciences Publishing House, Leningrad [in Russian].

Fersman, A. E. 1940. *Pegmatites. V.I. Granitic Pegmatites* (3rd edition: corrected and supplemented). USSR Academy of Sciences Publishing House, Moscow-Leningrad [in Russian].

Flowers, R. M., Mahan, K. H., Bowring, S. A., Williams, L. M., Pringle, M. S. & Hodges, K. V. 2006. Multistage exhumation and juxtaposition of lower continental crust in the western Canadian Shield: Linking high-resolution U–Pb and 40Ar/39Ar thermochronometry with pressure-temperature-deformation paths. *Tectonics*, 25, TC4003, doi: 10.1029/2005TC001912.

Fontan, D., Schouppe, M. A., Hunziker, J., Martinotti, G. & Verkaerken, J. 2000. Metamorphic evolution, 40Ar–39Ar chronology and tectonic model for the Neelum valley, Azad Kashmir, NL Pakistan. *In: Khan, M. A., Treloar, P. J., Searle, M. P. & Jan, M. Q. (eds) Tectonics of the Nanga Parbat Synaxis and of the Western Himalaya*. Geological Society, London, Special Publications, 170, 431–453.

Fraser, G., McDougall, I., Ellis, D. J. & Williams, I. S. 2000. Timing and rate of isothermal decompression in Pan-African granulites from Rundvågshetta, East Antarctica. *Journal of Metamorphic Geology*, 18, 441–454.

Friend, C. R. L. & Nutman, A. P. 2005. New pieces to the Archaean jigsaw puzzle in the Nuuk region, southern West Greenland: steps in transforming a simple insight into a complex regional tectonothermal model. *Journal of the Geological Society*, 162, 147–162.

Ginzburg, A. I. & Rodionov, G. G. 1960. On the depth of formation of granitic pegmatites. *Geologia Rudnykh Mestorozhdeniy (Geology of Ore Deposits)*, 1, 45–54 [in Russian].

Ginzburg, A. I., Timofeyev, I. N. & Feldman, L. G. 1979. *Principles of Geology of the Granitic Pegmatites*. Nedra, Moscow [in Russian].

Haapala, I. 1995. Metallogeny of the rapakivi granites. *Mineralogy and Petrology*, 54, 141–160.

Hacker, B. R., Yin, A., Christie, J. M. & Snoke, A. W. 1990. Differential stress, strain rate, and temperatures of mylonitization in the Ruby Mountains, Nevada: Implications for the rate and duration of uplift. *Journal of Geophysical Research*, 95, 8569–8580.

Hames, W. E., Tracy, R. J. & Bodnar, R. J. 1989. Postmetamorphic unroofing history deduced from petrology, fluid inclusions, thermochronometry, and thermal modeling: an example from southwestern New England. *Geology*, 17, 727–730.

Hamner, S., Hamilton, M. A. & Crowley, J. L. 2002. Geochronological constraints on Paleoarchean thrust nappe and Neoarchean accretionary tectonics in southern West Greenland. *Tectonophysics*, 350, 255–271.

Harris, P. D., Robb, L. J. & Tomkinson, M. J. 1995. The nature and structural setting of rare-element pegmatites along the northern flank of the Barberton greenstone belt, South Africa. *South Africa Journal of Geology*, 98, 82–94.

Hessler, A. M. & Lowe, D. R. 2006. Weathering and sediment generation in the Archaean: An integrated study of the evolution of siliciclastic sedimentary rocks of the 3.2 Ga Moodies Group, Barberton Greenstone Belt, South Africa. *Precambrian Research*, 151, 185–210.

Hodges, K. V. 2003. Geochronology and thermochronology in orogenic systems. *In: Rudnick, R. L. (ed.) Treatise on Geochemistry*, The Crust, 3, Elsevier, New York, 263–292.

Hofmann, A. 2005. The geochemistry of sedimentary rocks from the Fig Tree Group, Barberton greenstone belt: Implications for tectonic, hydrothermal and surface processes during mid-Archaean times. *Precambrian Research*, 143, 23–49.

Holm, D. K., Snow, J. K. & Lux, D. R. 1992. Thermal and barometric constraints on the intrusive and unroofing history of the Black Mountains: Implications for timing, initial dip, and kinematics of detachment faulting in the Death Valley region, California. *Tectonics*, 11, 507–522.

Holtz, F., Behrens, H., Dingwell, D. B. & Johannes, W. 1995. Water solubility in haplogranitic melts: Compositional, pressure and temperature dependence. *American Mineralogist*, 80, 94–108.

House, M. A., Bowring, S. A. & Hodges, K. V. 2002. Implications of middle Eocene epizonal plutonism for the unroofing history of the Bitterroot metamorphic core complex, Idaho-Montana. *Geological Society of America Bulletin*, 114, 448–461.

