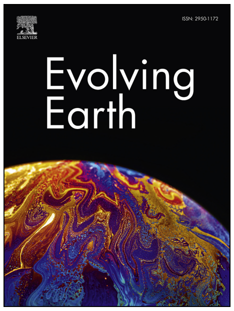

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Genesis, Classification, Tectonic Setting and Economic Potential of Global Granitic Pegmatites: A Review

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Declaration of interests

☒ The authors declare that they have no known competing financial or personal interests

Abstract

Granitic pegmatites have attracted a significant interest of petrologists and mineralogists due to their complex genetic characteristics and substantial economic value. Although numerous reviews addressed specific aspects such as genesis, classification, economic potential, and tectonic setting, a comprehensive review integrating all these facets in a single article has been lacking. This study fills that gap by offering a holistic review that synthesizes these key dimensions. It examines prevailing theories on pegmatite genesis, traces the evolution of classification systems, and assesses the economic importance of pegmatites as sources of industrial minerals, strategic metals, and gemstones, as well as their association with different tectonic settings.

The crystallization of the majority of granitic pegmatites is widely acknowledged to occur in the presence of a viscous, volatile-rich polymerized silicate magma, concomitant with supercritical aqueous fluids. Consequently, granitic pegmatites can frequently be conceded as a hybrid rock derived through igneous and hydrothermal processes. Various criteria, such as the emplacement depth of granitic pegmatites, their association with metamorphism and granitic plutons, geochemical signatures, and their relationships with host rocks, have traditionally been employed for the classification of granitic pegmatites. However, none of the classifications has been accepted unequivocally.

Pegmatitic melt is often rich in chemical constituents (including volatiles) that are incompatible in common rock-forming minerals. In addition to rare earth elements, pegmatites are commonly enriched in rare elements such as Li, B, Be, Cs, and Ta. Because they are incompatible in common rock-forming minerals, these elements tend to create specific mineral phases that can host themselves. As a result, the concentration of trace elements such as Li, B, Cs, Ta, and Bi in these bodies are enriched to levels as high as a thousand times higher than the typical crustal abundance.

Granitic pegmatites encompass a diverse array of economic commodities, serving as repositories for industrial minerals and sources for strategic metals like Ta, Cs, Sn, Nb, Be, Sb, W, Co, and rare earth elements. Furthermore, they contribute as reservoirs of fluxing components, including Li, P, F, and B, while serving as sources of precious and semi-precious gemstones and some radioactive minerals. The global distribution of pegmatite ages suggests a linkage between the age of pegmatite formation and supercontinent assembly. The Li, Ce and Ta enriched (LCT) granitic pegmatites form part of the orogenic suites associated with crustal shortening in zones of subduction and instances of continental collision. The Nb, Y, and F-enriched (NYF) pegmatites are mainly interpreted as products of anorogenic suites and formed mainly in extensional settings.

Keywords: Granitic pegmatites, pegmatite classification, industrial minerals, strategic metals, coloured gemstones

1 Introduction

Pegmatites are igneous rocks characterized by a granitic composition and are distinguished from other igneous formations by their exceptionally coarse yet variable grain size or by an abundance of crystals exhibiting skeletal, graphic, or other strongly directional growth habits (London 2008). While there is a prevailing consensus among petrologists regarding the spectrum from magmatic to magmatic-hydrothermal processes as the likely genesis of pegmatites (e.g., Hunt, 1871; Landes, 1937; Ramberg, 1952; Reitan, 1965; Roedder, 1981; Chiarenzelli et al., 2019; Xu et al., 2024), the complete understanding of their origin remains elusive. The historical definitions of pegmatites, as outlined in Table 1, reveal a progression from simple descriptive terms based on grain size and mineral composition to more complex and nuanced understandings that consider textural variability and growth habits.

Pegmatite occurrences, typically ranging from a few cm- to m-scale, can extend impressively to several Km^2 (e.g., London and Kontak, 2012; Dharmapriya et al., 2021; Li et al., 2023). These formations are ubiquitously distributed across continents, found both within ancient cratons and younger, marginal mountain belts (London, 2018; Jain et al., 2020; Wang et al., 2023). Composed of interlocking large crystals of varying sizes, pegmatites commonly host crystals exceeding 2.5 cm, with remarkable instances of crystals surpassing 10 m in length reported (Kozłowski, 2007; London and Kontak, 2012; William et al., 2024). Occasionally, these crystals exhibit systematic dispersion from the margins to the centres of the bodies, showcasing distinctive mineralogical zonation (London, 2014). Pegmatites are primarily granitic in composition, with quartz, feldspar, and mica as their main mineral constituents, making them a significant source of these minerals (Dill, 2015; London, 2015).

Granitic pegmatites have been a focal point of research among petrologists and mineralogists due to their profound economic significance. These formations serve as storehouses for essential industrial minerals, including quartz, feldspars and micas (Roedder, 1992; Glover et al., 2012; Kesler, 2012; London et al., 2012b; Dill, 2015; Cardoso-Fernandes et al., 2023). Furthermore, certain pegmatites exhibit rare earth element concentrations that are thousand times higher than crustal abundance, establishing granitic pegmatites as a noteworthy source for strategic metal-bearing minerals (Cerný, 1991; Simmons and Webber, 2008; Linnen et al., 2012; Bradley et al., 2017; Feng et al., 2023). Pegmatites also host precious and semiprecious gemstones (Simmons, 2007a, 2012b; McManus et al., 2008; Beurlen et al., 2013; Hussain et al., 2021; Mauthner, 2024). Additionally, pegmatites may contain fluxing components and energy minerals, further augmenting their geological importance (London and Kontak, 2012; Dill, 2015; London, 2018). Distinct from typical igneous rocks, pegmatites deviate due to their enigmatic origin as well as their unique mineralogy and textures.

The genesis of pegmatites has undergone investigation for over a century, with various proposed mechanisms. However, achieving a universal consensus on the formation of pegmatites remains elusive (e.g., Niggli, 1920; Merritt, 1924; Schaller, 1925; Landes, 1933; Cameron et al., 1949; Jahns and Burnham, 1969; Simmons and Webber, 2008; Simmons et al., 2024). Meanwhile, over the last six decades, the diverse spectrum encompassing mineralogical, textural, geochemical, and economic characteristics, along with their relation to the host rock of granitic pegmatites, has been the subject of numerous attempts at classification (Ginzburg and Rodionov, 1960; Ginsburg et al., 1979; Černý, 1991; Zargosky et al., 1999; Wise, 1999; Pezzotta, 2001; Černý and Erict, 2005; Martin and De Vito, 2005; Fuchsloch et al., 2017; Beskin et al., 2018).

Numerous previous work on granitic pegmatites have been published, examining (a) their genesis, elucidating the characteristics of the source magma, such as geochemical constituents, cooling rates, and the formation of coarse-grained texture, as well as the pressure-temperature (*P-T*) conditions of crystallization (e.g. London, 2005; Simmons and Webber, 2008; London et al., 2012; Černý et al., 2012), including discussions on heterogeneity and zonations (London, 2014); (b) classification schemes (Černý and Ercit, 2005; Simmons, 2005); (c) their economic potential as sources of industrial minerals (Glover et al., 2012), strategic metals (e.g., Linnen et al., 2012), and precious gemstones (e.g., Shigley and Kampf, 1984; Simmons, 2012); and (d) the tectonic settings in which pegmatites are formed (Martin and De Vito, 2005). Nevertheless, these review articles have often focused singularly on specific aspects, and a comprehensive review covering all these dimensions is notably absent.

In this paper, we aim to provide a comprehensive historical overview, compiling knowledge on granitic pegmatites from early studies to the latest research. We place particular emphasis on (a) mineralogy and anatomy, exploring the genesis of pegmatitic melt, its chemical constituents, and crystallization conditions; (b) attempts at classifying granitic pegmatites; and (c) their economic potential, as well as (d) the tectonic settings in which pegmatites are formed. These multifaceted aspects have not been collectively reviewed by previous researchers, contributing to a more holistic understanding of granitic pegmatites.

2. Mineralogical and Anatomical Features of Granitic Pegmatites

The granitic pegmatitic composition shares a simple mineral assemblage with granite, featuring dominant phases such as quartz, albite, and potassium feldspar—either orthoclase or microcline. Martin (1988) showed how the Al-Si ordering in K-feldspar does not follow a unique path and the K-feldspar, sanidine, orthoclase, and microcline are pseudomorphs of the magmatic phases. They have emphasized the discontinuous and continuous transformation of these phases during pegmatite formation. Muscovite and/or biotite 'books' commonly occur as minor mineral phases in many granitic pegmatites (London and Kontak, 2012; Ndikumana et al., 2020). In addition, granitic pegmatites may bear over fifty accessory mineral species such as iron oxides, zircon, fluorite, apatite, lepidolite, amblygonite, spodumene, tourmaline with variable colours, beryl, amber-coloured topaz, and rose and amethystine quartz (Fernald, 1906; Bastin, 1910; Abdelfadil et al., 2016; Chakraborty et al., 2024). Table 2 summarizes the occurrences of accessory minerals in some selected pegmatites in the globe. Minerals like sericite, epidote, and chlorite often occur as secondary phases (Abdelfadil et al., 2016; Georgieva et al., 2022).

The mineralogical composition as well as the elemental concentration in pegmatite is influenced by the distance of the pegmatite from the source (e.g., Simmons and Webber, 2008; Fredriksso, 2017). For instance, the Na/K ratio in feldspars correlates well with the source rock (Stugard, 1958; Tang et al., 2018). K-rich pegmatites are found closest to the source, while Na-rich pegmatites are situated at the most distal end of the pegmatite source (Černý et al., 2012). The K/Rb ratio decreases with the advancing crystallization of K-feldspar (Černý et al., 1985). These fractionation patterns play a significant role in determining the exotic mineralogy of the pegmatite body (Trueman and Černý, 1982; Breaks et al., 2005). Post-crystallization processes, for example, the subsolidus reaction of solids with aqueous fluids (primary or secondary) facilitate the formation of new mineral phases in granitic pegmatites (e.g., Černý, 1972). A detailed summary of the formation of industrial minerals, strategic metals, gemstones, and energy minerals is provided in Section 7.

Pegmatites can exhibit zonation or remain unzoned based on mineralogy and rock fabrics. Two types of zoned pegmatites can be recognized based on their scale: (i) regional zonation, where chemical complexity increases with distance from granitic or other thermal sources, and (ii) internal zonation (Fig. 1), representing mineralogical and textural changes within individual pegmatite bodies (London, 2014). Cameron et al. (1949) proposed the nomenclature and defined the patterns of internal zonation within granitic pegmatites that are in use today. The four main zones defined by these authors are (i) border (ii) wall (iii) intermediate and (iv) core (Fig. 1). As summarized by London (2014): (i) The border zone is a thin layer (mm to cm thick) that surrounds the pegmatite body in contact with its host rocks. The grain size is fine-grained (~2–5 mm), and the texture is hypidiomorphic granular. (ii) The wall zone is a thick margin surrounding the pegmatite bodies, containing plagioclase and predominantly quartz, along with minerals such as muscovite, biotite, K-feldspar, garnet, tourmaline, apatite, primary chlorite, beryl, and columbite as accessory minerals (London, 2014; Pattison and De Buhr, 2015). (iii) The intermediate zone is symmetrically or asymmetrically distributed through the pegmatite, carrying exceedingly coarse-grained and blocky textures. These textures result from the buildup of fluxing components in a boundary layer of liquid that advances into the pegmatite along the crystal growth front (Kesler et al., 2012). The zone contains perthitic microcline, plagioclase, muscovite, spodumene, and petalite with quartz. (iv) The core of a pegmatite constitutes its innermost unit within the zoned structure, characterized by a singular mass or multiple repetitions of the same mineralogy at the same structural or sequential position. In granitic pegmatites, monominerallic quartz cores are frequently encountered (London, 2014). Nevertheless, in certain pegmatites (Li-rich pegmatite), in addition to quartz, the latest primary units may occasionally exhibit a quartz-poor composition, featuring albite and lepidolite, along with various typically rare minerals. The proportions of core material to the other zones can be exceedingly variable, and monomineralic quartz cores predominate in granitic pegmatites (Brögger, 1890).

Brögger (1890) proposed zonation as the simultaneous interaction of silicate melt and aqueous fluid, a concept subsequently widely described by Jahns and Burnham (1969). This type of pegmatite is concentrically zoned and graded inward from zones rich in feldspar, through intermediate zones containing rare element minerals, toward a core strongly enriched in quartz (Cameron et al., 1949; Jahns, 1955; Kesler et al., 2012). Grain size increases drastically inwards due to more rapid diffusion through the low-viscous, flux-enriched liquid (Bartels et al., 2011). Unzoned pegmatites tend to occur in association with host rocks of high metamorphic grade in the stability field of kyanite and spodumene (Horton et al., 1987). This type of pegmatite is relatively narrow and long (1 m to 50 m wide and up to 1 km in length; Lentz, 1992), and the distribution of grain sizes and mineralogy is largely homogeneous (Kesler, 1961; Swanson, 2012; London, 2014).

3. Genesis of Granitic Pegmatitic melts, and their syn-crystallization modification, and post-crystallization evolution

3.1 Unraveling the Genesis of Granitic Pegmatitic Melts

The complexity of pegmatite genesis defies a universally accepted model, as its diverse features resist a singular explanation (Simmons and Webber, 2008). The pegmatite-forming process constitutes a continuous physiochemical transition, commencing from magma, traversing through granitic magma cooling, and extending into hydrothermal conditions (Thomas and Davidson, 2007). Previous studies, employing mineral chemistry, fluid inclusions, and experimental phase equilibria, introduced varied models reliant on bulk compositions, depths, and cooling histories (London and Kontak, 2012).

Since the late 19th century, efforts to elucidate the formation processes of intricate textures and uncommon minerals in massive granite-derived pegmatites have persisted. Brögger (1890) proposed internal reactions between a silicate melt and water vapour in the magmatic chamber as the genesis of most pegmatites. A model proposed by Cameron et al. (1949) attributed pegmatite chemical evolution to fractional crystallization. Later, two concepts emerged to explain granitic pegmatite formation (London, 2008). According to experimental findings, pegmatites form via equilibrium crystallization of coexisting granitic melt and hydrous fluid at or slightly below the hydrous granite liquidus (Jahns and Burnham, 1969; Jahns and Tuttle, 1963). Richard and Jahns (1970, 1980) suggested pegmatites evolve from residual granitic melts comprised of coexisting water vapour and silicate melt. Currently, three hypotheses explain pegmatite origin as outlined below.

3.1.1 Fractional Crystallization of a Granitic Magma at the Pluton Scale

The most widely accepted model of pegmatite formation is by fractional crystallization of a granitic pluton (Cameron et al. 1949; Jahns, 1953; Jahns and Burnham, 1969; Černy1991; London, 2005, 2008; Yuan et al., 2018; William et al., 2024). In the fractional crystallization process, the primary minerals that crystallize from the melt are alkali feldspar, sodic plagioclase, quartz, and smaller amounts of muscovite or biotite. As these silicate minerals form, they incorporate elements such as Si, Al, K, Na, O, and minor amounts of Fe and H_2O from the melt. During fractional crystallization, fluid (H_2O), flux (B, F, and P), and incompatible element concentrations surge in the residual magma (Trueman, 1982; Linnen, 1998; Breaks et al., 2005; London, 2005, 2018; Simmons and Webber, 2008). The late residual fraction of silicic melt, viewed as the medium forming pegmatite, is expelled from crystal-laden mush to create small pools within the source granite or extruded dikes (London, 2005). This model implies even upper mantle-derived mafic magma contribution during the final stage of fractional crystallization.

3.1.2 Direct Formation via Anatexis of Rocks in the Presence of Fluids

Crustal anatexis, prevalent during high-grade metamorphism in the lower to middle crust, has been linked to direct pegmatite formation during the anatexis of adjacent metasedimentary and metaigneous rocks (Simmons et al., 1995; 1996; Falster et al., 1997, 2005; Roda et al., 1999; London, 2005; Martin and De Vito, 2005; Lv et al., 2021). Peraluminous pegmatitic melts are supposed to be derived directly from metasediments, not genetically tied to fractional crystallization of magma. Evaporite-rich metasedimentary rocks may introduce fluxing components such as boron and lithium into pegmatitic melts. Thus, low-degree partial melts can directly generate pegmatite melts. Some pegmatites in the Oxford County pegmatite field in western Maine were formed by direct anatexis (Webber et al., 2019, 2023; Simmons et al., 2016, 2024). This anatexis model is supported by trace element analyses of Li, Rb, Sr, Cs, B, and Be, as well as chemical modelling of metasediments and leucosomes from migmatites in western Maine, as detailed by Simmons et al. (2022) and Webber et al. (2023). Additionally, high overburden pressure during orogenesis has been proposed as another potential cause of anatexis (e.g., Mukherjee 2013).

