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Chapter 8.2.2.

Kimberlite, carbonatite and alkaline magmatic systems: definitions, origins, and strategic mineral resources

*Sebastian Tappe [a], Charles D. Beard [b], Anouk M. Borst [c, d], Emma R. Humphreys-Williams [e],
Benjamin Walter [f], Gregory M. Yaxley [g]

[a] Institute for Mineralogy, Technical University Bergakademie Freiberg, D-09599 Freiberg, Germany
[b] Department of Earth Sciences, Utrecht University, 3584 CB Utrecht, The Netherlands
[c] Department of Earth and Environmental Sciences, KU Leuven, 3001 Heverlee, Belgium
[d] Geodynamics and Mineral Resource Unit, Royal Museum for Central Africa, 3080 Tervuren, Belgium
[e] Department of Earth Sciences, University College London, London WC1E 6BT, United Kingdom
[f] Department of Geosciences, University of Tübingen, D-72076 Tübingen, Germany
[g] Research School of Earth Sciences, The Australian National University, 2601 Canberra, Australia

** sebastian.tappe@mineral.tu-freiberg.de*

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Abstract

This Chapter provides an overview of the nature and origin of kimberlites, carbonatites and alkaline silicate rocks, with emphasis on the strategic mineral deposits linked to these types of mantle-derived volcanism (e.g., diamonds and rare-earth elements). It is demonstrated that many of the critical raw materials hosted by carbonatites and alkaline silicate rocks are essential for the transition to a renewable energy based global economy. Both geoscientific and metallurgical research will be of paramount importance to better understand these volatile-rich magmatic systems and their economic mineralization.

Keywords

critical metals, critical minerals, sustainable development, igneous rock classification, kimberlites, carbonatites, alkaline rocks, deep carbon cycle, diamond, volatile-rich magmatism, mineral systems approach, mineral carbonation

Sustainable Development Goals

Quality Education *(4)*, Affordable and Clean Energy *(7)*, Decent Work and Economic Growth *(8)*, Climate Action *(13)*

1. Introduction

1.1. Scope and definitions

Kimberlites, carbonatites and alkaline rocks occur mainly on thick continental lithosphere (Fig. 1), with the oldest known examples having Late Archean ages (Table 1). These igneous rocks represent some of the most extreme compositions among mantle-derived magmas, which is a consequence of near-solidus melting at ≥65 km depths in the presence of complex volatile mixtures (e.g., H_2O, CO_2, CH_4, SO_2, F, Cl), followed by reactive magma transport to Earth's surface. Scientific interest in kimberlite, carbonatite and alkaline magmatism – through the lenses of volcanologists, petrologists, geochemists and geodynamicists – has outweighed its volumetric significance. The high interest is not least caused by the tremendous economic opportunities that these rock types provide, ranging from gem-quality diamond to the strategically important rare-earth elements. The rare-earth elements and several other critical raw materials extracted from carbonatites and alkaline rocks play key roles in technologies for the decarbonization of energy and transport infrastructure. Understanding the genesis of carbonatites and alkaline rocks – including their economic mineralization – is therefore at the core of 'Climate Action' as stipulated by the UN Sustainable Development Goals [1]. Even the kimberlite rock tailings from diamond mines are being investigated for their feasibility to store atmospheric CO_2 via mineral carbonation, with the goal to offset the carbon emissions of the energy-intensive mining industry [2] – a commendable step towards a more sustainable future.

This Chapter gives an overview of volatile-rich mantle magmatism whose products were erupted at or emplaced near the Earth's surface, as is recorded by the global distribution of kimberlite, carbonatite and alkaline silicate rock occurrences (Fig. 1; Supplementary Information). Attention is paid to the mineral resources associated with this deep-sourced and often explosive volcanism, and examples of projects at the exploration and mining stages are presented. Given that many of these mineral systems host key resources of critical metals needed to decarbonize society, 'case-study' boxes are presented to elaborate on this particularly rewarding human interface with volcanoes.