Huang, W. L. & Wylie, P. J. 1981. Phase relations of S-type granite with H2O to 35 kbar: muscovite granite from Harney Peak, South Dakota. *Journal of Geophysical Research*, 86, 10515–10529.
Martignole, M., Nediari, S. 2002. Contrastive evolution of low-P rare metal granites from two different terranes in the Hoggar area, Algeria. Journal African Earth Sciences, 34, 247–257.

Kotrová, J., Gerebs, A., Párhí, R. R. & Novák, M. 2007. Clasts of Variscan high-grade rocks within Upper Viséan conglomerates – constraints on exhumation history from petrology and U–Pb chronology. Journal of Metamorphic Geology, 25, 781–801.

Kratz, K. O. (ed.) 1984. Principles of the Metallogeography of the Precambrian Metamorphic Belts. Nauka, Leningrad [in Russian].

Kudrin, V. S., Stavrov, O. D. & Shuriga, T. N. 1994. New spodumene type of tantallum-bearing rare metal granites. Petrologia, 2, 88–95 [in Russian].

Landes, K. K. 1932. The Baringer Hill pegmatite. American Mineralogist, 17, 381–390.

Landes, K. K. 1935. Age and distribution of pegmatites. American Mineralogist, 20, 81–105, 153–175.

Lazarenko, E. K., Pavlishin, V. I., Latysh, V. T. & Sokkin, Y. G., 1973. Mineralogy and Genesis of the Chamber Pegmatites of Volynia. Vischa Shkola, Lvov [in Russian].

Lindh, A. 2005. Origin of chemically distinct granites in a composite intrusion in east-central Sweden: geochemical and geothermal constraints. Lithos, 80, 249–266.

London, D. 2005. Granitic pegmatites: an assessment of current concepts and directions for the future. Lithos, 80, 281–303.

London, D. 2008. Pegmatites. The Canadian Mineralogist, Quebec. Special Publication, 10.

Luth, W. C. 1976. Granitic rocks. In: Bailey, D. K. & MacDonald, R. (eds) The Evolution of Crystalline Rocks. Academic Press, London, 335–417.

Luth, W. C., Jahn, R. H. & Tuttle, O. F. 1964. The granite system at pressures of 4 to 10 kilobars. Journal of Geophysical Research, 69, 759–773.

Makrygina, V. A., Makagon, V. M., Zagorsky, V. E. & Smakin, B. M. 1990. Granitic Pegmatites. V.1: Mica-bearing Pegmatites. ‘Nauka’, Novosibirsk [in Russian].

Maphalala, R. M. & Trumbull, R. B. 1998. A geochemical and Rb/Sr isotopic study of Archaean pegmatite dykes in the Tin Belt of Swaziland. South African Journal of Geology, 101, 53–65.

Maphalala, R. M., Kröner, A. & Kramers, J. D. 1989. Rb–Sr ages for Archaean granitoids and tin-bearing pegmatites in Swaziland, southern Africa. Journal of African Earth Sciences, 9, 749–757.

Mareschal, J.-C. 1994. Thermal regime and post-oreogenic extension in collision belts. Tectonophysics, 238, 471–484.

Martignole, J. & Reynolds, P. 1997. 40Ar/39Ar thermochronology along a western Quebec transect of the Grenville Province, Canada. Journal of Metamorphic Geology, 15, 283–296.

Martin, Y. J. 1964. The Bikita tinfield. Southern Rhodesia Geological Survey Bulletin, 58, 114–143.

Martínez, F. J., Julivert, M., Sebastian, A., Arboleya, M. L. & Gil Ibaguchi, J. I. 1988. Structural and thermal evolution of high-grade areas in the northeast-western parts of the Iberian Massif. American Journal of Science, 288, 969–996.

Moorbath, S., Taylor, P. N. & Goodwin, R. 1981. Origin of granitic magma by crustal remobilisation: Rb–Sr and Pb/Pb geochronology and isotope geochemistry of the late Archaean Qoqut Granite Complex of southern West Greenland. Geochimica et Cosmochimica Acta, 45, 1051–1060.

Morteani, G., Preinfalk, A. & Horn, A. H. 2000. Classification and mineralization potential of the pegmatites of the Eastern Brazilian Pegmatite Province. Mineralium Deposita, 35, 638–655.

Mosher, S., Levine, J. S. F. & Carlson, W. D. 2008. Mesoproterozoic plate tectonics: a collisional model for the Grenville-aged orogenic belt in the Llano uplift, central Texas. Geology, 36, 55–58.

Nutman, A., Allaart, J., Bridgwater, D., Dimroth, E. & Rosing, M. 1984. Stratigraphic and geochemical evidence for the depositional environment of the Early Archaean Isa Supracrustal belt, southern West Greenland. Precambrian Research, 25, 365–396.