3.1.3 Liquid Immiscibility

Liquid immiscibility in hydrous silicate melts, considered a key process in the formation of pegmatitic parental magmas, arising from the parent magma during its differentiation (Peretyazhko, 2010; Thomas et al., 2012; Zagorsky et al., 1999). This process is distinct from the magmatic-hydrothermal model, demonstrating that immiscible hydrous silicate melt evolves during differentiation. Immiscibility, demonstrated by experimental and fluid inclusion studies, plays a role in enriching granitic melts with volatile and fluxing components (Thomas et al., 2000; Veksler et al., 2002; Thomas and Davidson, 2013). Initial closed-system crystallization involves crystallization of feldspar and quartz, forming graphic intergrowths. Further cooling induces enrichment in volatiles, leading to melt-melt immiscibility, especially involving H_2O, Li, Be, B_2O_3, CO_2, F, Cl, S, Rb, Cs, W, Nb, Ta, and Sn (Thomas et al., 2006). Gradual transformation results in the formation of hydrosilicate liquids, silicate, and silica gel. Liquid immiscibility, crucial for extreme rare element enrichment, typifies late-stage pegmatite formation (Smirnov, 2015).

3.2 Evolution with Syn and Post-Crystallization Modification

Several studies emphasize syn-crystallization modification of pegmatitic melt and post-crystallization evolution happens due to hydrothermal fluid interaction (Černý, 1972; Linnenand Williams-Jones, 1994; Badanina et al., 2015; Dostal et al., 2015; Ballouard et al., 2020a, b; Kaeter et al., 2018; Wu et al., 2018; Ballouard et al., 2022a). This modification, is pivotal for the genesis of economically significant minerals in pegmatites (Černý, 1972), involves hydrothermal fluid interaction at late pegmatite genesis stages, potentially yielding metasomatic transformation. Hydrothermal fluids play a vital role in the formation of pegmatite-related rare metal deposits by mobilizing elements around granitic intrusions through processes like exsolution and metasomatism (Thomas et al., 2011; Badanina et al., 2015; Dostal et al., 2015; Ballouard et al., 2020a; Kaeter et al., 2018; Wu et al., 2018). Often, magma and hydrothermal fluids are immiscible (Ballouard et al., 2020a, b). Components like H_2O, B, F and P play a significant role in lowering pegmatite formation temperature, enhancing miscibility among less soluble constituents (London, 1986a; London et al., 1987, 1997; Keppler, 1993; Thomas et al., 2000; Sowerby and Keppler, 2002). Melt-fluid immiscibility during the magmatic-hydrothermal transition proves to be an effective mechanism for rare metal mineralization in pegmatites (Zhou et al., 2023). Table 3 summarizes some selected studies on syn and post-crystallization modifications in global Pegmatites.

In contrary, London (1992, 2005) argued against the necessity of a hydrous vapour phase for pegmatitic texture development. Others (e.g., Masoudi and Yardley, 2005 and Daele et al., 2018 from the Borujerd complex of western Iran and Rwamagana-Musha-Ntunga pegmatite-quartz vein field of East Rwanda, respectively) suggested the potential mixing of magmatic and metamorphic fluids during pegmatite development. Additional reports indicate the mixing of meteoric fluid with magma at low pressure for pegmatite production (Thomas et al., 2003; Kaeter et al., 2018).

4. Geochemical Signature of Granitic Pegmatitic Melt

The chemical composition of granitic pegmatitic melt includes fluids, volatile fluxing components, iron, and molecules, including rare elements.

4.1 Role of Fluids, Volatiles, and Fluxing Components

The majority of granitic pegmatites represent “simple” chemical compositions that represent the system $NaAlSi_3O_8$–$KAlSi_3O_8$–Al_2O_3–SiO_2 (London et al., 2012). The profound role of water in pegmatite formation, as emphasized by Niggli (1920), finds support in the evidence provided by Jahns & Tuttle (1963) and Jahns & Burnham (1969), suggesting pegmatite formation occurs through equilibrium crystallization of coexisting granitic melt and hydrous fluid. Using the phase relations in the peralkaline [$Al_2O_3 < (Na_2O + K_2O)$] portion of the Na_2O–Al_2O_3–SiO_2–H_2O system, Mustart (1972) further elucidated the potential of peralkaline melts to dissolve elevated concentrations of H_2O as temperature decreases. However, experimental, fluid inclusion, and melt inclusion studies emphasize the pivotal influence of water in pegmatitic melt, impacting mineral melting and crystallization dynamics (e.g. Fig. 02). Water, along with other volatiles (H_2O, F and Cl) and semi-volatiles (B_2O_3 and P_2O_5), not only reduces the viscosity of aluminosilicate melts but also enhances ion transfer through the melt to promote unusually large crystal growth (see Tomas et al., 2006, for details). The ability of water to transport incompatible elements contributes to the crystallization of residual melt until the separation of water vapour and silicate-rich melt (Fig. 02). The interaction between water and silicate-rich melt during this phase yields distinctive igneous textures (Simmons and Webber, 2008). Additionally, volatile components like carbonate/bicarbonate/CO_2 play a significant role in granitic pegmatite evolution (Thomas et al., 2011; Liu et al., 2022a; 2022b), facilitating low-viscosity melts and fostering large crystal growth (Thomas et al., 2011).

Granitic magma, recognized as a complex suspension, composes dissolved gases, including ~6% water in residual magma (Roedder, 1992), along with Cl, F, and S gases and acids (Tyrrell, 1978). Some magmatic-hydrothermal fluids display a dominance of coexisting vapour and complex multiphase brines (Kamenetsky et al., 2002; Webster et al., 2004). Continuous cooling of volatile-enriched pegmatitic melt leads to the development of two or more melt fractions due to their propertis of liquid immiscibility, characterized by lower viscosity, facilitating separation from the larger magma body (Thomas et al., 2006). Below 750 °C, granitic melt typically contains substantial fluxing components, especially F and alkali metals (Smirnov, 2015). Experimental studies by Kovalenko (1979) showed that under H_2O-saturated conditions, the granite minimum temperature drops to 550–600 °C with the addition of fluorine. Dingwell et al. (1996) suggested a decrease in the temperature of the quartz-albite-orthoclase (haplogranite) system equilibrium with an aqueous fluid containing 7–9 wt.% B2O3 to 620–640 °C. Several authors (e.g. Taylor et al., 1979; Linnenand Williams-Jones, 1994; Webster et al., 1997; Zaw, 1998; Thomas et al., 2006,2012; Fei et al., 2018; Wanget al., 2023; Lv et al., 2024) highlighted the incorporation of a number of potential flux components, which may include H_2O, OH^-, CO_2, HCO^{-3}, CO_2^{-3}, SO_2^{-4}, PO_3^{-4}, H_3BO_3, F, and Cl, as inclusions in different mineral phases in pegmatites. Fluid inclusion analysis by Smirnov (2015) revealed that pegmatite minerals often contain trapped melts with high alumina contents, with F playing a crucial role in the glasses of melt inclusions, reaching up to 5 wt.%.

4.2 Rare Elements Concentrations

During the continuous fractional crystallization of magma, the residual melt eventually becomes saturated in incompatible elements and fluxing components (Thomas et al., 2006: Thomas et al., 2019; Troch et al., 2022). The high enrichment in granitic pegmatites makes them a classic source for a diverse range of trace elements such as Li, Rb, Cs, Be, Ga, Sc, Y, REE, Sn, Nb, Ta, U, Th, Zr, and Hf. Fractional crystallization may enrich trace elements in pegmatites up to some wt.% levels (e.g. London, 2008; Fredriksson, 2017). Most pegmatites, though compositionally similar to simple granites, lack evidence for fluxing components such as B, P and F, as well as high concentrations of normally rare elements like large-ion lithophile elements (LILE) and high field-strength elements (HSFE) (London, 1992). For instance, a positive correlation exists between Li, Rb, Cs, F, Be, Ga, B, P, Mn, Nb, Ta, Zr, Hf, Sn, and U with increasing differentiation of peraluminous pegmatites (e.g., Černy et al. 1985). Shaw et al. (2016) argued that rare-metal pegmatites are also distinguished by their bulk-rock and mineral chemistry, including a more peraluminous character and enrichments in Rb, Li, Cs, Be, Nb and Ta. London and Kontak (2012) suggested that the process of rare-element enrichment in pegmatites appears to occur in a closed system, originating from a small fraction of residual silicate liquid derived from a much larger magma body. In contrast, hydrothermal fluids in chemically open systems can contribute to the addition of rare elements (London et al., 2012).

5 Cooling History, Rates of Crystal Growth and P-T Evolution of Granitic Pegmatites

5.1 Influence of Cooling Rate on the Formation of Pegmatitic Textures

The dynamics of cooling in shallow-level pegmatites, as proposed by various researchers, span from weeks to years based on their respective sizes (e.g., Chakoumakos and Lumpkin, 1990; Webber et al., 1997, 1999; Morgan and London, 1999; London et al., 2012b). It is widely acknowledged that the principles governing mineral crystallization in conventional magmatic systems can not be directly applied to pegmatites. For instance, experimental studies on quartz and feldspar crystallization in granitic melts (e.g., Swanson & Fenn, 1986, 1992; MacLellan & Trembath, 1991; Fenn, 1977, 1986) have revealed that skeletal and graphic morphologies denote rapid crystal growth from a highly undercooled melt (Simmon and Webber, 2008). Textural relationships in pegmatites, such as needle-like, skeletal, branching, and wedge-shaped crystals, indicate rapid cooling, higher growth rates, and larger degrees of undercooling, with fewer nucleation sites. In contrast, tabular to equant crystal forms are indicative of slower cooling rates, lower growth rates, smaller degrees of undercooling, and abundant nucleation sites. Notable examples include the Harding pegmatite in New Mexico (Chakoumakos and Lumpkin, 1990) and the George Ashley, Mission, Stewart, and Himalaya dikes in San Diego County, California (Webber et al., 1997, 1999), which underwent rapid cooling, reflecting unusual growth rates in the cooling history of the pegmatites (Simmon and Webber, 2008). Cerný (2005) suggested that even the Tanco pegmatite, with a thickness of 100m appears to have cooled rapidly within decades to a maximum of a few hundred years. Webber and Simmons (2007) utilized a cooling model along with the parameters provided by Černý (2005), which include a thickness of 100 meters, an emplacement temperature of 700 °C, an emplacement depth of 10 kilometers, and country rock temperatures ranging from 25 °C to 300 °C. They calculated that the Tanco pegmatite cooled from 700 °C to approximately 450 °C over a period of about 700 to 1,000 years.

Rapid cooling rates in pegmatites challenge the prevalent notion that large crystals in pegmatites result from very slow rates of cooling and crystal growth. The phenomenon of rapid cooling with the formation of large crystals (e.g., Mc Birney and Russell, 1987; London, 1992, 1999; 2005) is explained by the concentration of incompatible elements, fluxes, volatiles and rare earth elements in the residual magma during the initial stages (e.g., Dingwell et al., 1996; Linnen et al., 2012). The presence of fluxes and volatiles, lowering crystallization temperature, reducing nucleation rates, melt polymerization and viscosity, and increasing diffusion rates and solubility, is considered crucial for the development of large crystals and pegmatitic textures (Simmons and Webber, 2008). Consequently, granitic pegmatite grains exhibit extreme variation in size (Jahns, 1955; Sinclair, 1995). Given that a majority of pegmatites form at shallow depths of the crust (around 3–5 km, Kozlowski, 1978), the H_2O solubility in aluminosilicate melts strongly depends on pressure over temperature (e.g., Holtz et al., 2001). Holtz et al (2001) noted that at the eutectic point (2 kbar/ 680 °C) in the quartz-albite-orthoclase-water system, the H_2O content is about 6.4 wt% and the viscosity is approximately 105.5 Pa.s. Under such high-viscosity conditions, the formation of typically coarse-grained or giant-textured pegmatites becomes challenging (Simmon and Webber, 2008). Pegmatite evolution involves disequilibrium crystallization from an undercooled, flux-bearing granitic melt, which is necessarily flux-rich and does not require the presence of an aqueous vapour phase (Morgan and London, 1999; London, 2005b; Simmon and Webber, 2008).

London (1999, 2005, 2018) proposed a model to address the viscosity problem, assuming a boundary layer enriched in incompatible fluxing components along the margins of the growing crystal front. Fluxing components aid in lowering the viscosity and solidus of granitic melts, enhancing the solubility of components that would otherwise precipitate as accessory minerals, and promoting the rapid growth of large, perfect silicate crystals (London et al., 2012). The enrichment of fluxing components in the pegmatitic melt not only facilitates rapid diffusion of elements into the crystal structure of the growth surfaces (London, 2009; London, 2014) but also influences the formation of fluid inclusions in the larger host crystals (London, 2015). When crystal formation begins, incompatible components, including the fluxes, are rejected at the growing crystal interface of quartz and feldspar (London (2005). These components then concentrate along the margins of the growing crystal front, forming a fluxed boundary layer of melt. Although the overall concentration of fluxes in the pegmatitic melt is low, a boundary layer can form if the diffusion of excluded components through the melt is slower than the rate of crystal growth. This boundary layer, being richer in fluxes, may exhibit a lower solidus temperature and increased silicate-H_2O miscibility. A model based on the high viscosity of the pegmatite-forming medium (supercooled silicate liquid, gel, or glass) was proposed by London (2005) to explain why granitic pegmatites are more common than those of basic or alkaline composition. The author suggested that due to the higher viscosity of granitic liquids, the diffusion of elements necessary to nucleate crystals decreases. The higher viscosity of granitic liquids also impedes the diffusion of excluded components back into the melt, promoting boundary layer formation that results in pegmatite textures. Alternatively, considering metasomatic processes, Beus (1983) proposed a model for the giant growth of crystals embedded in hard rock.

After conducting experimental studies at 480–700 °C and 220–960 MPa using H_2O-saturated melts of a typical granitic composition as well as granitic melts enriched in lithium (8800 ppm), Maneta and Anderson (2018) showed that euhedral quartz and alkali feldspar megacrysts crystallized over a period of hours with an average growth rates ranging from 3 (Li-free charges) to 41 cm/year (Li-enriched charges) for quartz and from 18 (Li-free samples) to 58 cm/year (Li-enriched samples) for alkali feldspars. Using the above experimental data, these authors argued that crystals formed from a silicate melt in the presence of a coexisting aqueous phase crystallize rapidly. In addition, crystals formed in miarolitic pockets can also crystallize rapidly.

5.2 Pressure-Temperature *(P-T)* Evolution

Determining the *P-T* evolution of many granitic pegmatites is challenging due to the absence of proper mineral assemblages for thermobarometry. Nevertheless, several studies attempted to speculate on the *P-T* evolution of pegmatites using experimental data, stable isotopic data, fluid inclusion analysis, and thermal modelling (e.g., Jahns and Burnham, 1969; Babu, 1969; Taylor et al., 1979; Kosukhin et al., 1984; Simmons et al., 1987; Konovalenko, 1988; Chakoumakos and Lumpkin, 1990; Nabelek, 2003a, b; London et al., 2012; Fei et al., 2021; Wang et al., 2023). Table 4 summarises the estimated P-T condition of selected pegmatites/pegmatite fields in the world throughout the history of pegmatite studies.

Jahns and Burnham (1969) proposed that pegmatites crystallize at or near the hydrous minimum melt temperatures of about 600 °C. Babu (1969), based on the chemical compositions of associated granitic gneisses, argued that crystallization of different mineral varieties in the pegmatite in the Nellore Mica Belt, Andhra Pradesh, India, occurred between 760 – 300 °C. The colour-temperature relationships and the associated minerals in the pegmatite prompted Babu (1969) to propose that the crystallization temperature range for tourmaline is between 760 °C and 450 °C, while for beryl, it is between 500 °C and 400 °C. Muscovite indicates a minimum crystallization temperature of 500 °C and 435 °C, based on the relationship within the paragonite-muscovite solid solution (Babu, 1969). Stable isotopic studies on pegmatites from San Diego County, California, led Taylor et al. (1979) to argue that temperatures of pegmatite melt emplacement ranged between 730 °C and 700 °C. The authors suggested that supersolidus crystallization began with the basal aplite zone in that pegmatite occurring around 565 °C, whereas the subsolidus formation of gem-bearing pockets took place at approximately 565–525 °C in a nearly closed system.

Kosukhin et al. (1984) demonstrated that the minerals of granite pegmatites, containing mineralized cavities (miaroles), crystallize between 515–750 °C and 0.5–3.0 kbar. Bakumenko and Konovalenko (1988) reported the lowest temperatures of homogenization of melt inclusions (505–510 °C) from miarolitic quartz from the Vezdarinskaya vein (SW Pamirs). Chakoumakos and Lumpkin (1990) found that the crystallization of the Harding pegmatite in Taos County, New Mexico, USA, occurred at 650°C and 3.3 to 3.5 kbar and subsequently cooled isobarically to 550°C. This conclusion was based on experimentally determined liquidus and solidus temperatures, phase assemblages for a bulk sample of the pegmatite, and isochores for CO_2-H_2O-NaCl fluid inclusions in quartz and beryl.