Admittedly, the terminology of these mineralogically-diverse igneous rock types is cumbersome and burdened with jargon. It is therefore necessary to offer a set of simplified and updated definitions for future use (Box 1):

BOX-1 START *[text-only]*

- **Kimberlites** are hybrid ultramafic rocks that contain primary magmatic minerals (e.g., olivine, spinel, phlogopite, perovskite, apatite, monticellite, calcite) and abundant xenocrystic components (including diamond and Cr-pyrope garnet, as well as various megacryst phases), plus hydrothermal minerals such as serpentine. The textures of coherent magmatic kimberlites are typically inequigranular, characterized by up to centimetre-sized subrounded olivine macrocrysts set in a serpentine- or carbonate-rich groundmass (i.e., macrocrystic kimberlite). Compositionally, kimberlites can be described as SiO_2-undersaturated ultrabasic rocks (≤40 wt.% SiO_2 and ≥20 wt.% MgO) that may be enriched in K_2O (up to 3 wt.%) and volatiles (e.g., up to 10 and 20 wt.% H_2O and CO_2, respectively), but they are conspicuously deficient in Al_2O_3 (≤4 wt.%) and Na_2O (≤0.5 wt.%). The MgO-rich nature of kimberlites is coupled to an extreme enrichment in incompatible trace elements, which clearly sets them apart from other ultramafic volcanic rocks such as picrites and komatiites. The above definition refers to archetypal kimberlite, and the old subdivision into "Group I" and "Group II" kimberlites is now obsolete, because the ultrapotassic rocks known formerly as Group II kimberlite have no petrogenetic relationship to archetypal kimberlite, and they are more appropriately referred to as orangeite or lamproite.

- **Carbonatites** are plutonic or volcanic rocks that contain significant amounts of primary magmatic carbonate minerals, most commonly calcite and dolomite, but also ankerite, siderite and economically important alkali-REE-carbonates such as burbankite and carbocernaite. Bastnäsite and other REE-(fluoro)carbonate minerals are increasingly recognized in carbonatites, but typically these phases have secondary origins. The amount of magmatic carbonate minerals required to classify an igneous rock as carbonatite has been lowered over the last four decades, from ≥70 vol.% during periods of the 20th Century, through ≥50 vol.% to ≥30 vol.% [3]. Here, we recommend **25 vol.% as the cut-off value** for modal carbonates in carbonatites, regardless of SiO_2 content, which may become elevated via wall-rock contamination. This adjustment is in keeping with modern petrological concepts, which recognize that a carbonatite-forming magma will – through differentiation – create a suite of related rocks characterized by highly variable modal carbonate contents over short distances (i.e., at centimetre-to-decimetre scale). In other words, the exact amount of carbonate present is less important than the fact that a suite of carbonate-bearing rocks may be derived from a common parental carbonatite magma [3].
Igneous rocks with ≥25 vol.% primary carbonates do not exceed 45 wt.% SiO_2, which underscores the ultrabasic chemical composition of carbonatites and thus the non-arbitrary

nature of the recommended 25 vol.% threshold. Related igneous rocks with **5-25 vol.%** magmatic carbonates may be marked with the modifying prefix **'carbonate-rich'** – for example, 'carbonate-rich pyroxenite'. If the content of magmatic carbonate falls between **1-5 vol.%**, then the modifying prefix **'carbonate-bearing'** may be applied – for example, 'carbonate-bearing pyroxenite'. This simplified classification scheme is better suited for petrogenetic modelling and mineral resource evaluations, and it obviates the need for the rather uninformative term 'silicocarbonatite'.

Once an igneous rock with ≥25 vol.% primary magmatic carbonate has been identified as a true carbonatite, it is preferably described by its modal mineralogy, and the following petrographic varieties have been recognized in nature: calcite carbonatite, dolomite carbonatite, calcite–dolomite carbonatite, ankerite carbonatite, siderite carbonatite, nyerereite–gregoryite carbonatite, burbankite–bastnäsite carbonatite, etc. These petrographic varieties correspond broadly to the following respective geochemical groups: calciocarbonatite, magnesiocarbonatite, ferrocarbonatite, natrocarbonatite, and REE-carbonatite. The geochemical terminology is, however, inadequate for petrogenetic studies and should only be applied if a carbonatite rock is too fine-grained for an accurate mineral abundance to be determined. For the endorsed petrographic terminology, the name of the most significant non-carbonate mineral (the classification allows up to 75 vol.% of silicates, oxides, phosphates, halides, etc.) may be added as a modifying prefix to a carbonatite root name to specify its nature (e.g., amphibole calcite carbonatite, clinopyroxene calcite carbonatite, fluorite siderite carbonatite). This will improve the communication of large modal variations within a given suite of carbonatite and related rocks, such as frequently encountered in intrusive complexes. The use of historical type locality names, especially those with grain size connotations, is discouraged (e.g., sövite *versus* alvikite), because it serves no petrogenetic purpose.