Nutman, A. P., Bennett, V. C., Friend, C. R. L. & McGregor, V. R. 2000. The early Archaean Isaa Gneiss Complex of southern West Greenland: the importance of field observations in interpreting age and isotopic constraints for early terrestrial evolution. Geochimica et Cosmochimica Acta, 64, 3035–3060.

Nutman, A. P., Friend, C. R. L. & Bennett, V. C. 2001. Review of the oldest (4400–3600 Ma) geological record: glimpses of the beginning. Episodes, 24, 93–101.

Nutman, A. P., Friend, C. R. L. & Bennett, V. C. 2002. Evidence for 3650–3600 Ma assembly of the northern end of the Isaak Gneiss Complex, Greenland: implications for early Archaean tectonics. Tectonics, 21, 10.1029/2000TC001203.

O’Reilly, S. Y. & Griffin, W. L. 1996. 4-D lithospheric mapping: a review of the methodology with examples. Tectonophysics, 262, 3–18.

Ovcinnikov, L. N., Voronovskiy, S. N. & Ovchinnikova, L. B. 1975. Radiogeochronologic of granitic pegmatites. Doklady of the USSR Academy of Sciences, 223, 1202–1205 [in Russian].

Ovcinnikov, L. N., Voronovskiy, S. N. & Ovchinnikova, L. B. 1976. Radiogeochronologic of granitic pegmatites. In: Studies on Geological Petrology. Nauka, Moscow, 319–326 [in Russian].

Partington, G. A., McNaughton, N. J. & Williams, I. S. 1995. A review of the geology, mineralization, and geochronology of the Greembushes pegmatite, Western Australia. Economic Geology, 90, 616–635.

Pedroso-Saores, A. C., De Campos, C. P. et al. 2011. Late Neoproterozoic–Cambrian granitic magmatism in the Araçuaí orogen (Brazil), the Eastern Brazilian Pegmatite Province and related mineral resources. In: Sial, A. N., Bettencourt, J. S., De Campos, C. P. & Ferreira, V. P. (eds) Granite-Related Ore Deposits. Geological Society, London, Special Publications, 350, 25–51.

Pericival, J. A. 1979. Kyanite-bearing rocks from the Hackett River area, N.W.T.: Implications for Archaean geothermal gradients. Contributions to Mineralogy and Petrology, 69, 177–184.
EVOLUTION OF GRANITIC PEGMATITIC METALLOGENY

Poudjom Djomani, Y. H., O’Reilly, S. Y., Griffin, W. L. & Morgan, P. 2001. The density structure of subcontinental lithosphere through time. Earth and Planetary Science Letters, 184, 605–621.

Rossoskii, L. N., Chmyrev, V. M. & Salakh, A. S. 1976. Genetic relationship of apatitic spodumene dikes to lithium-pegmatite veins. Doklady of the USSR Academy of Sciences, Earth Science Section, 226, 170–172.

Rougvie, J. R., Carlson, W. D., Copeland, P. & Connelly, J. N. 1999. Late thermal evolution of Proterozoic rocks in the northeastern Llano Uplift, central Texas. Precambrian Research, 94, 49–72.

Ruppel, C. & Hodges, K. V. 1994. Pressure-temperature-time paths from two-dimensional thermal models: prograde, retrograde, and inverted metamorphism. Tectonics, 13, 17–44.

Sal’ye, V. E. & Glebovitsky, V. A. 1976. Metallogenic Specialisation of Pegmatites on the East of the Baltic Shield. Nauka, Leningrad [in Russian].

Scaillet, S., Cuney, M., Le Carlier de Veslud, C., Cheilletz, A. & Royer, J. J. 1996. Cooling patterns and mineralization history of the Saint Sylvestre and Western Marche leucogranite plutons, French Massif Central. II. Thermal modelling and implications for the mechanisms of U-mineralization. Geochimica et Cosmochimica Acta, 60, 4673–4688.

Schneiderhön, H. 1961. Die Erzlagerstätten der Erde. Bd. 2. Die Pegmatite. Gustav Fisher Verlag, Stuttgart.

Schoene, B., De Wit, M. J. & Bowring, S. A. 2008. Mesochronaean assembly and stabilization of the eastern Kaapvaal craton: A structural-thermochronological perspective. Tectonics, 27, TC5010, doi: 10.1029/2008TC002267, 1–27.

Seacher, K., Steenfelt, A. & Garde, A. A. 2008. Pegmatites and their potential for mineral exploration in Greenland. Geology and Ore, 10, 2–12.

Searle, M. P., Waters, D. J., Rex, D. C. & Wilson, R. N. 1992. Pressure, temperature and time constraints on Himalayan metamorphism from eastern Kashmir and western Zanskar. Journal of the Geological Society, 149, 753–773.