Simmons and Webber (2008) summarized the crystallization temperature conditions of various granitic pegmatites. They noted that the intermediate zones of pegmatites in South Platte, Colorado, recorded temperatures between 550 °C and 500 °C (Simmons et al., 1987) based on two-feldspar thermometry. In contrast, the Little Three pegmatite in California showed temperatures of approximately 400 °C to 435 °C near the margins, 350 °C to 390 °C close to the pegmatite pocket zone, and a sharp decline to 240 °C to 275 °C within the pockets (Morgan and London, 1999) Nabelek et al. (1992a, b) estimated that the equilibration temperatures for coexisting quartz and K-feldspar in the cores of various pegmatites in the Black Hills, South Dakota, were about 350 °C. For the Li-bearing Tin Mountain pegmatite in the same region, Sirbescu and Nabelek (2003a, b) proposed that crystallization occurred from fluid-rich, compositionally complex melts at around 400 °C to 350 °C. The lower crystallization temperatures were attributed to the combined fluxing effects of Li, B, P, H_2O, and carbonate anions (Sirbescu and Nabelek, 2003a, b). Evidence from thermal models combined with mineral compositions, as reported by London et al. (2012a), indicates that crystallization in pegmatites starts at approximately 450 °C, which is about 200–250 °C lower than the liquidus temperature where crystallization is expected to begin. The Lijiagou spodumene pegmatites in the Songpan-Garze Fold Belt, southwestern Sichuan Province, China crystalized at a high-temperature of 660 -740 °C at moderate pressure of 4.7 - 6 kbar (Fei et al.,2021). Wang et al. (2023) attempted to study the *P*-*T* condition of pegmatite after considering the *P*-*T* evolution of host metamorphic rocks and the mineralogical evolution of pegmatites in Lhozhag, eastern Himalaya. The authors interpreted the crystallization temperature of the pegmatite to be around 580 °C at a pressure of 4 kbar.

6. Classification Scheme for Granitic Pegmatites

In 1920, the Swiss mineralogist Niggli proposed a classification system for pegmatites that consisted of two levels: types and sub-types, which were categorized according to their chemical composition and mineral content. He highlighted that the distinct mineralogy and texture of pegmatites are mainly influenced by the type and amount of volatile components found in the pegmatite melts. Throughout the last century, efforts were made by Russian and French authors to classify pegmatites. Later, in 1930, Russian mineralogist and petrologist Fersman introduced the first internationally recognized classification of pegmatites. This classification was based on the thermal evolution of pegmatite melts.

Notably, Landes (1933) categorized pegmatites into acid, intermediate, and basic varieties. Subsequent classifications considered emplacement depth, metamorphic relationships, granitic activity, geochemistry, and the association with pegmatite-host rocks (e.g. Ginzburg and Rodionov, 1960; Ginsburg et al., 1979; Černý, 1991; Zargosky et al., 1999; Fuchsloch et al., 2017). Additionally, internal characteristics such as zoning, feldspar composition, grain size variation, and textural patterns were also included as classification criteria (Abdelfadil et al., 2016). The classification of granitic pegmatites is acknowledged to be more intricate than that of other igneous rocks. The following section provides a summary of each proposed classification scheme. A summary of existing classification schemes of granitic pegmatite is available in Table 5.

6.1 Classification based on Depth of Emplacement

Pegmatites were classified into four geological classes, considering factors such as depth of emplacement, metamorphic relationships, and their associations with granitic plutons (Ginzburg and Rodionov 1959; Ginsburg et al. 1979).

These classifications include (i) very deep-seated uranium–REE pegmatites (emplacement deeper than 10–11 km), (ii) great depth for mica-bearing pegmatites (emplacement at 7–11 km), (iii) moderate depth for rare-metal pegmatites (emplacement at 3.5–7.0 km), and (iv) minor depths for rock-crystal pegmatites (emplacement depth at 1.5–3.5 km) (Beskin and Marin, 2018).

6.2 Classification based on *P-T* Conditions

Fersman (1930) attempted to classify granitic pegmatites based on crystallization temperature, considering the thermal stability of minerals and mineral assemblages. The above author referred to the crystallization temperature starting with ~700 °C as 'ordinary allanite-monazite pegmatites' and the lower limit representing ~200 °C as 'zeolite pegmatites'. Crucial temperature boundaries at ~600 °C and ~400 °C, known as the β-/α- quartz transition (Fig. 4) and the critical point of water, respectively, were used to divide the thermal evolution into pegmatitic (600 °C), pneumatolytic (600–400 °C), and hydrothermal (400 °C) stages (Fig. 4). Fersman (1930) outlined ten main thermal types of pegmatites and added an eleventh type for "fillings in miarolitic cavities." He also described three compositional lineages: pure (no wall rock interaction), contaminated (affected by wall rocks), and migmatitic (intense interaction with wall rocks). These classifications are further divided based on mineralogy. He outlined ten main thermal types of pegmatites, with the pure (non-contaminated) lineage subdivided into 18 sub-types based on mineralogy and the contaminated lineage divided into 9 sub-types (Fig. 4). He added an eleventh type for "fillings in miarolitic cavities." He described three compositional lineages: pure (no wall rock interaction), contaminated (affected by wall rocks), and migmatitic (intense interaction with wall rocks). Fersman (1930) used temperature-indicative minerals and mineral parageneses to establish the temperature of pegmatites, noting that 250 indicative minerals could be organized into a "mineralogical-thermal evolution scheme" for this purpose. Despite the value of temperature-based classification, he acknowledged that factors like melt origin and contamination also play a significant role. Therefore, he superimposed compositional lineages on the temperature classification for a more comprehensive framework, recognizing that most pegmatites crystallize across multiple thermal intervals.

Černý (1991a) revised the above classification, considering emplacement depth, metamorphic grade, and minor element content, resulting in four main categories: Abyssal, Muscovite, Rare Earth element, and Miarolitic pegmatites (Table 6). Abyssal is associated with high grade, intermediate to high pressure facies, analogous to an upper amphibole facies environment (~700 – 800 °C and 4-9 Kbar), while muscovite is linked to high-pressure and low-temperature facies similar to amphibolite-barrovian kyanite facies (580 - 650 ℃ and 5 -8 kbar; Fig. 3). Rare elements are associated with low temperature and low pressure, resembling amphibolite - upper greenschist facies (500 - 600 ℃ and 2-4 kbar), and the miarolitic class indicates shallow levels (1 to 2 kbar; Fig. 3).

6.3 Classification based on Composition

Niggli (1920) proposed a two-level classification based on chemical characteristics and mineralogy. In this classification, the specific mineralogy and texture of pegmatites primarily determine the type and amount of volatile components. Niggli introduced fifteen pegmatite types associated with granitic and syenitic magmas. In Černý's (1991) classification, rare-element pegmatites were further categorized into two families (Table 6): the LCT family (Li, Ce, and Ta enrichment) and the NYF family (Nb, Y, and F enrichment). These families were then subdivided based on mineralogical or geochemical characteristics. For instance, the LCT family was divided into four subtypes: beryl, complex (rare-element), albite-spodumene, and albite types, with the beryl type further subdivided into two sub-types: beryl-columbite and beryl-columbite-phosphate (Table 6). Zargosky et al. (1999) suggested that miarolitic facies can occur to varying degrees in almost any pegmatite sequence. Based on the type of granite plutons, the classification has three main classes based on the aluminum saturation of the parent granite and mineralogy. Based on mineralogy, Wise (1999) classified NYF pegmatites into three classes named peralkaline, metaluminous, and peraluminous. This classification includes six types and nine sub-types. Pezzotta (2001) attempted to classify pegmatites according to old depth-related classification schemes but modified the older classification after considering pegmatite mineralogy, which relates to the pegmatite bulk chemistry. Pezzotta (2001) classified pegmatites into three classes based on emplacement depth structure and mineralogy. According to this classification: *Class I*: Abyssal (K-feldspar and corundum subtypes) is formed at low pressure and high temperature, poorly mineralized but rich in ceramic minerals such as quartz and feldspars. *Class II*: The rare-element class (Beryl, rare earth type, and complex subtypes are included) is the most mineralized and most mineralogically diverse. This class includes NYF and LCT types. *Class III*: NYF miarolitic is formed under a low-pressure regime. This type of pegmatites can be seen at shallow levels, and miarolitic cavities are common. Ercit (2004) found a low degree of correlation between accessory mineralogy and depth of emplacement for NYF pegmatites. Ercit (2004) divided NYF pegmatites into four categories: Abyssal, Muscovite–rare element class, and Miarolite classes. The abyssal class was further divided into two subclasses as the allanite monazite-uraninite subtype and the (Y, REE)-Nb-oxide subtype. Meanwhile, the rare-element class also has three subclasses, namely allanite-monazite, euxenite, and gadolinite.

Černý and Ercit (2005) proposed a classification for pegmatites by combining the systems proposed by Černý (1991a) and Ercit (2004), with several changes that address NYF pegmatite classification and a petrogenetic classification of pegmatites derived from plutons. In Černý and Ercit's (2005) classification, three families were distinguished: (i) the NYF family with progressive accumulation of Nb, Yand F, fractionated from sub-aluminous to meta-aluminous A- and I-granites, (ii) the LCT family, commonly accumulated Li, Cs, and Ta and derived mainly from S-granites and rarely form I-type granite, (iii) mixed by the NYF + LCT family of diverse origin.

Recently, Wise et al. (2022) proposed a classification scheme for pegmatites based on their association with primary accessory rock-forming and rare-element minerals (Table 7), and pegmatites are divided into three main groups as Groups 1, 2 and 3. According to the Wise et al. (2022) classification, the pegmatites of Group 1 are typically enriched in Li, Rb, Cs, Be, Ga, Sn, Ta, Nb, B, P and F. Three types of pegmatites in this group are frequently recognized as (i) beryl phosphate-bearing types, (ii) Li-rich spodumene- or petalite-dominant types, and (iii) bodies with lepidolite or elbaite as the principal Li phase. In the pegmatite Group 2, the most common granitic pegmatites are composed mainly of quartz and feldspar. The presence of small but significant amounts of accessory minerals, such as helvite, magnetite, fluorite or hyalite opal, serves to distinguish them from Group 1 pegmatites. The pegmatite Group 3 is strongly peraluminous, with essential K-feldspar, quartz, and plagioclase, and accessory biotite, muscovite, garnet or tourmaline. The authors suggested that pegmatites belonging to Groups 1 and 2 are generated from the residual melts of S-, A-, and I-type granite magmatism and are also direct products of anatexis. The Group 3 pegmatites are only derived by anatexis.

In evaluating the classification of pegmatites, it is essential to recognize the complexities and evolving nature of the classification systems proposed over the last century. The early works (e.g., Niggli, 1920; Fersman in 1930; Landes, 1933) laid the groundwork by focusing on chemical composition, mineralogy, and thermal evolution. Although pioneering, these schemes were limited by the scientific knowledge of their time. As subsequent researchers built on these foundations, classifications became increasingly sophisticated, incorporating not only chemical and mineralogical data but also emplacement depth, geochemical signatures, and tectonic settings (e.g., Černý, 1991a; Wise, 1999; Ercit, 2004).

However, this increasing complexity also presents challenges. For instance, the classification systems by Černý and Ercit (2005) and Wise et al. (2022) aim to address the diverse origins and geochemical signatures of pegmatites, but still fall short in accommodating all variations observed in nature, particularly in hybrid or mixed-origin pegmatites. Additionally, the distinction between LCT and NYF families, while useful, may oversimplify the spectrum of pegmatite compositions, especially where there is significant overlap or transitional characteristics. Furthermore, some recent classifications, which categorize pegmatites based on their primary accessory minerals and rare-element enrichment, introduce a more detailed perspective yet still require further empirical validation across different geological settings. A critical gap in current classification schemes is their application to field-based studies, where variability in pegmatite formation conditions can be highly localized and influenced by factors not fully accounted for in generalized models. This highlights the need for more flexible, region-specific classification frameworks that can adapt to local geological contexts while still providing a coherent structure for broader comparison.

7. Economic Potential of Granitic Pegmatites

Granitic pegmatites continue to be crucial sources of industrial minerals, coloured gemstones, rare metals, and energy minerals (e.g., Glover et al., 2012; Simmons et al., 2012; London et al., 2012b; London, 2014). Despite industrial minerals such as quartz, feldspar, and mica being found in the majority of rocks commonly as medium to fine grains, the exceptionally coarse grain size of these minerals in granitic pegmatites significantly enhances their economic value up to a potential mining grade. The exotic nature of the pegmatitic melt, enriched with incompatible elements and fluxing components, provides favourable conditions for the crystallization of coloured gemstones, rare metals, and energy minerals, as seen globally.

7.1. As a Source of Industrial Minerals

As granitic pegmatites primarily consist of quartz and feldspars, with some mica-family minerals, certain ore-grade deposits approach almost 100% minable—a rare advantage in the mining industry (London et al., 2012b). On average, pegmatites consist of feldspar (65%), quartz (25%), mica (5-10%, muscovite 6%), and accessory minerals such as spodumene, tourmaline, oxides, garnet, topaz, and muscovite (Kesler et al., 2012; Eagle et al., 2015). Some pegmatites contain kaolin as a secondary mineral, derived from the intense acidic alteration of primary feldspar (Glover et al., 2012). Pegmatites sought primarily for their industrial minerals are referred to as barren, indicating a lack of gems and rare earth elements, and are also known as ceramic pegmatites (Glover et al., 2012).

Various varieties of quartz, such as milky, water-clear, and smoky quartz, are prominent in granitic pegmatites. Quartz is typically purer than other minerals (Glover et al., 2012), although it may sometimes contain Li, Na, K, Ti, and certain impurity minerals (e.g., Müller et al., 2015). Based on purity levels, pegmatitic quartz is divided into two types: high-purity quartz (<100 ppm total impurities by weight) (Larsen et al., 2000; Glover et al., 2012), containing less than 50 ppm trace element concentration (Harben, 2002), and low concentrations of elements other than silicon and oxygen (>99.995% SiO_2) (Larsen et al., 2000; Müller et al., 2015). However, a few pegmatite bodies contain quartz pure enough to produce high-purity quartz powder [High-purity quartz powder is produced from pegmatites in Brazil, India, Australia, Madagascar, Norway, China, the USA, and Russia (Glover et al., 2012)]. The chemistry of pegmatite quartz is mainly controlled by the origin (chemistry of the source rock) of pegmatitic melts and, occasionally, the geodynamic setting of the pegmatite fields and provinces (Müller et al., 2020).

Albite and microcline are common feldspar varieties found in ceramic pegmatites. These alkali-feldspars crystallize over extensive periods of pegmatite solidification and serve as suitable hosts for Rb, Cs, Ca, Sr, Pb, Ba, Eu, La, and Li in the alkali position, and B, Ga, Ge, Fe, Mg, Ti, and P in tetrahedral sites (Oyarzábal et al., 2009). The different feldspar types have various applications related to their melting temperature and are used in various industries according to specific requirements.

Kaolin deposits and occurrences could be primary or secondary depending on their genesis (Bailey, 1980; Dixon, 1989; Murray, 1999 a, b, c; Ekosse, 2010). Primary deposits form due to the alteration of crystalline rocks, while secondary deposits result from the erosion of primary deposits (Prasad et al., 1991). Most kaolinite deposits form as a consequence of hydrothermal leaching or weathering of feldspars. Additionally, acid leaching helps remove ions such as calcium, sodium, and potassium, resulting in the production of kaolin or kaolinite (Glover et al., 2012).

7.2. As a Source of Strategic Metals

Strategic metals, pose the greatest risk of supply disruptions globally and play a vital role in a country's economy or defense (Linnen et al., 2012). The availability of strategic metals varies from country to country, with key metals including Li, Ta, Nb, Be, Sb, W, Co, and rare earth elements. These metals are commonly found in rare-element pegmatites, which are mineralogically complex and enriched with incompatible elements (Cerný, 1991; Webber, 2008; Linnen et al., 2012; Bradley et al., 2017; Lv et al., 2020; Rosing-Schow et al., 2023). Notably, many of these incompatible elements align with strategic metals (Linnen et al., 2012; Hulsbosch et al., 2014).

During the initial stages of the crystallization process, quartz and feldspar dominate. Because strategic metals are highly incompatible with these minerals, their concentrations in residual melts experience a significant increase. For example, caesium (Cs) as a highly incompatible element due to its larger ionic radius, except in mica, feldspar, and cordierite, where it is moderately incompatible (London, 2005). Rubidium is also an incompatible element but is compatible with mica and alkali feldspar. Beryllium is normally incompatible due to its small ionic radius but is compatible with cordierite and muscovite (Linnen et al., 2012). This type of pegmatite also contains minerals with Li, Nb, W, Cs, and Be, which are by-products of tin- and tantalum-pegmatite mining (Sweetapple et al., 2002).

The most granitic pegmatites containing quartz and feldspar have a low concentration of rare-earth elements (Stilling et al., 2006; Linnen et al., 2012). Rare-earth element pegmatites make up only 1% - 2% of all pegmatites (London, 2014). They consist of elements such as Be, Li (spodumene and petalite), Sn (cassiterite), Rb (lepidolite), Ta and Nb (Ta-oxide minerals), Cs (pollucite), REEs, and U, which may be highly enriched in pegmatites compared to the bulk continental crust. Some enrichment factors (relative to the continental crust) exceed 10,000 to 100,000 (Galeschuk and Vanstone 2007; Cerný 1990, 1991; Cerný and Ercit 2005). These are typically found in geographically restricted areas (Hotel, 2010), along large regional-scale faults in greenschist and amphibolite facies metamorphic terrains (Selway et al., 2005). The rare-element-rich granitic pegmatites represent the youngest, most fractionated end-members of layered sequences in upwardly differentiated plutons (Cerný et al., 2005; Linnen and Cuney, 2005; Melcher et al., 2015). Additionally, rare-element pegmatites contain fluxing compounds, which lower the viscosity and solidus temperature of the magma (Suwimonprecha et al., 1995; Melcher et al., 2015). Consequently, crystal-melt fractionation and the concentration of rare metals occur in residual melts, aided by liquid-liquid separation and fluid separation at the magmatic-hydrothermal transition (Cerný et al., 2005; Melcher et al., 2015).