- **Alkaline silicate rocks** are characterized by molar excess of alkalis relative to aluminium ([Na + K]/Al > 1). Alkaline igneous rocks therefore contain minerals with a similar excess of alkalis over that required to form feldspar, such as feldspathoids (nepheline, sodalite, cancrinite, kalsilite, leucite, etc.). Sodic clinopyroxene and sodic amphibole are also diagnostic of alkaline rocks. The modal presence of feldspathoids or their normative appearance justifies rock classification as *alkaline,* whereas the term *peralkaline* is reserved for igneous rocks that meet the above chemical definition, which even includes certain types of SiO_2-saturated rocks such as peralkaline granites. Peralkaline igneous rocks are separated into 'miaskitic' and 'agpaitic' types, the former having REE–HFSE dominantly hosted by zircon and titanite, and the latter containing complex Na-Ca-HFSE silicate minerals, commonly with structurally required halogens, such as eudialyte- or rinkite-group minerals [4]. Among the alkaline silicate rocks, it is the peralkaline agpaitic rocks that are most commonly associated with economic REE–HFSE mineralization. More generally, an igneous rock may be classified as alkaline if it contains more alkali metals (total alkalis; Na_2O+K_2O in weight-%) than is considered average for the group of rocks to which it belongs – for example, an alkali basalt. Contrary to common belief, kimberlites – as defined above – are not alkaline rocks. Their closest compositional equivalents among truly alkaline rocks are the carbonate-rich ultramafic lamprophyres, also known as aillikites.

BOX-1 END *[text-only]*

1.2. From scientific legacy to world-class resources of critical metals

Kimberlites, carbonatites and alkaline silicate rocks are rich in information about the inner workings of Earth (Table 1). Most of our understanding of the origin and evolution of continental roots is linked to the emplacement of kimberlite magmas at the Earth's surface. This knowledge is gleaned from the emplacement patterns and compositions of kimberlites through time, but more directly from the study of entrained mantle cargo, which includes peridotite and eclogite xenoliths, a wide variety of megacrysts such as centimetre-sized mantle zircon and, with a bit of luck, diamond. The engrained petrology concept of 'xenoliths' and 'xenocrysts' could not be any more obvious than during inspection of a typical magmatic kimberlite sample, where these foreign objects reach official status as part of the definition of kimberlite rock. Mantle cargo in kimberlites is used to constrain magma ascent rates applying hydrogen- and argon-based diffusion speedometers to olivine and phlogopite, respectively. Results suggest that kimberlite magmas can traverse 200 km of continental lithosphere within a few hours (less than a day), which is much faster than the ascent speed of alkali basalt magmas, and probably crucial for the preservation of precious diamond xenocrysts [5]. Once kimberlite magmas reach the surface, relatively thin feeder dykes, less than one meter in thickness, are often thought to interact explosively with groundwater over the course of multiple eruptions to create the characteristic maar–diatreme volcanoes [6]. Alternatively, some researchers prefer a model whereby the energetics of eruption are guided entirely by the violent degassing of magmatic volatiles [5]. Regardless of the importance of phreatomagmatism, the kimberlite volcanic pipe excavation and infill process is pictured in virtually every textbook of volcanology, petrology and economic geology. Kimberlites also serve as probes into the deep carbon cycle, given the abundance of primary groundmass carbonate, a feature they have in common with carbonatites and some alkaline rocks such as aillikites. Much debate surrounds the origin of these carbonates, and it may well be that both recycled crustal and primordial mantle derived carbon components preconditioned the sources of kimberlites, carbonatites and alkaline silicate rocks [3]. Kimberlite, carbonatite and alkaline silicate magmas result from, but also provide the agents for, mantle metasomatism, which is a paradigm for chemical and mineralogical enrichment of depleted mantle rocks such as peridotite. In a similar way, metasomatic reactions occur between intruding carbonatite and alkaline silicate magmas and their quartzofeldspathic country rocks in the upper crust. This fluid-mediated mass transfer process is known widely as 'fenitization', named after the carbonatite type locality at Fen in Norway. The spatiotemporal partnership between carbonatites and alkaline silicate rocks in numerous intrusive complexes worldwide is typically explained by melt–melt unmixing from a common carbonated silicate parent magma after protracted fractional crystallization – a process known as liquid immiscibility. Geological and experimental evidence suggests that liquid immiscibility is an important prerequisite to economic rare-earth element mineralization in carbonatites [7], which is one of the main reasons why these peculiar rocks are currently enjoying a renaissance. For example, >50% of the world's current supply of rare-earth elements come from Bayan Obo – a single, poorly understood carbonatite deposit in China. Data from the Web of Science (Clarivate) illustrate how the annual research output for 'carbonatites' and related topics has more than doubled over the past ten years, with many of the new research contributions exploring links between petrogenesis and ore-forming processes.