Shmakin, B. M., Zagorsky, V. E. & Makagon, V. M. 2007. Granitic Pegmatites, v.4: Rare-earth Pegmatites. Pegmatites of Unusual Composition. ‘Nauka’, Novosibirsk [in Russian].

Shroder, J. F., Jr. & Bishop, M. P. 2000. Unroofing of the Nanga Parbat Himalaya. In: Khan, M. A., Treloar, P. J., Searle, M. P. & Jan, M. Q. (eds) Tectonics of the Nanga Parbat Synaxis and the Western Himalaya. Geological Society, London, Special Publications, 170, 163–179.

Sinclair, H. D., Gibson, M., Naylor, M. & Morris, R. G. 2005. Asymmetric growth of the Pyrenees revealed through measurement and modeling of orogenic fluxes. American Journal of Science, 305, 369–406.

Söderlund, U., Hellström, F. A. & Kamo, S. L. 2008. Geochronology of high-pressure mafic granulite dykes in SW Sweden: tracking the P–T–t path of metamorphism using Hf isotopes in zircon and baddeleyite. Journal of Metamorphic Geology, 26, 539–560.

Solodov, N. A. 1985. Metallogeny of Rare-metal Formations. Nedra, Moscow [in Russian].

Stillings, A., Černý, P. & Vanstone, P. J. 2006. The Tanco pegmatite at Bernie Lake, Manitoba. XVI. Zonal and bulk compositions and their petrogenetic significance. Canadian Mineralogist, 44, 599–623.

Sweetapple, M. T. & Collins, P. L. F. 2002. Genetic framework for the classification and distribution of Archaean rare metal pegmatites in the North Pilbara craton, Western Australia. Economic Geology, 97, 873–895.

Taylor, S. R. & McLennan, S. M. 1985. The Continental Crust: Its Composition and Evolution. Blackwell, Oxford.

Thompson, A. B. 1999. Some time–space relationships for crustal melting and granitic intrusion at various depths. In: Castro, A., Fernandez, C. & Vigneresse, J. L. (eds) Understanding Granites: Integrating New and Classical Techniques. Geological Society, London, Special Publications, 168, 7–25.

Trachev, A. V. & Gershenskoy, A. Sh. 1997. Mineral Raw Materials. Mica. Handbook. ZAO ‘Geoinformmark’, Moscow [in Russian].

Trachev, A. V., Sapozhnikova, L. N., Zhukova, I. A. & Zhukov, N. A. 1998. Location and generation conditions of the sheet muscovite deposits with large reserves and high quality of raw materials. Otechestvennaya Geologiya (Domestic Geology), 4, 35–39 [in Russian].

Trumbull, R. B. 1993. A petrological and Rb/Sr isotopic study of an early Archaean fertile granite-pegmatite system: the Sinceni Pluton in Swaziland. Precambrian Research, 61, 89–116.

Trumbull, R. B. 1995. Tin mineralization in the Archaean Sinceni rare element pegmatite field, Kaapvaal Craton, Swaziland. Economic Geology, 90, 648–657.

Trumbull, R. B. & Chausson, M. 1998. Chemical and boron isotopic composition of magmatic and hydrothermal tourmalines from the Sinseni granite-pegmatite system in Swaziland. Chemical Geology, 153, 125–137.

Vermeech, P., Miller, D. D., Graham, S. A., De Grave, J. & McWilliams, M. O. 2006. Multiphase detrital thermochronology of the Great Valley Group near New Idria, California. Geological Society of America Bulletin, 118, 210–218.

Zagorsky, V. E., Makagon, V. M., Shmakin, B. M., Makrygina, V. A. & Kuznetzova, L. G. 1997. Granitic Pegmatites, v.2: Rare-metal Pegmatites. ‘Nauka’, Novosibirsk [in Russian].

Zagorsky, V. E., Peretyazhko, I. S. & Shmakin, B. M. 1999. Granitic Pegmatites, v.3: Miarolitic Pegmatites. ‘Nauka’, Novosibirsk [in Russian].

Zagorsky, V. E., Makagon, V. M. & Shmakin, B. M. 2003. Systematics of granitic pegmatites. Russian Geology and Geophysics, 44, 422–435.

Zeh, A., Klemd, R., Buhlmann, S. & Barton, J. M. 2004. Pro- and retrograde P–T evolution of granulites of the Beit Bridge Complex (Limpopo Belt, South Africa): constraints from quantitative phase diagrams and geotectonic implications. Journal of Metamorphic Geology, 22, 79–95.

Zulauf, G., Dorr, W., Flitra, J., Kotkova, J., Maluski, H. & Valverde-Vaquero, P. 2002. Evidence for high-temperature diffusional creep preserved by rapid cooling of lower crust (North Bohemian shear zone, Czech Republic). Terra Nova, 14, 343–354.