The geochemical signatures of rare-element pegmatites are particularly intriguing as they represent the extremes of fractionation trends encountered in the final stages of magmatic differentiation, found in very leucocratic, high-silica, metaluminous to peraluminous granites, and rhyolites (Cerný et al., 1985). Rare earth, which represents LCT and NYF pegmatites, hosts several strategic metals (Cerný et al., 1991).

7.2.1. LCT Pegmatites as a Source of Strategic Metals

LCT (Lithium-Cesium-Tantalum) pegmatites are of significant importance in industrial applications (Melcher et al., 2015), encompassing elements such as Li, Cs, Ta, and Be, tantalum oxide, Be, B, F, P, Mn, Ga, Rb, Nb, Sn, and Hf, as well as ceramic-grade feldspar and quartz (London, 2017; Goodenough et al., 2019). LCT pegmatites are characterized by a high K/Na ratio, low Ca and Na content, high Al content, and a low oxidation state (Selway et al., 2005). Consequently, these pegmatites typically exhibit the presence of biotite and muscovite, while lacking hornblende (Selway et al., 2005). The LCT pegmatites as crystallized from magma resulting from the partial melting of preexisting sedimentary source rocks (Shelley, 1993; Selway et al., 2005).

Presently, most authors believe that LCT family pegmatites originate from S-type (Černý et al., 2012) peraluminous [Al-rich Al/(Na + K + 2Ca) > 1], quartz-rich granites associated with the melting of schists and gneisses of sedimentary origin (Černý et al., 2012, London, 2017). S-type granites typically form in areas with crustal thickening and are associated with subduction and continental collision. Additionally, LCT pegmatites are believed to be derived in syntectonic areas, as the sources of these granitic pegmatites lack foliation or pervasive deformation (Černý et al., 2012).

LCT pegmatites play a crucial role in supplying rare and strategic elements globally, contributing to one-third of the world's lithium production, most of the tantalum, and the entire world's caesium requirements (Geological Survey of the U.S., 2011). These pegmatites host various rare-element-bearing mineral phases, including beryl, spodumene, elbaite, columbite-tantalite, pollucite, and lithium phosphates (Dwight and Andrew, 2013). Dwight and Andrew (2013) estimated millions of tonnes (Mt) of ore, with indicated percentages of tantalum and lithium oxides (Ta_2O_5 and Li_2O), from larger scale deposits such as Tanco in Canada (2.1 Mt at 0.215 percent Ta_2O_5), Greenbushes in Australia (70.4 Mt at 2.6 percent Li_2O), and Bikita in Zimbabwe (12 Mt at 1.4 percent Li_2O).

7.2.2. NYF Pegmatites as a Source of Strategic Metals

NYF (Niobium-Yttrium-Fluorine) pegmatites encompass the rare-element-REE and miarolitic-REE subclasses (Selway et al., 2005) and are enriched with Nb, Y, F and occasionally contain economic resources of these metals (Goodenough et al., 2019; Eby 1990; Cerný and Ercit 2005; Martin and De Vito 2005; Simmons and Webber, 2008). These pegmatites are derived from A- and I- type granites (Cerný and Ercit 2005; Martin and De Vito 2005). The A-type of granites, which are subaluminous to metaluminous (Černý, 1991a; Selway et al., 2005), can be found in the lower crust but vary with the amount of added mantle components (King et al., 1997; Christiansen et al., 2007). The NYF-type granitic pegmatites form during peralkaline igneous rocks that fractionate directly from mantle sources. This type of granite has high K+Na, F, and Zr, low Al and Ca, and high Fe/Mg (Martin and De Vito, 2005; Selway et al., 2005), characterized by the presence of Fe-rich mica, pyroxene, and amphiboles (Selway et al., 2005). This family contains chemically complex oxides and silicates carrying heavy elements Ti, U, Th, and Nb over Ta, including minerals such as euxenite/aeschynite [$(Y,Ca,Ce,U,Th)\ (Nb,Ta,Ti)_2O_6$], allanite-(Y) [$CaYFe^{2+}Al_2Si_3O_{12}(OH)$], samarskite/fergusonite [$(Y,Fe^{3+},Fe^{2+},U,Th,Ca)\ (Nb,Ta)O_4$], and gadolinite [$(Y,Ca)_2Fe^{3+}Be_2Si_2O_{10}$] (Černý et al., 2012). Abundant fluorite or topaz in NYF pegmatites reflects their enrichment in fluorine (Černý et al., 2012).

7.4. As a Source of Gem-Bearing Minerals

All significant gem-bearing pegmatites share a granitic composition (Keller, 1990; Simmons et al., 2012), hosting more than 50 familiar gemstones occurring in various oxide, phosphate, and silicate minerals (London, 2008). Granitic pegmatites are vital sources for coloured gemstones such as tourmaline, topaz, spessartine polychrome, various coloured varieties of beryl like aquamarine, emerald, morganite, amazonite, the green variety of microcline, and coloured spodumene crystals like kunzite and hiddenites (Pezzotta and Laurs 2011; Simmons et al., 2012; Simpson, 1952; Hickman, 1983; Sweetapple, 2000). Although gem crystals are typically smaller than industrial minerals like quartz and feldspar, exceptional cases with larger gem-quality crystals like topaz, aquamarine, and tourmaline extending to several meters have been recorded

(Keller, 1990). Minas Gerais, Brazil, is renowned for large crystal findings, including the largest aquamarine crystal (110.5 kg) discovered in 1910 at the Papamal mine (Keller, 1990). Table 8 summarises the major gem-pegmatite districts and notable gem minerals in the world.

While gemstone pegmatites primarily originate in-situ, secondary gem pockets formed during transportation and erosion are also observed (Proctor, 1985). Brazil is particularly famous for gem-bearing pegmatites, with notable finds of blue-green, blue, and red tourmaline (Proctor, 1985). The Olhod'Água mine and Aracuai-Itinga district mining are well-known locations for gem-bearing pegmatites in the Eastern Brazilian pegmatite province (Proctor, 1985).

Sri Lanka, with its high abundance of gemstones, often associated with high-grade metamorphic rocks like granitic gneiss and pegmatites, is another significant source. Studies (e.g., Herath, 1980, 1985; Munasinghe and Dissanayake, 1980) suggest that the source of precious gem varieties found in gem gravels in Sri Lanka are likely pegmatite. Central Madagascar is also characterized by the presence of gem-bearing pegmatites (Pezzotta, 2001a, 2005b). Some gem minerals (e.g., tourmaline group minerals, beryl group minerals, spessartine) are more frequently hosted in granitic pegmatites than in other rock types. Cornwall, United Kingdom, and Phuket, Thailand are renowned localities for gem-quality tourmaline-bearing pegmatites (Suwimonprecha et al., 1995). Boron, an incompatible trace element, can be highly enriched in granitic melts and aqueous fluids, facilitating the crystallization of the tourmaline group minerals. The Beryl group contains various gem varieties because many ions can substitute the beryl crystallographic structure (McManus et al., 2008). Gem-quality beryl varieties like Aquamarine, Emerald, and Goshenite are found in well-known localities such as Sparrow Pluton (Yellowknife, NWT, Canada) and Helen Beryl Black Hills, USA (Abduriyim, 2006; Simmons et al., 2012). Spessartine, a gem-quality garnet variety, can be found in some pegmatites, with famous localities like Alto Mirador pegmatite, Escondido pegmatite, and Poaiá pegmatites in Brazil (Eeckhout et al., 2002). Ramona, California, is also renowned for spessartine gem-bearing pegmatites.

7.4.1. Origin of Gem-Forming Melts

Gem minerals are believed to form in pegmatites from rare elements that cannot be incorporated into the crystal structure of initially crystallized minerals like quartz, feldspar, and mica (Keller, 1990). Most gem-quality minerals originate during the final stages of the crystallization of a parent granitic melt (Simmons et al., 2012). The pegmatite-forming melts with the potential for crystallizing gem minerals mostly originate during the crystallization of a parental granitic melt (Simmons et al., 2012). During the fractional crystallization of magma, if the concentration of volatiles, mainly H_2O, exceeds its solubility limit in the remaining residual melt, an aqueous fluid exsolves from the melt, promoting the formation of a miarole or some primary pockets (e.g., Simmons et al., 2012). Keller (1990) points out that the abundance of hydrous minerals, including clays and zeolites, together with gem-quality minerals in pockets in Gem Pegmatites of Minas Gerais, Brazil, serves as evidence of the role of fluid in the final stages of crystallization. Sometimes gem-quality minerals intergrow with feldspar, quartz, and mica. Corundum is poorly associated with quartz but predominantly occurs in feldspars. The exact mechanism for the formation of these unusual melts of pegmatites is still unclear. One hypothesis suggests that these unusual pegmatites were stripped off silica from the surrounding rocks, allowing the formation of corundum. Rare gem varieties sometimes occur at the contact between the pegmatite body and its host rocks, interpreted as open system formation due to metasomatism (Schwarz et al. 2002; Zwaan et al. 2005; Groat et al. 2007; Keller, 1990).

7.4.2 Relations with Occurrences of Gem-Quality Minerals in relation to Pegmatite Anatomy

Gemstones typically constitute only a small fraction of pegmatite bodies (Shigley and Kampf, 1984). Gem-quality minerals are primarily found in the interiors of zoned pegmatites or reaction zones adjacent to pegmatites and host rocks. Three main settings within pegmatites where gemstones are commonly found have been summarized by Simmons et al. (2012): (i) *Setting 1*: Crystals "frozen" in massive quartz or feldspar in the core or margin of the core of a pegmatite: This setting includes minerals such as aquamarine and tourmaline in massive quartz or feldspar, usually with only parts of these crystals occurring as gem-quality. (ii) *Setting 2*: Crystals in reaction zones surrounding pegmatites that intrude mafic rocks: These crystals are produced from the interaction of Be-rich pegmatitic fluids with chromium-bearing mafic country rocks. Examples include alexandrite and emerald found in the Ural Mountains of Russia, the Mananjary area of Madagascar, and the Kafubu area of Zambia. (iii) *Setting 3*: Crystals in miarolitic cavities (pockets): Miarolitic cavities are crystal-lined voids or crystal-filled and clay-packed masses within igneous hosts. These pockets, comprising about 5% of the total pegmatite volume, contain gem-quality minerals. Factors influencing minerals precipitated in these cavities include source rock composition, the amount and composition of volatiles present, and pressure and temperature conditions during melt solidification. Gem varieties such as beryl, spessartine, spodumene, topaz, and tourmaline can occasionally be found in miarolitic cavities, along with minerals like kaolin, smectite, chlorite, oxides, carbonates, and other phyllosilicate group minerals.

The size of miarolitic cavities varies, and examples from around the world include gem pockets in the Dunton pegmatite on Newry Mountain, Maine, USA (up to 2 × 3 × 7 m sizes) containing bicolored, green, and red elbaite tourmaline crystals, and gem pockets in the Mount Mica pegmatite in Maine, USA (1.5 × 2 × 7 m sizes) producing large tourmaline crystals. Notable gem pockets are also found in the Jonas pegmatite (Conselheiro Pena district of Brazil), Volodarsk-Volynsky, Ukraine, and various other locations globally. Figure 5 represents a flow chart summarizing the development of gem quality crystals in miaroles in pegmatite. Figure 6 is a schematic diagram showing a cross-section through a pegmatite with gem pockets (modified after Petrel al., 2012).

7.5. As a Source of Energy Minerals

The identification of energy minerals is crucial for environmental stewardship, human health, and sustainable growth (Ferrero et al., 2013). Uranium and thorium, characteristic energy minerals, are often present in igneous rocks as part of the late-stage differentiation of granite or syenite (Virk and Singh, 1977). Besides industrial minerals, strategic metal-bearing minerals, and gemstones, certain pegmatites contain high levels of U- and Th-bearing minerals like uraninite and thorianite, are serving as potential sources of nuclear energy (Dill, 2015).

Uranium minerals in pegmatites can be categorized as primary and secondary based on their origin. Primary uranium minerals include simple oxides, multiple oxides (columbium, tantalum, and titanium), miscellaneous compounds, and carbonaceous material. Most primary uranium minerals undergo partial or complete alteration to secondary minerals such as hydrated oxides, phosphates, arsenates, sulfates, silicates, and carbonates (Tyler, 1950; Nete, 2009). Examples of uranium mineralization in pegmatites include uraninite, uranophane, beta-uranophane, and uraniferous pyrochlore. In some instances, uranium occurs as small grains embedded with mica flakes or garnet grains.

Significant uranium deposits in Gascoyne Province, South Africa, demonstrate pegmatites and aplite containing disseminated uraninite betafite and secondary uranium minerals (Berning et al., 1976). Madagascar is also renowned for uranium-bearing pegmatites, with locations like Ambatofotsy, Abatohasana, and Ambatofampykely containing large pegmatites averaging 12% - 15% U_3O_8 (Bourre, 1988). Studies on the Guangzhou uranium deposit in China reveal the enrichment of uraninite hosted in granitic pegmatites, occurring commonly as mineral inclusions in quartz, K-feldspar, and biotite. Post-tectonic magmatism and hydrothermal activities are suggested to contribute to the remobilization and circulation of uranium into pegmatites (Wu et al., 2021).

8. Tectonic Settings of Granitic Pegmatites

The global distribution of pegmatite ages implies a connection between the age of pegmatite formation and supercontinent assembly (Bradley, 2011; Bradley et al., 2017). Bradley et al. (2017) highlighted the correlation of LCT (Lithium-Cesium-Tantalum) pegmatite age distribution with supercontinent assembly throughout geological history. According to the authors, pegmatite ages from the Appalachians in North America and the Hercynides and Uralides in Eurasia, ranging from 370 to 275 million years ago (Ma), signify collisional events during the assembly of Pangea.

Overlapping peaks at 530 and 485 Ma in pegmatites from Brazil, Australia, India, Namibia, and Ethiopia coincide with the assembly of Gondwana. In the Precambrian, LCT pegmatites globally exhibited peaks at 960 Ma, 1800 Ma, and 2650 Ma, corresponding to the assembly of Rodinia, Nuna, and Superia supercontinents, respectively. Subduction of oceanic crust and continental-continental collision associated with supercontinent amalgamation provides favourable conditions for the formation of large igneous p phosphates, arsenates, sulfates, silicates, and carbonates (Tyler Rovinces (Klausen, 2020).

Tkachev (2011) noted that the occurrence of continental granitic magmatism and the emplacement of LCT pegmatites followed a consistent cycle of about 800 million years. Significant peaks in the global age distribution of LCT pegmatite fields were observed at 2650–2600 Ma, 1900–1850 Ma, 1000–950 Ma, and 300–250 Ma (Fig. 7). Tkachev (2011) also identified additional peaks at 2850–2800 Ma, 2100–2050 Ma, 1200–1150 Ma, and 550–500 Ma. These periods correspond with heightened continental crust magmatism and align well with Bradley's (2011) proposed supercontinent cycles. While continental crust magmatism dates back to 3850 Ma, no LCT pegmatites have been recorded prior to around 3150 Ma (Fig. 7). Each period of increased LCT pegmatite emplacement lasted approximately 50 million years.

Martin and De Vito (2005) extended the tectonic classification of granite magmatism to the field of granitic pegmatites. According to Martin and Vito (2005), LCT granitic pegmatites are part of orogenic suites associated with crustal shortening, occurring in zones of subduction and instances of continental collision. The parental magma is derived by partial melting in the mantle wedge within a tectonic environment characterized by overall compression, extending from the upper mantle to the lower crust, experiencing assimilation and fractional crystallization. Magma upwelling facilitates the partial melting of muscovite- and biotite-bearing metasedimentary sequences of the crust, resulting in contrasting magmas juxtaposed due to magma mixing and mingling. The extreme fractional crystallization of these magmas produces metaluminous to peraluminous granitic pegmatites involving both crust and mantle source materials to varying extents. Bradley (2019) noted the lack of a comprehensive study addressing the detailed plate tectonic settings of LCT pegmatites. However, they may form in possible tectonic settings involving partial melting of the continental crust, such as (a) subduction beneath a continental-margin arc; (b) slab failure (breakoff, delamination, or rollback) during or after collision; (c) overthickening of continental crust during collision, including variants involving shear heating and radiogenic heating; (d) extensional collapse during or after collision; and (5) rise of a mantle "plume" below a recently formed orogenic belt (Bradley, 2019). Bradley et al. (2017) argued that the 7 Ma LCT pegmatites in Elba Island, Italy formed through a combination of (b) and (d) when slab rollback during the Apennine collision caused the mantle to well up and partially melt beneath an extending collisional orogeny.