Clearly, kimberlites, carbonatites and alkaline rocks have inspired the world of science, and they have contributed their fair share of stunning mineral specimens to natural history museum collections – but all legacy aside – where, when and how do these magmas form?

2. Deep mantle melting in the presence of volatiles

2.1. Kimberlites

A striking feature of kimberlite magmatism is the strong coupling to cratons characterized by cold and thick lithospheric mantle roots that reach down to 250 km depth (Fig. 1). The nature of primary kimberlite melt is only poorly understood, with *(i)* carbonate melts, *(ii)* carbonated silicate melts, and *(iii)* volatile-poor ultrabasic silicate melts being debated [8]. There is consensus, however, that these melts originate from below the lithosphere by low-degree partial melting of mainly peridotite (Fig. 2). The primary liquids evolve by reactive melt transport and digestion-fractional-crystallization during ascent through the cratonic lithosphere, where kimberlite magmas ultimately develop their diagnostic hybrid nature [9].

Although kimberlites formed as early as 2.85 Ga ago, more than 60% of known kimberlite clusters formed during only 4% of Earth history between 250 and 50 Ma [10]. This Mesozoic–Cenozoic 'kimberlite bloom' occurred globally and has been associated with breakup of the Gondwana–Pangea supercontinent, when plate-tectonic reorganisation caused kimberlite magma ascent from the lithosphere–asthenosphere boundary and possibly deeper. Less than 2% of global kimberlites formed after 50 Ma. These rare 'young' kimberlites are located on cratons that are currently exposed to active plate-tectonic forces such as subduction in western North America (e.g., Eocene kimberlites on the Slave craton) and continental rifting in eastern Africa (e.g., Oligocene kimberlites on the Congo craton). The ca. 12 ka Igwisi Hills monogenetic volcanoes on the Tanzania craton are regarded as the youngest kimberlites on Earth, although their groundmass mineralogy deviates slightly from that of archetypal kimberlites. Kimberlite is one of the few volcanic rock types without historically active volcanoes during the Holocene, which means that understanding the physical properties and eruption behaviour of these highly explosive magmas requires detailed petrological research. This research is commonly carried out on drill-core obtained during diamond exploration, but analogue laboratory experiments have also been conducted. Extrusive kimberlites, in the form of lava flows and lava lake deposits, are increasingly recognized in some maar–diatreme structures, which further challenges theories in which explosive kimberlite volcanism is controlled mainly by magmatic volatile degassing [5].

The late appearance of kimberlites in the geological record and the episodic nature of this volatile-rich ultramafic volcanism have been explained by a combination of secular mantle evolution and increased tectonic activity during supercontinent cycles [10]. Decreasing temperatures and increasing volatile plus redox budgets of the ambient convecting mantle, due to heat-loss and more efficient subduction recycling through time, promote low-degree 'incipient' melting – the mechanism required for kimberlite magma petrogenesis (e.g., redox melting). On the Phanerozoic Earth, such incipient low-degree melts may have been common beneath thick continental roots, and their success rate to reach Earth's surface increased significantly during times of global tectonic unrest. A plate-tectonics control helps to explain the overlap between peak kimberlite magmatic activity and supercontinent assembly and breakup stages. Although continental breakup may be causally linked to mantle plume activity, evidence from petrology and isotope geochemistry points to the ambient convecting mantle as the source for most kimberlites, without significant contributions from anomalously hot or compositionally extreme deep Earth reservoirs such as the core–mantle boundary [8, 11]. However, the quest for exotic components in plume sources of kimberlite magmas, such as remnants of primordial mantle, is an active research direction sparked by the occasional discovery of isotopic anomalies [12].