According to Martin and Vito (2005), NYF (Niobium-Yttrium-Fluorine) pegmatites are interpreted as products of anorogenic suites and mainly form in an extensional setting. These pegmatites represent the culmination of protracted fractional crystallization of basaltic magma. Crustal extension occurs above zones of anomalous upper mantle and active degassing, producing a mixture of H_2O, CO_2, and CH_4. Trace elements, including alkalis, rare earths, Nb, Ta, Zr, Ti, U, and Th, are mobilized upward into the attenuated crust. The lower and middle crusts are locally alkali-metasomatized before anatexis. Crustal anatexis produces a metaluminous to mildly peralkaline melt. Simultaneously, the rising mantle-derived Oceanic Island Basalt-type magma promotes the progress of anatexis. The uncontaminated basaltic magma may fractionate in a classic way or mix with derivative magmas through the partial melting of the metasomatized crust. The final product may be an NYF-type pegmatite. However, recent investigations (Müller, 2017) concluded that NYF pegmatites are not necessarily formed in an anorogenic tectonic setting in relation to A-type magmatism. They can result from both compressional or extensional orogenic settings unrelated to pluton-scale magmatism.

The tectonic settings of "Mixed" type (LCT + NYF) pegmatites have received little attention throughout history. Several authors (e.g., Černý, 1991b) have pointed out the association of LCT pegmatites with S-type granites (Chappell and White, 1992) and NYF pegmatites with A-type granites (Eby, 1990). Both these granite types are tectonically controlled, with A-types associated with extensional rift environments within continental cratons and S-types associated with the melting of accreted marine sediments in the waning stages of continental collisions (London, 2022). One possible tectonic environment where these two contrasting granite magma types might arise in close association is in the back-arc basins behind subduction-related volcanic arcs that extend into continental margins to produce mixed-type pegmatite (London, 2022).

Figure 8 represents the geological distribution of 127 pegmatite provinces and districts, distinguishing between LCT and NYF suites and some with hybrid affiliation mainly after Sánchez-Muñoz et al. (2017) and some from Liet al., (2023). These regions are predominantly associated with orogenic belts linked to crustal convergence, indicating a compressional tectonic regime. However, granitic pegmatites can also form in tectonically stable environments, signifying an extensional regime. Most large, mineralized pegmatites are found in collisional orogens, within districts containing numerous individual pegmatitic bodies. Such extensive pegmatite fields are generally absent in non-collisional settings, such as the western part of the American continent and the eastern part of Australia, due to the lack of thickened continental crust required to generate the necessary heat and pressure for melt formation (Tkachev, 2011; Sánchez-Muñoz et al., 2017).

Current research has advanced our understanding of pegmatite formation, yet the variability in tectonic settings—especially the role of specific processes like slab rollback, mantle plumes, and extensional collapse—remains an area for further exploration. Understanding these tectonic drivers in greater detail could provide further deeper insights into the conditions necessary for the emplacement of economically significant pegmatite bodies. Therefore, future studies should aim to refine these models, taking into account of regional geological context and integrating multidisciplinary data to better predict pegmatite occurrences and their mineralization potential.

9. Summary

Granitic pegmatites are a widespread rock type characterized by a granitic composition (Bradley and McCauley, 2017), representing chemical compositions of $NaAlSi_3O_8$–$KAlSi_3O_8$–Al_2O_3–SiO_2 (London et al., 2012). This rock type is texturally distinctive due to large crystals and is often marked by zonation, reflecting variations in grain size, mineral assemblage, crystal habit, or rock fabric. Most granitic pegmatites usually contain quartz, feldspar, mica, and over about fifty accessory mineral species (Bastin, 1910), including secondary minerals and rare earth elements.

During the crystallization of felsic magma, quartz and feldspar minerals become enriched in the residual melt. Concurrently, the content of rare earth elements, exotic components, and volatile components increases. These compositions are excluded in the initial crystallization of quartz and feldspar, leading to a residual melt eventually saturated in minerals containing these exotic components. Granitic pegmatites, therefore, exhibit high concentrations of various elements, including Li, Rb, Cs, Be, Ga, Sc, Y, REE, Sn, Nb, Ta, U, Th, Zr, and Hf, typically present at trace levels. Fractional crystallization may further enrich these elements in pegmatites up to higher levels, typically observable in few wight percentages (London, 2008; Fredriksson, 2017).

The strategic elements present in granitic pegmatites are a few thousand times higher than their crustal abundance (Cerný, 1991; Simmones and Webber, 2008; Linnen et al., 2012; Bradley et al., 2017). This enrichment of exotic chemical constituents facilitates the crystallization of economic commodities such as strategic metal enriched-minerals, precious and semiprecious gemstones, and energy minerals. However, rare element-enriched pegmatites are rare in nature, and the majority of pegmatites contain only industrial minerals, often referred to as barren or ceramic pegmatite.

Three main hypotheses propose the formation of pegmatites: (i) Fractional crystallization, (ii) direct formation by anatexis of surrounding rocks, typically in the presence of fluids, and (iii) Liquid immiscibility. Recent studies suggest that granitic pegmatite formation occurs in the presence of viscous, polymerized silicate magma and supercritical aqueous fluid (e.g. Dharmapriya et al., 2021). The pegmatite formation process is a continuous physico-chemical transitional process, progressing from magmatic to hydrothermal temperatures and pressures, making it challenging to precisely determine which hypothesis is involved.

Using experimental data, stable isotopic data, fluid inclusion analysis, and thermal modelling, researchers have attempted to speculate on the pressure-temperature (P-T) evolution of pegmatites (e.g., Jahns and Burnham, 1969; Babu, 1969; Taylor et al., 1979; Kosukhin et al., 1984; Simmons et al., 1987; Konovalenko, 1988; Chakoumakos and Lumpkin, 1990; Nabelek, 2003a,b; London et al., 2012). Results indicate that the crystallization temperature conditions range from 700 to 300 °C at 2 to 5 kbar, with a rapid crystallization process compared to typical plutonic rocks.

Pegmatites are classified based on depth of emplacement, P-T conditions, and geochemical signatures (Ginsburg et al., 1979; Černý, 1991; Zargosky et al., 1999; Wise, 1999; Pezzotta, 2001; Černý and Erict, 2005; Martin and De Vito, 2005; Fuchsloch et al., 2017; Beskin et al., 2018). Geochemical classification, mainly based on rare earth element concentration, has improved over the last thirty years but remains somewhat complicated. Overlapping geochemical signatures among different pegmatites can occur due to the complexity of the parent magma (Martin and De Vito, 2005). Additionally, the geochemical signatures of original pegmatitic magma may be overprinted by contamination from surrounding country rocks. The classification schemes also do not encompass the most dominant ceramic/barren pegmatites, constituting over 99% of all known pegmatite bodies worldwide (London, 2008).

Studies reveal that granitic pegmatites predominantly form in tectonically active settings, particularly during supercontinent cycles and associated collisional or extensional events. The observed correlation between pegmatite ages and supercontinent assembly underscores the influence of large-scale tectonic processes on pegmatite formation. Despite significant advances, there is a need for further research to elucidate the roles of specific tectonic processes, such as slab rollback and mantle plumes, in pegmatite genesis. Future investigations should integrate diverse geological data to refine models and enhance predictions of pegmatite occurrences and their economic potential worldwide.

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Figure and Table captions

Figure captions

Figure 01. Internal structure of zoned pegmatites in a schematic horizontal section (modified after Černý 1991a).

Figure 02. A schematic diagram showing the crystallization of granitic magma and withdrawn residual silicate liquid at increasing degrees of consolidation of a melt initially containing 0.2 wt. % H_2O (Modifiedafter Jahns and Burnham, 1969).

Figure 03. *P-T* field of environment hosting pegmatite population of the Abyssal (AB), Muscovite (MS), and Rare element (RE) Aluminosilicate boundaries from Robie and Hemingway (1984), cordierite in metapelites after Schreyer and Seifert (1969), Spodumene-petalite from London (1984), granite liquidus-solidus after Jahns (1982), Geothermal gradients from Winkler (1976,1967) (modified after Černý, 1991).

Figure 04. Classification of ten pegmatitiee phases (residual pegmatites) based on the thermal evolution of pegmatite melt after Fersman (1930). The right column shows the lineage of non-contaminated pegmatite subdividing the 10 thermal phases into 18 pegmatite types modified after Fersman (1967). The Figure modified after Müller et al., 2022)

Figure 6. A schematic diagram showing a across-section through a pegmatite with gem pockets (modified after Petrel al., 2012). Labeling after Petrel al. (2012). Different mineral assemblages on the hanging-wall part (large crystals of smoky quartz I up to 25 cm long + albite I) and the footwall part of the pocket (large corroded crystals of K-feldspar up to 30 cm in size muscovite II + subparallel crystals of smoky quartz II + tourmaline).

Figure 07. Age distribution patterns of significant LCT (Lithium-Cesium-Tantalum) pegmatite fields and associated granites (Pe and Gr) based on data from Condie et al. (2009) and McCauley and Bradley (2014). The age distributions of major LCT pegmatite fields are derived from Tkachev (2011). Colored bars indicating the estimated assembly periods of supercontinents. The figure is modified after Dittrich et al. (2019).

Figure 08. Map of the world showing the geological distribution of 127 pegmatite provinces and districts, distinguishing between LCT (Lithium-Cesium-Tantalum) and NYF (Niobium-Yttrium-Fluorine) suites, as well as some with hybrid affiliations. The map of orogenic belts is based on Utsunomiya et al. (2007) and Rino et al. (2008). Legend: 1. Bighorn Mt. (WY, USA); 2. Copper Mt. (WY, USA); 3. South Pass (WY, USA); 4. Black Hills (SD, USA); 5. Haystack Range (WY, USA); 6. Routt Plutonic Suite (CO, USA); 7. Trout Creek Pass (CO, USA); 8. White Picacho (AZ, USA); 9. Mohave Co (AZ, USA); 10. Laramie Mt (WY, USA); 11. Berthoud Plutonic Suite (CO, USA); 12. Burro Mtn (NM, USA); 13. Adirondack Highlands (NY, USA); 14. North New Mexico fields (USA); 15. Rockford (AL, USA); 16. Kings Mountain (NC, USA); 17. Spruce Pine (NC, USA); 18. Amelia (VA, USA); 19. New England districts (USA); 20. Brazil Lake (Nova Scotia, Canada); 21. Southern California (USA); 22. Cat Lake – Winnipeg River (Manitoba, Canada); 23. Wekusko Lake (Manitoba, Canada); 24. Yellowknife Basin (NW Territories, Canada); 25. NW Ontario fields (Canada); 26. Superior Lake (ON, Canada); 27. Lac Simard (QC, Canada); 28. Preissac – Lacorne (QC, Canada); 29. Birch Portage – Hanson Lake (SK, Canada); 30. Granville Lake (Manitoba, Canada); 31. SW Grenville southern fields (ON, Canada); 32. SW Grenville northern fields (ON, Canada); 33. Mt Laurier and Gatineau fields (QC, Canada); 34. Lac Turgeon – Johan Beetz (QC,

Canada); 35. Pikes Peak (CO, USA); 36. Llano-Burnet (TX, USA); 37-38. Sveconorwegian Province in S Norway and SW Sweden (Evje-Iveland, Froland, Glamsland, Arendal, Søndeled, Kragerø, Tørdal, Østfold-Halland); 39. Uttö-Mysingen (Sweden); 40. Varuträsk (Sweden); 41. Bothnian Basin (Sweden); 42. Falun (Central Sweden); 43. Kemiö – Orijärvi (Finland); 44. Eräjärvi (Finland); 45. Seinäjoki (Finland); 46. Ladoga Lake (Russia-Finland); 47. Chupa-Ijona (Karelia, Russia); 48. Keivy Massif, Kola Peninsula (Russia); 49. Leinster (Ireland); 50. El Muerto pegmatites (Oaxacan Complex, Southern Mexico); 51. Borborema Province (RGN, PB, Brazil); 52. Eastern Brazilian Province (MG, BA, ES, Brazil); 53. Sta Maria de Itabira (MG, Brazil); 54. Damara Province (Namibia); 55. Namaqualand (South Africa); 56. Kaapvaal (South Africa); 57. Natal districts (South Africa); 58. Panpean Pegmatite Province; 59. SW Nigeria Province (Ago-Iwoye, Keffi, Nassarawa, Komu, Wamba districts); 60. Giraul (SW Angola); 61. Caxito (NW Angola); 62. Bikita (Zimbabwe); 63. Alto Ligonha (Mozambique); 64. Ruanda; 65. Burundi; 66. Kobokobo, Kamituga area, South Kivu (Congo); 67. Kapiri Mposhi (Zambia); 68. Lundazi (Zambia); 69. Choma-Kalomo (Zambia) and Kamativi (Zimbabwe); 70. Itremo (Madagascar); 71. Southeastern Desert Province (Egypt); 72. Manono-Kittolo (Shaba, Congo); 73. Pilbara (WA, Australia); 74. Lake Moore - Dalgaranda (WA); 75. King Leopold (WA); 76. Greenbushes (WA); 77. Coolgardie – Norseman (WA); 78. Mukinbudin (WA); 79. Gascoyne (WA); 80. Mt. Isa (QNL); 81. Olary and Broken Hill, Curnamona Province (SA-NSW); 82. Bihar Mica Belt (India); 83. Nellore Mica Belt (India); 84. Tamil Nadu Belt (India); 85. Bastar-Malkagiri Belt (India); 86. Rajasthan Belt (India); 87. Nuuk Region (Greenland); 88. Gardar Province (Greenland); 89. Volta Grande, Sao Joao de Rei (Brazil); 90. Strange Lake (Labrador, Canada); 91. Alakha (Russia); 92. Kolmozero-Voronya (Kola Peninsula, Russia); 93. Yenisei Ridge (Russia); 94. Highland Complex (Sri Lanka); 95. Ghaha-Cote d'Ivoire; 96. Bohemian-

Moldanubicum Belt; 97. Taimyr Fold Belt (Russia); 98. Altai Belt (China, Kazakhstan, Russia); 99. Mongol-Okhotsk Fold Belt; 100. East Sayan Mt. (Siberia, Russia); 101. Southern Tuva (Kamar-Daba Fold Belt); 102. East Transbaikalia (Russia); 103. NW Baikal (Russia); 104. Mama-Chuya, North Baikal Highland (Russia); 105. Iberian Peninsula; 106. Creus Cap (Spain); 107. Hagendorf-Pleystein (Babaria, Germany); 108. Paranesti (NE Greece); 109. Koralpe (Austria); 110. Afghanistan-Pakistan Belt; 111. Little Nahanni (NW Territories, Canada); 112. Jiajika, Kangdi (Sichuan Province, China); 113. Nanping (China); 114. Cattlin Creek – Cocanarup, Ravensthorpe (WA); 115. Tak (Thailand); 116. Phuket (Thailand); 117. Lao Cai and Phu Tho (Northern Vietnam); 118. Kenticha (Ethiopia); 119. Nimnyr Block, Central Aldan (Russia); 120. Middle Urals (Russia); 121. South Urals (Russia); 122. Southern Japan (Japan); 123. Central Alps; 124. Qiongjiagang in Tibet; 125. Heiyingshan in Xinjiang (China); 126. Jing’erquan in Xinjiang (China), 127. Chaka in Qinghai (China)

Table Caption

Table 1. Selected examples of the definition of “pegmatite” through history.

Table 2. Occurrences of accessory minerals in some selected pegmatite/ pegmatite fields in the world

Table 3. Some selected Studies on Syn and Post-Crystallization Modifications in Pegmatites

Table 4. Estimated *P*-*T* condition of selected pegmatites/pegmatite fields in the world

Table 5. Summary of existing classification schemes of granitic pegmatitie (modified after Muller et al., 2022)

Table 6. Classification of granitic pegmatites (this table is a modified version after Černý' (1991) and Černý' and Erit. 2005)

Table 7. Typical examples of accessory mineral assemblages in pegmatite groups (after Wise et al, 2022)

Table 8. Major gem-pegmatite districts and notable gem minerals

Table 1. Selected examples of the definition of “pegmatite” through history.

Publication	**Definition**
Bastin(1911) (Geology report)	*“The granite-pegmatites, in which are. found feldspar, quartz, mica, and gem minerals, are composed of the same mineral constituents as the ordinary granites of the State, and differ from these principally in their greater coarseness and in their very uneven texture.”*
Landes (1933) (Original research article)	The word 'pegmatite" was coined originally by Haüy in the early years of the 19th century as a descriptive term applied to rocks with a graphic texture. In 1849 "pegmatite" was used by Delesse (Williams,1885) to designate a very coarse granite and gradually the use of the term was extended to include all types of exceptionally coarse plutonic rocks whether graphic or not.
Jahns (1953) (Research article)	“The term pegmatite is applied by geologists mainly to those intrusive igneous rocks that are holocrystalline and at least in part very coarse-grained. Marked irregularities in grain size are characteristic, and in some masses of pegmatite, the coarsest crystals are enormous.”
Gillespie and Styles (1999) (a rock classification scheme)	*“The term pegmatite should be used only for veins comprised of coarse or very coarse crystals. Such rocks should be named by listing the component minerals as qualifiers in front of the term pegmatite, for example biotite-quartz-feldspar pegmatite.”*
Best (2003) (Textbook)	*“Pegmatite: Rocks of highly variable grain size in which individual crystals can be as much as several m. Usually refers to a textural variety of leucocratic granite.”*
London (2008) (Textbook)	*“Pegmatites are essentially igneous rock, mostly of granitic composition, that is distinguished from other igneous rocks by its extremely coarse but variable grain-size, or by an abundance of crystals with skeletal, graphic, or other strongly directional growth habits.”*
London et al. (2012) (Review article)	*“Pegmatites are texturally distinct variants of the more common and more voluminous plutonic igneous rocks, including gabbros, granites, syenites, etc. Whereas common plutonic bodies tend to be mineralogically and texturally homogeneous throughout large volumes of rock, pegmatites are precisely the opposite. Exceedingly coarse crystal size is a hallmark of pegmatites for most geoscientists but gigantic crystal size is not the sole or even a necessary defining factor.”*
Simmons et al. (2024) (Review article)	*“Pegmatites are among the most intriguing of all the igneous rock types. They form from the crystallization of an igneous melt and show very coarse-grained to gigantic-sized textures in intrusive igneous rocks, with crystals generally more than 2.5 cm.”*

Table 2. Occurrences of accessory minerals in some selected pegmatite/pegmatite fields in the world

pegmatite/ pegmatite field	**Major mineral**	**Minor/Accessory minerals**	**Age (Ma)**	**References**
The Tanco pegmatite at Bernic Lake, Manitoba, Canada	Albite, microcline, quartz, mica	Hafnium zircon, tantalite, lepidolite, pollucite, petalite, amblygonite	~ 2631	Černý and Siivola 1980; Černý et al. 2005; Brown 2001; Camacho et al., 2012; Brown et al., 2017London 2024;
Greenbushes, Western Australia	Albite, Spodumene, quartz, biotite,	Tourmaline, Ta-ilmenite, Ta-rutile, cassiterite, apatite	~2530	Partington et al., 1995; Klementová, Rieder, 2004; Phelps-Barber et al., 2022
Wodgina, Western Australia	Albite, quartz, muscovite	Columbite, cassiterite, petalite, beryl, wodginite	~2800	Tindle et al., 1998; Sweetapple and Collins, 2002; Richter et al., 2014; Jacobson, 2021
Tochka deposit, Karagoin–Saryozek zone in East Kazakhstan	Microcline, albite, quartz	Zircon with disseminated U phase (uraninite), Rutile, Ni phase (bunsenite?), cassiterite, tantalite, columbite, native tin, pyrolusite	~292	Zimanovskaya et la., 2022
Bikita, Zimbabwe	Quartz, albite, Mica	Tourmaline, zinnwaldite, garnet, tourmaline and epidote, beryl, petalite, tantalite, amblygonite	~2600	Cerny et al., 2003;Dittrich et al., 2019; Gogodo, 2021; Chagondah et al., 2024
Manono-Kitotolo, Democratic Republic of Congo	Quartz, muscovite, microcline and albite	Beryl, spodumene, columbite, tantalite, cassiterite, thoreaulite, wodginite and tapiolite, fluorite, zircon, thoreaulite, arsenopyrite, pyrite, ilmenite, autunite	~930	Thoreau, 1950; Bassot and Morio, 1989; Melcher et al., 2008; Dewaele et al., 2016
Alto Ligonha, Mozambique	Quartz, albite K-feldspar, muscovite	Zinnwaldite, lepidolite, polylithionite, beryl, garnet, tourmaline, cassiterite, gahnite, columbite, tantalite, rutile, some uranium minerals	~480	Gomes et al., 2009; Neiva et al., 2013;Cronwright, 2014; Gomes and Neiva, 2022; Altieri et al., 2024
Mina Gerais, Brazil	Quartz, feldspar, mica	Tourmaline, aquamarine, topaz, beryl, bergslagite, herderite, lithiophosphate, lithiotantite, kosnarite	~580 and ~520	Murata et al., 1959; Federico et al., 1998; Scholz et al., 2011: De Mello and Bilal, 2012; Menezes et al., 2016; Aranha et al., 2018

Jiajika Pegmatite, China	Microcline, albite, muscovite, quartz	Lepidolite, spodumene, beryl, columbite-group minerals; cassiterite, tourmaline, garne	~200	Jiankang et al., 2013; Wang et al., 2021; Zhu et al., 2023; Gao et al., 2024
Fregeneda–Almendra pegmatitic field in Spain and Portugal	quartz, K-feldspar, albite, muscovite,	Tourmaline, andalusite, garnet, Fe-Mn phosphates, montebrasite, petalite, spodumene, lepidolite	~300	Roda et al., 2007; Roda-Robles et al., 2010; 2023

Table 3. Some selected Studies on syn and post-crystallization modifications in pegmatites

Author/s and year	Locality or pegmatite field	Key interpretation related to modifications in Pegmatites
Chadwick (1958)	New England, USA	The various pegmatites developed by several mechanisms. Furthermore, individual pegmatites commonly exhibit evidence for more than one process.
Černý (1972)	Tanco pegmatite	Late hydrothermal activity produced a series of secondary minerals in the spodumene-rich zones of the Tanco pegmatite
Kinnaird (1985)	Jurassic alkaline anorogenic granitic ring complexes in central Nigeria	The contribution of post-crystallization acid metasomatic fluid to the mineralization of ore minerals such as monazite, zircon, and ilmenite, followed by cassiterite, wolframite, columbite, siderite, and rutile.
Shearer et al. (1986)	Black Hills, South Dakota. USA	Migration of pegmatite-derived fluids out of pegmatite bodies and into the surrounding country rock can form extensive alteration aureoles.
Morgan and London (1987)	Tanco pegmatite, Bernic Lake, Manitoba	The occurrences of metasomatic alteration of pegmatite under the presence of fluids
Linnen and Williams-Jones (1994)	Nong Sua, Thailand	The two dominant stages of Sn-W-Ta-Nb mineralization are: (1) the early stage, which forms directly from the magmatic melt, and (2) the later stage, which involves the formation of W-rich minerals through Fe-Mn-rich fluids.
Uher et al. (1998)	Prasivá granitic pegmatites, Slovakia	The hydrothermal overprint of the Prasivá pegmatites was caused by external, low-temperature CO_2-rich aqueous solutions of magmatic-metamorphic origin
Thomas et al. (2003)	Ehrenfriedersdorf Complex, Germany	Detailed analyses of melt and fluid inclusions combined with an electron-microprobe survey of boron-bearing minerals reveal the evolution of boron in a highly evolved peraluminous granite-pegmatite complex and the associated high- and medium-temperature ore-forming hydrothermal fluids.
Palinkaš et al. (2014)	Mina do Santino and the Jacú pegmatites in the Borborema Pegmatite Province in northeastern Brazil	After minerals such as feldspars and beryl were deposited under magmatic conditions, they interacted with hydrothermal fluids, becoming unstable. This interaction led to the formation of a new hydrothermal mineral paragenesis, which included euclase, bertrandite, kaolinite, and quartz, that overprinted the earlier magmatic minerals
Rao et al. (2017)	The Nanping No. 31 pegmatite (Fujian Province, southeastern China)	The study suggests that the re-enrichment of Li during the post-magmatic stage likely increased the solubility of Ta in the hydrothermal fluids of the granitic pegmatite. This implies that the hydrothermal alteration processes not only affect the distribution of Li but also influence the behavior of other rare elements like Ta in the pegmatite system.
Kaeter et al. (2018)	Southeast Ireland	The study reveals that hydrothermal fluids significantly impact the mineralogical evolution during the magmatic-hydrothermal transition. Ta and Nb are fractionated differently during the hydrothermal process. Ta is preferentially concentrated and redeposited by halogen-rich aqueous fluids,

		leading to Ta-enriched mineralization. This fractionation is observed in both the magmatic and hydrothermal stages, with hydrothermal fluids playing a crucial role in the remobilization and redeposition of Ta.
Wu et al. (2018)	Shangdan domain of the North Qinling Orogenic Belt, China	Hydrothermal alteration of primary uraninite involved multiple stages of fluid-assisted metasomatism.
Yin et al. (2020)	The Bailongshan pegmatite district in south Xinjiang Province, NW China,	Hydrothermal fluids caused the development of Cs- or Na-enriched veinlets in beryl and Mn- and Ta-enriched veinlets in columbite-group minerals. These alterations reflect the complex chemical interactions between the pegmatitic magma and the hydrothermal fluids, which were rich in Li, F, Cs, and Ta.
Ballouard et al. (2020a)	Orange River pegmatite belt, Namaqualand, South Africa	The study emphasizes the complex interplay between magmatic and hydrothermal processes in the formation of pegmatite-related rare metal deposits. It indicates the alteration of primary minerals in the pegmatites through hydrothermal fluids in several stages.
Bhandari et al. (2023)	Yamrang Pegmatite, Eastern Nepal	Alkali-enriched fluids exsolved from the pegmatite, leading to the formation of miarolitic cavities and the crystallization of aquamarine. This process resulted in the development of hydrothermal beryl, characterized by its sodic to Li-Cs composition, tabular to acicular habit, and weak compositional zoning. The hydrothermal stage significantly influenced the pegmatite's mineralogy and texture.
Xu et al. (2024)	Songpan–Ganzê orogenic belt, Eastern Tibet	The hydrothermal alteration, particularly metasomatism, leads to the release of elements like Ta, Nb, Sn, and W from primary minerals (spodumene, alkali feldspar, and columbite-group minerals) into the reactive media (melt/fluid). This redistribution results in the depletion of these elements in early-formed minerals but sets the stage for their subsequent mineralization.

Table 4. Estimated *P-T* condition of selected pegmatites/pegmatite fields in the world

Reference	**Pegmatite/ pegmatite field**	***Used methods***	***Crystallization conditions***	
			P	***T***
Brotzen (1959)	Stockholm, Sweden	Considering mineral solid solutions' stability	-	600°C
Jahns and Burnham (1969)		Experimental studies	**-**	Hydrous minimum melt temperatures of about 600 °C
Babu (1969)	Nellore Mica Belt, Andhra Pradesh, India	Mainly based on mineral compositions	-	760 – 300 °C Crystallization of: -Tourmaline 760 to 450 °C - Beryl is 500 °C to 400 °C - Muscovite 500 to 435 °C
Taylor et al. (1979)	San Diego Co., California	Considering oxygen isotope fractionations between minerals	-	730 to 700 °C - Aplite zone in that pegmatite occurring around 565 °C - Gem pocket formation in the relatively narrow temperature range of about 40 °C (approximately 565–525 °C)
Kosukhin et al. (1984)	Russia	Fluid and melt inclusion analysis	3.0 – 0.5 kbar	Granite pegmatites which contain mineralized cavities (miaroles) crystallize at 750 - 515°C
Simmons et al. (1987)	South Platte, Colorado	Two-feldspar thermometry		550–500 °C
Bakumenko and Konovalenko (1988)	SW Pamirs	Fluid and melt inclusion analysis		Crystallization of miarolitic quartz 510 – 505 °C
Chakoumakos and Lumpkin (1990)	Harding pegmatite. Taos county,	I) the experimentally determined	3.3 -3.5 kbar	650°C

	New Mexico, USA	liquidus, solidus and phase assemblages for a bulk sample of the pegmatite, 2) isochores for CO_2-H_2O-NaCl fluid inclusions in quartz and beryl		
Morgan and London, (1999)		Two-feldspar thermometry.		
Frindt et al. (2002)	Gross Spitzkoppe Granite Stock, Namibia	Fluid inclusion analysis	0.9 kbar	Crystallization and fluid entrapment during cooling from high temperatures (around 550°C) down to lower temperatures around 170°C
Kontak et al. (2002)	Peggys Cove, Nova Scotia, Canada	Fluid inclusion analysis	3–3.5 kbar	600–650°C
Sirbescu and Nabelek, (2003a)	Black Hills, South Dakota	Fluid inclusion analysis	2.7 kbar	~400–350 °C
Li et al. (2013)	Jiashan–Xiangshui Fault, china	Ti in zircon thermometry		730–840 °C
Fei et al. (2021)	Songpan-Garze Fold Belt, southwestern Sichuan province, China	Fluid inclusion analysis	>4.7 kbar	741 to 658 °C
Knoll et al. (2023)	Austroalpine Unit Pegmatite Province (Eastern	Studying of metamorphic history of surrounding metamorphic rocks	3 – 4 kbar	500–570 °C

	European Alps)			
Wang et al. (2023)	Lhozhag, eastern Himalaya	Studying of metamorphic history of surrounding metamorphic rocks	3.1	720 – 650°C

Table 5. Summary of existing classifications schemes of granitic pegmatites (modified after Müller et al., 2023)

Author/s	**1st order classification criterion**	**2nd order classification criterion**	**3rdorder classification criterion**	**Evolution**
Niggli (1920)	Geochemistry	Mineralogy	-	Not well constrained
Fersman (1930)	Temperature	Geochemistry	Mineralogy	Temperature along not sufficient
Landes (1933)	Geochemistry	Structure	Mineralogy	Too simple, not well-constrained
Bjørlykee (1937)	Geochemistry, genesis	Mineralogy	Mineralogy	Limited to rare metal pegmatites of the NYF family
Vlasov (1952, 1961)	Structure, mineralogy	Mineralogy		Too simple, not well-constrained
Ginsburg and Rodionov (1960); Ginsburg (1979; 1984)	Emplacement depth	Structure (presence or absence of miarolitic cavities)	Geochemistry	It presupposes that the metamorphic grade of the wall rock corresponds to the P-T conditions at the time of pegmatite emplacement
Schneiderhöhn (1961)	Genisis (anatectic vc. Magmatic)	Geochemistry	Geochemistry	Too simple
Černý (1991a) and Černý and Erict (2005)	Mixture of emplacement depth structure and mineralogy	Geochemistry	Mineralogy	A mixture of different types of classification criteria in the 1st order
Wise (1999)	Mineralogy	Mineralogy	Mineralogy	Classification of NYF-type pegmatites only
Pezzotta (2001)	Mixture of emplacement depth structure and mineralogy	Mineralogy, geochemistry	Mineralogy	Limited to pegmatites occurring in Madagascar
Zagorsky et al. (1999, 2003)	Emplacement depth	Mineralogy (structure)	Mineralogy (structure)	Emplacement depth is commonly uncertain
Dill (2016)	Mixture of emplacement style, deformation, host	Structure	Chemistry	The distinction of different pegmatite types

	rock type, chemical and mineralogical features			(1st order) is difficult to draw in several classes
Wise et al. (2022)	Mineralogy	Geochemistry	Genesis (magmatic vs. anatectic)	Fall short in accommodating all variations observed in nature, particularly in hybrid or mixed-origin pegmatites

Table 6. Classification of granitic pegmaitte (the table is a modified version after Černý’, 1991 and Simmons and Webber, 2008)

Class	**Family**	**Pegmatite types**	**Pegmatite sub types**	**Geochemical signature**	**Metamorphic environment**	**Relation to granite**	**Structural features**	**Example**
Abyssal				U, Th, Zr, Nb, Ti, Y, REE, Mo poor (to moderate) mineralization	(Upper amphibolite to) low- to high-P granulite facies ~4-9 kb ~700-800°C	None (segregations of anatectic leucosome)	Conformable to mobilized cross-cutting veins	Rae and Hearne Provinces, Sask. (Tremblay, 1978); Aldan And Anabar Shields, Siberia (Bushev and Koplus, 1980)
Muscovite				Li, Be, Y, REE, Ti, U, Th, Nb>Ta poor (to moderate) mineralization, micas and ceramic minerals	High-P, Barrovian amphibolite facies (kyanite-sillimanite) ~5-8 kb ~650-580°C	None (anatectic bodies) to marginal and exterior	Quasi-conformable to cross-cutting	White Sea region, USSR (Gorlov, 1975); Appalachian Province (Jahns *et al.*, 1952); Rajahstan, India (Shmakin, 1976)
Rare – Element	LCT Poor to abundant mineralization,	Beryl	Beryl – columbite	Be, Nb>< Ta (± Sn, B) Beryl-(columbite -				

	gem stock industrial minerals			tantalite)				
			Beryl - columbite phosphate	Be, Nb>< Ta, P (Li, F ± Sn, B) Beryl, columbite-tantalite, triplite, triphylite				
		Complex	Spodumene	Li, Rb, Cs, Be, Ta ><Nb (Sn, P, F ± B) spodumene (amblygonite), beryl (lepidolite), tantalite (pollucite)	Low-P, Abukuma amphibolite to upper greeenschistfacies (andalusite-sillimanite) ~2-4 kb ~650-500°C	(Interior to marginal to) exterior	Quasi-conformable to cross-cutting	Yellowknife field, NWT (Meintzer, 1987); Cat Lake-Winnipeg River field, Manitoba (Černý*et al.*, 1981)
			Petalite	Li, Rb, Cs, Be, Ta >Nb (Sn, Ga, P, F ± B) petalite (amblygonite) tantalite, beryl				
			Lepidolite	F, Li, Rb, Cs, Be Ta >Nb (Sn, P ± B) lepidolite microlite beryl,				

				topaz				
			Amblygonite	P, F, Li, Rb, Cs Be, Ta >Nb (Sn ± B) amblygonite (lepidolite), beryl (pollucite), tantalite				
		Albite spodumene		Li (Sn, Be, Ta ><Nb ± B				
		Albite		Ta ><Nb, Be (Li ± Sn, B)				
	NYF	REE	Allanite - monazite					
			Gadolinite					
Miarolitic				Be, Y, REE, Ti, U, Th, Zr, Nb> Ta, F Poor mineralization, gemstock	Shallow to sub-volcanic ~1-2 kb	Interior to marginal	Interior pods and cross-cutting dikes	Pikes Peak, Colorado (Foord, 1982); Korosten pluton, Ukraine (Lazarenko*et al*., 1973)

Table 7. Typical examples of accessory mineral assemblages in pegmatite groups (after Wise et al, 2022)

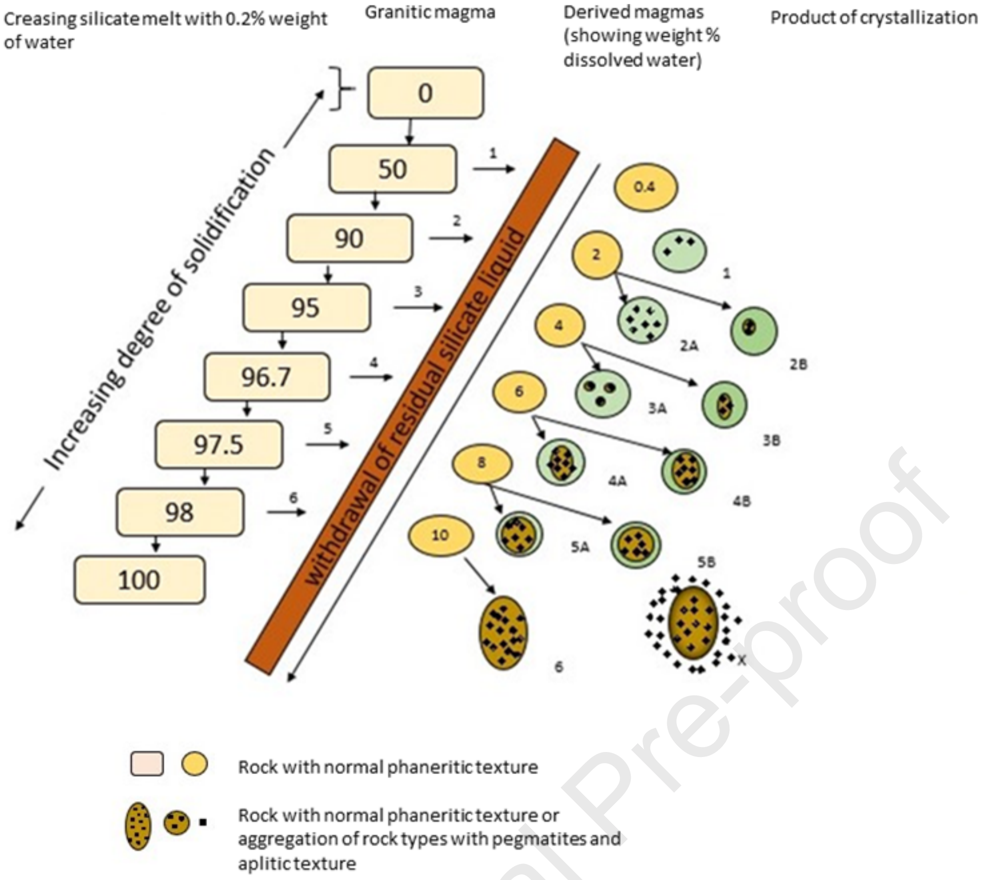
Group 1	Group 2	Group 3
Beryl ±columbite ± triphylite ± graftonite	Magnetite ± epidote ± titanite	Andalusite ± sillimanite ± corundum
Beryl ± amblygonite ± columbite	Uraninite ± magnetite	Kyanite ± rutile
Amblygonite ± spodumene	Fayalite ± magnetite	Cordierite ± beryl
Spodumene ±petalite	Arfvedsonite-riebeckite ± fayalite	Sapphirine ± sillimanite
Spodumene ±elbaite	Arfvedsonite-riebeckite ± aegirine ± fluorite	Chrysoberyl ± beryl ± schorl
Spodumene ± lepidolite ± elbaite	Allanite ± gadolinite	Dumortierite ± grandidierite ± borasilite
Spodumene ± albite	Allanite ± monazite ± euxenite	Dumortierite ± schorl
Elbite ± danburite ± hambergite ± zeolites	Allanite + beryl ± columbite	
	Gadolinite ± euxenite ± fergusonite	
	Beryl ± phenakite	
	Microcline (var. amazonite ± fluorite	
	Microcline (var. amazonite ± topaz ± fluorite	

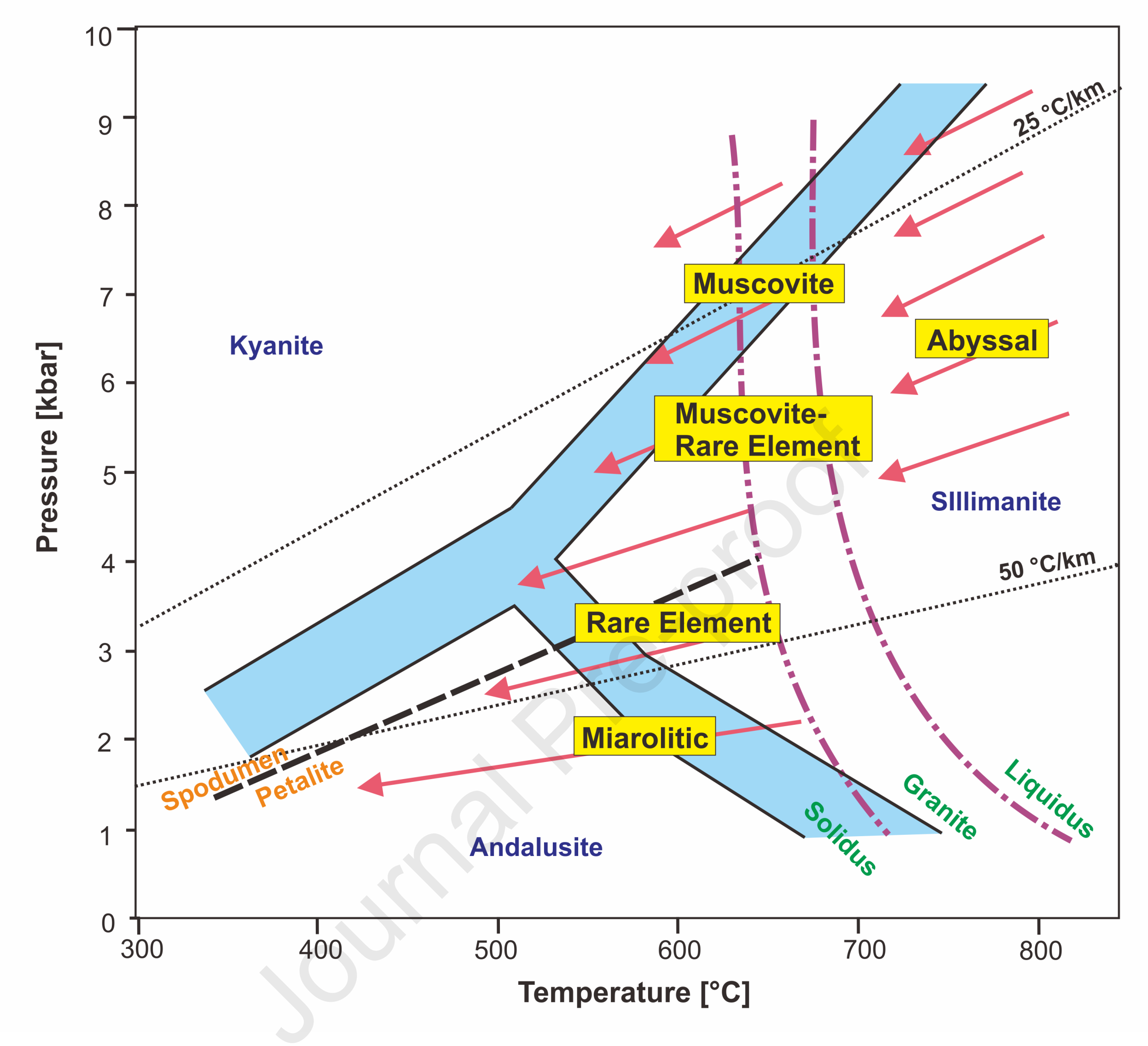
Table 8. Major gem-pegmatite districts and notable gem minerals in the world

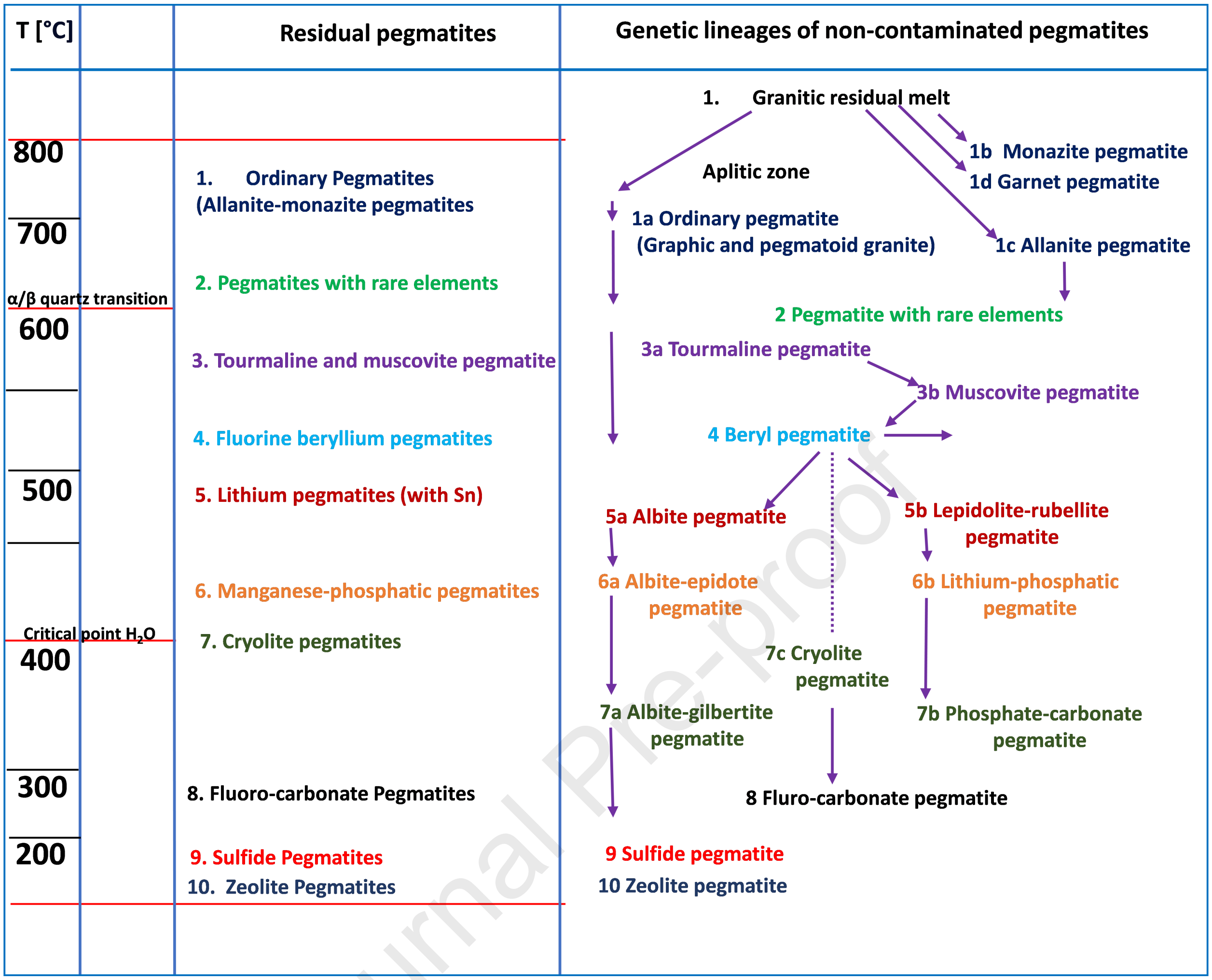
Pegmatite/pegmatite field and the country	**Notable gem minerals**
Minas Gerais in Brazil	Beryl (aquamarine, morganite, heliodor), topaz, tourmaline, spodumene (kunzite, hiddenite), brazilianite, euclase, quartz (rose, smoky, citrine), amblygonite–montebrasite
Paraíba, Rio Grande do Norte in Brazil	Tourmaline (Paraíba variety), beryl (aquamarine, morganite, heliodor), euclase, quartz (rose, smoky, citrine), spessartine
Central and Eastern Zambia	Tourmaline, beryl (emerald), spessartine
Katanga, Nord Kivu, D. R. Congo:	Tourmaline
Coast in Kenya	Tourmaline
Erongo in Namibia	Tourmaline, beryl (aquamarine), topaz, jeremejevite
Edo, Kaduna, Kwara, Nassarawa, Oyo, Plateau in Nigeria	Tourmaline, chrysoberyl, topaz, beryl, phenakite, spodumene, spessartine
Mashonaland West, Zimbabwe	Beryl (aquamarine), topaz, tourmaline
Nampula, Tete, Zambezia in Mozambique	Tourmaline, spessartine, beryl (aquamarine), topaz, spodumene (kunzite), quartz (rose)
Matale, Awissawella, Balangoda, Meetiyagoda in Sri Lanka	Saphire, amazonite, topaz, moonstones
Arusha, Morogoro, Tanga in Tanzania:	Tourmaline, spessartine
Antananarivo, Fianarantsoa, Toamasina in Madagascar	Tourmaline, beryl (aquamarine), spodumene (kunzite), londonite, pezzottaite, hambergite, orthoclase/ sanidine, quartz (rose), spessartine, danburite, phenakite
California, Mesa Grande, Pala, Ramona in United States	Tourmaline, beryl (aquamarine, morganite), spessartine, spodumene (kunzite)
Colorado Collegiate Range in the United States	Beryl (aquamarine), phenakite
New England, Maine, New Hampshire, and Connecticut in the United States	Tourmaline, beryl (aquamarine), pollucite
Zhytomyr in Ukraine	Beryl (heliodor, aquamarine), topaz

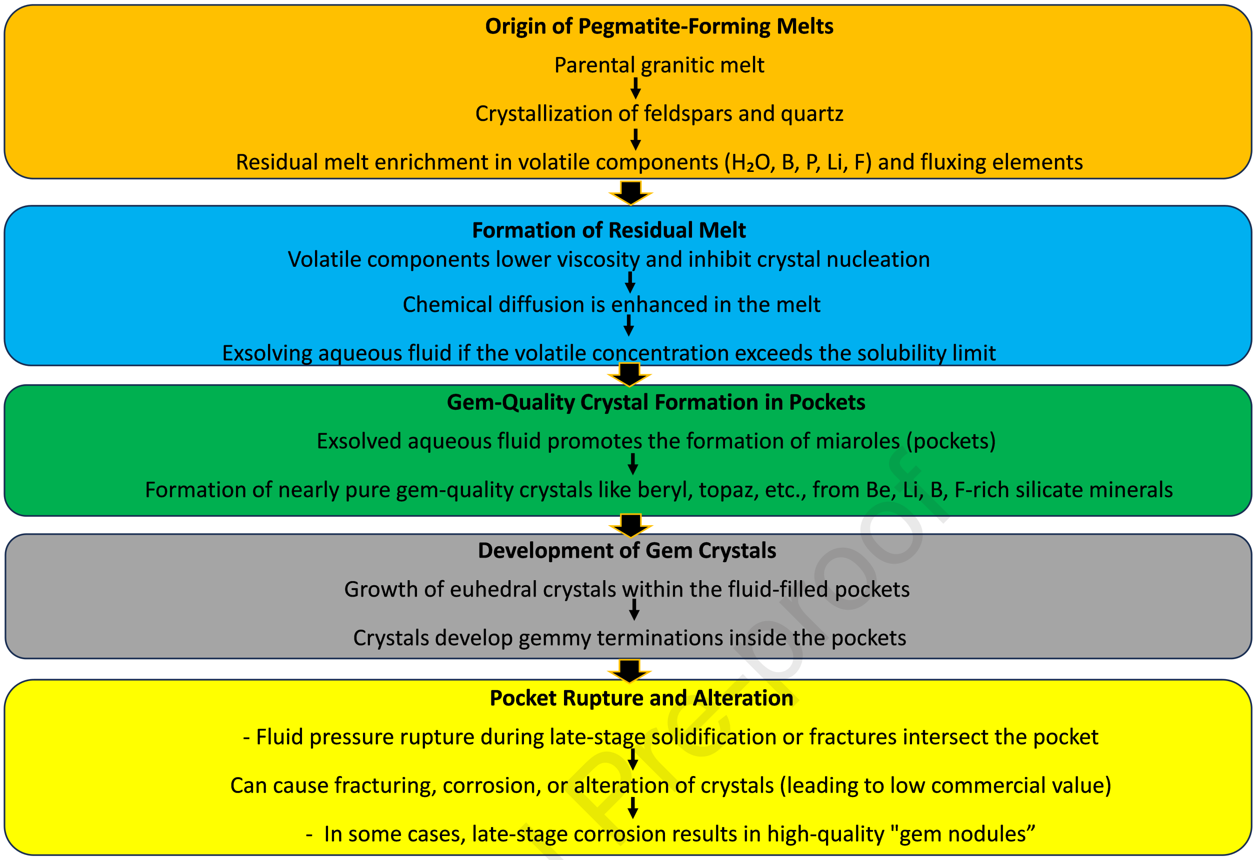
Ural Mountains in Russia	Beryl (aquamarine, emerald), topaz, chrysoberyl (alexandrite)
Zabaykalsky (Transbaikalia) in Russia	Tourmaline, danburite, pollucite, spessartine, beryl (aquamarine), topaz
Elba in Italy	Tourmaline, pollucite
Luumäki (South Karelia) in Finland	Beryl (aquamarine), tourmaline
Mogok, Momeik in Myanmar	Tourmaline, phenakite, petalite, pollucite
Yen Bai in Vietnam	Tourmaline, microcline (amazonite)
Xinjiang Uygur in China	Beryl (aquamarine), tourmaline
Yunan in China	Tourmaline, beryl (aquamarine), spessartine
Tamil Nadu in India	Beryl (aquamarine)
Kunar, Nuristan in Afghanistan	Tourmaline, spessartine, topaz, beryl (emerald, aquamarine), spodumene (kunzite)
Gilgit-Baltistan in Pakistan	Tourmaline, spessartine, beryl (aquamarine), topaz

Sources: Pezzotta (2001); Falster et al. (2002, 2005); Groat et al. (2007); Simmons(2007); Cornejo and Bartorelli (2010); Shigley et al. (2010); Clifford et al. (2011), Dharmapriya et al. (2020)


Creasing silicate melt with 0.2% weight of water
Granitic magma
Derived magmas (showing weight % dissolved water)
Product of crystallization
Increasing degree of solidification
0
50
90
95
96.7
97.5
98
100
1
2
3
4
5
6
withdrawal of residual silicate liquid
0.4
1
2
2A
2B
4
3A
3B
6
4A
4B
8
5A
5B
10
6
X
Rock with normal phaneritic texture
Rock with normal phaneritic texture or aggregation of rock types with pegmatites and aplitic texture


Pressure [kbar]
10
9
8
7
6
5
4
3
2
1
0
300
400
500
600
700
800
Temperature [°C]
Kyanite
SIllimanite
Andalusite
Muscovite
Abyssal
Muscovite-
Rare Element
Rare Element
Miarolitic
Spodumen
Petalite
Solidus
Granite
Liquidus
25 °C/km
50 °C/km


T [°C]
Residual pegmatites
Genetic lineages of non-contaminated pegmatites
800
700
α/β quartz transition
600
500
Critical point H_2O
400
300
200
1. Ordinary Pegmatites (Allanite-monazite pegmatites
2. Pegmatites with rare elements
3. Tourmaline and muscovite pegmatite
4. Fluorine beryllium pegmatites
5. Lithium pegmatites (with Sn)
6. Manganese-phosphatic pegmatites
7. Cryolite pegmatites
8. Fluoro-carbonate Pegmatites
9. Sulfide Pegmatites
10. Zeolite Pegmatites
1. Granitic residual melt
Aplitic zone
1b Monazite pegmatite
1d Garnet pegmatite
1a Ordinary pegmatite (Graphic and pegmatoid granite)
1c Allanite pegmatite
2 Pegmatite with rare elements
3a Tourmaline pegmatite
3b Muscovite pegmatite
4 Beryl pegmatite
5a Albite pegmatite
5b Lepidolite-rubellite pegmatite
6a Albite-epidote pegmatite
6b Lithium-phosphatic pegmatite
7c Cryolite pegmatite
7a Albite-gilbertite pegmatite
7b Phosphate-carbonate pegmatite
8 Fluro-carbonate pegmatite
9 Sulfide pegmatite
10 Zeolite pegmatite



Origin of Pegmatite-Forming Melts

Parental granitic melt

↓

Crystallization of feldspars and quartz

↓

Residual melt enrichment in volatile components (H_2O, B, P, Li, F) and fluxing elements

Formation of Residual Melt

Volatile components lower viscosity and inhibit crystal nucleation

↓

Chemical diffusion is enhanced in the melt

↓

Exsolving aqueous fluid if the volatile concentration exceeds the solubility limit

Gem-Quality Crystal Formation in Pockets

Exsolved aqueous fluid promotes the formation of miaroles (pockets)

↓

Formation of nearly pure gem-quality crystals like beryl, topaz, etc., from Be, Li, B, F-rich silicate minerals

Development of Gem Crystals

Growth of euhedral crystals within the fluid-filled pockets

↓

Crystals develop gemmy terminations inside the pockets

Pocket Rupture and Alteration

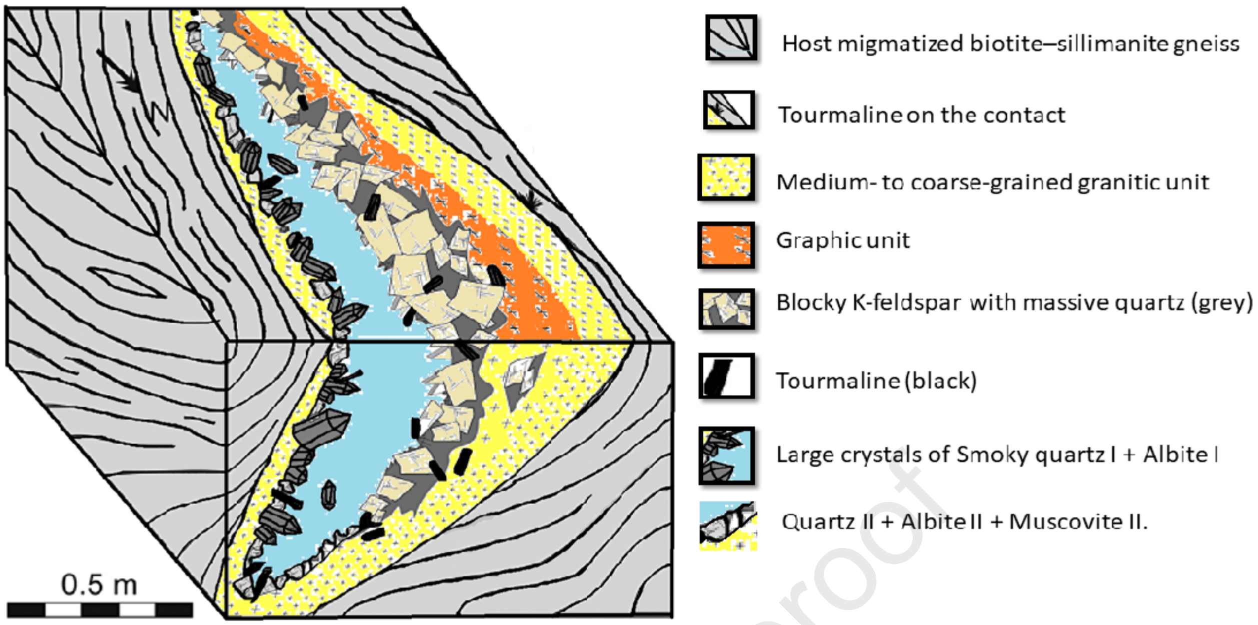
- Fluid pressure rupture during late-stage solidification or fractures intersect the pocket

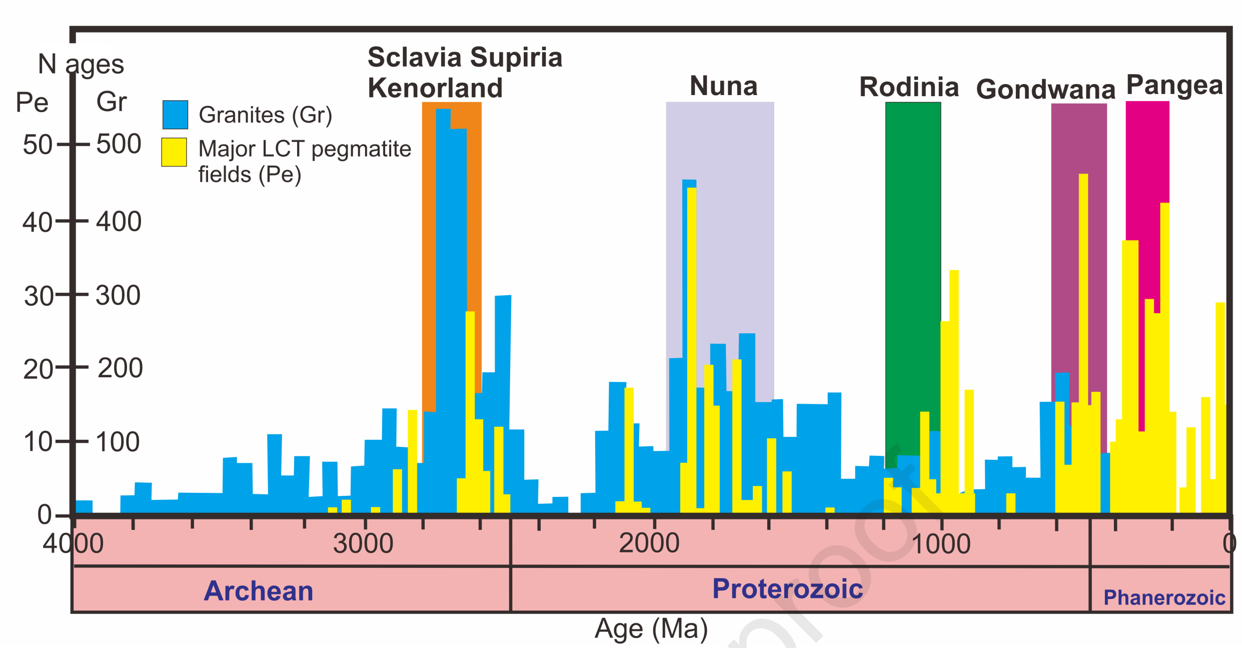
↓

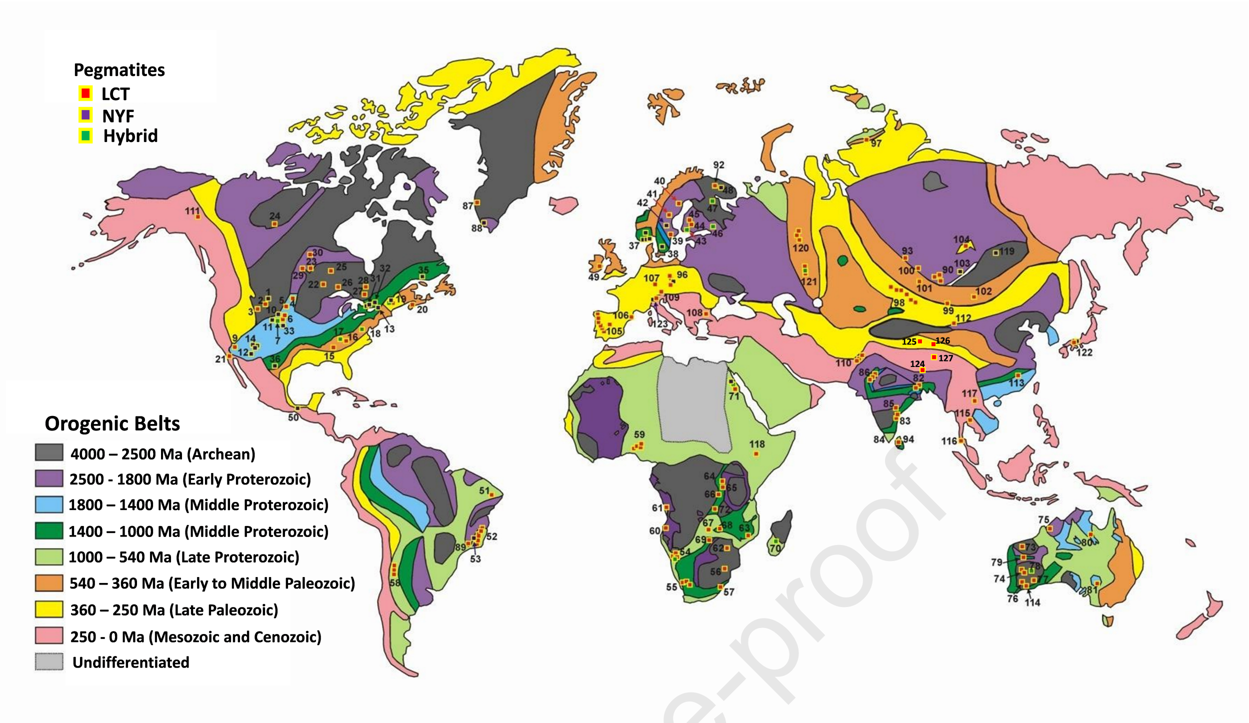
Can cause fracturing, corrosion, or alteration of crystals (leading to low commercial value)

↓

- In some cases, late-stage corrosion results in high-quality "gem nodules"


0.5 m
Host migmatized biotite–sillimanite gneiss
Tourmaline on the contact
Medium- to coarse-grained granitic unit
Graphic unit
Blocky K-feldspar with massive quartz (grey)
Tourmaline (black)
Large crystals of Smoky quartz I + Albite I
Quartz II + Albite II + Muscovite II.


N ages
Pe
Gr
50
500
40
400
30
300
20
200
10
100
0
Granites (Gr)
Major LCT pegmatite fields (Pe)
Sclavia Supiria
Kenorland
Nuna
Rodinia
Gondwana
Pangea
4000
3000
2000
1000
0
Archean
Proterozoic
Phanerozoic
Age (Ma)


Pegmatites
LCT
NYF
Hybrid
Orogenic Belts
4000 – 2500 Ma (Archean)
2500 - 1800 Ma (Early Proterozoic)
1800 – 1400 Ma (Middle Proterozoic)
1400 – 1000 Ma (Middle Proterozoic)
1000 – 540 Ma (Late Proterozoic)
540 – 360 Ma (Early to Middle Paleozoic)
360 – 250 Ma (Late Paleozoic)
250 - 0 Ma (Mesozoic and Cenozoic)
Undifferentiated
111
24
30
23
29
25
22
26
28
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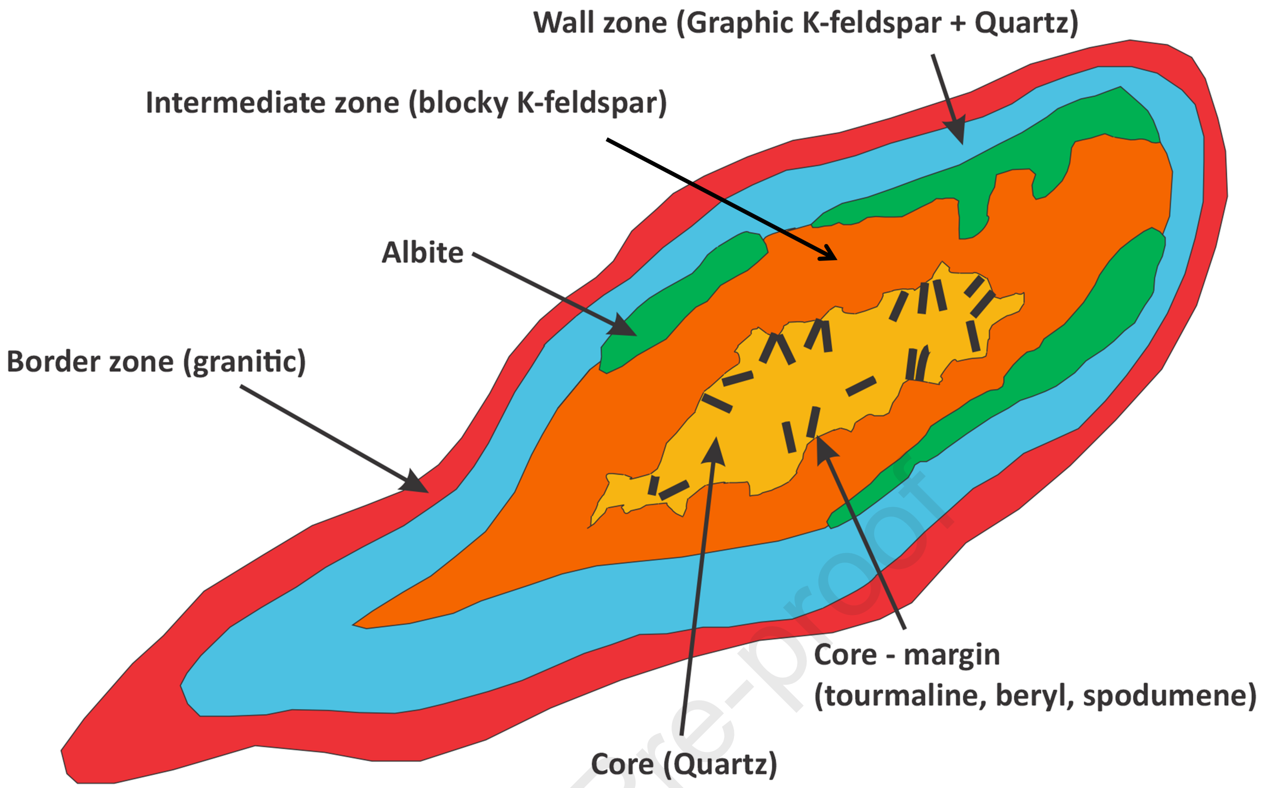
114
81


Wall zone (Graphic K-feldspar + Quartz)
Intermediate zone (blocky K-feldspar)
Albite
Border zone (granitic)
Core - margin
(tourmaline, beryl, spodumene)
Core (Quartz)

Declaration of interests

☐ The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

☒ The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:

P.L. Dharmapriya reports financial support was provided by National Research Council, Sri Lanka. If there are other authors, they declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.