Kimberlite magmas are rich in dissolved volatile components, mainly H_2O and CO_2, which is shown in part by the extreme velocity of their ascent through the shallow lithosphere, driven by volatile exsolution. This implies the presence of volatile components in the deep mantle sources of the melts parental to kimberlite magmas. High-pressure experimental petrology has been an important approach to understanding the partial melting processes in the deep upper mantle. Experimental campaigns have explored variables such as pressure, temperature, source rock composition and volatile abundance plus type (H_2O, CO_2, CH_4, etc.) in attempts to produce melt compositions matching those of natural kimberlites [8]. Alternative approaches have searched for the conditions of pressure, temperature and volatile abundance plus type at which kimberlite melts are saturated in mineral assemblages such as Mg-rich olivine, enstatite-rich orthopyroxene and Cr-pyrope garnet. Unfortunately, both approaches are hampered by two factors: *(i)* kimberlites have been compositionally modified through fractional crystallisation, contamination with mantle debris and devolatilization during magma ascent also causing alteration, and *(ii)* the experimental strategies assume that partial melting of the deep upper mantle is a relatively simple process at some unique equilibrium condition of pressure, temperature and volatile content. However, the formation of kimberlite melt in nature is far more complex.

Constraints from the extinct ^{182}Hf-^{182}W isotope system, combined with data from extant radiogenic isotope systems such as ^{147}Sm-^{143}Nd and ^{176}Lu-^{176}Hf, suggest that the main source components of parental kimberlite melts reside in the asthenosphere directly beneath the cratonic lithosphere [11]. The asthenosphere is chemically fertile and reduced in oxygen fugacity so that CHO volatile components exist as diamond + H_2O, perhaps with minor CO or CO_2. Partial melting of the ambient convecting mantle beneath cratons may be triggered by upwelling, leading to decompression-related oxidation of the volatile components (such as diamond) to form CO_2 + H_2O-rich fluids, which lower the peridotite solidus causing localized melting [8]. Experimental studies have shown that such melts would be alkali-bearing carbonatites near the solidus or ultramafic lamprophyres at slightly higher temperatures, but their stability and mobility in the fairly reduced asthenosphere is uncertain [13]. If these low-volume melts infiltrate the refractory and even more reduced cratonic mantle lithosphere, they may undergo redox freezing, a process by which continental roots become re-enriched and more oxidized [14]. Subsequent asthenosphere-derived melt may interact with such metasomatized zones at the lithosphere–asthenosphere boundary, which increases K and Ca contents approaching compositions reminiscent of the parental melts to kimberlites. Further magma modification takes place during interaction with more refractory portions of the cratonic mantle lithosphere [9], which may lead to entrainment of diamonds (Section 4.1).

2.2. Carbonatites

In contrast to kimberlites, several Quaternary carbonatite volcanic systems are known, with Oldoinyo Lengai in the East African Rift being presently active. The oldest known *(meta-)*carbonatite is a relatively small ca. 3 Ga intrusive sheet within Mesoarchean tonalitic gneisses in West Greenland (Supplementary Information); however, sizeable Neoarchean carbonatite intrusions occur on the Canadian–Greenland and Fennoscandian shields (2.75-2.57 Ga). As for kimberlites, the age distribution of global carbonatites is skewed towards the Mesozoic–Cenozoic, but it is more continuous with <50 Ma 'young' carbonatites being most abundant. Many Phanerozoic carbonatites were emplaced near craton margins because remobilized, faulted and thinned lithosphere aided magma ascent (Fig. 2). In contrast, craton interiors were the preferential sites of Phanerozoic kimberlite volcanism, which implies that primitive carbonatitic melts interacted vigorously with the thick lithospheric mantle roots, possibly giving rise to hybrid kimberlite magmas (Section 2.1). Alternative views suggest, however, that primary carbonate melts in the Earth's mantle cannot be parental to kimberlites, because their stability is limited to relatively shallow mid-lithospheric depth levels [16]. Although carbonatites occur mainly on the continents (Fig. 1), several localities are known from oceanic islands (Kerguelen, Cape Verde, Canary Islands), where minor carbonatite magmatism is associated with alkali basalts and their differentiation products. Unlike kimberlites, carbonatites have been reported from a variety of tectonic settings (stable cratons, reactivated craton margins, continental rifts, post-orogenic extension, oceanic plateaus, etc.) (Fig. 2), and the majority of carbonatites are associated with a rich diversity of alkaline silicate rocks (ultramafic lamprophyres, melilitites, nephelinites, phonolites, etc.). This reveals that carbonatites have a range of origins within the upper mantle (e.g., asthenospheric and lithospheric sources), and diverse carbonated melts ascend along distinctly different evolution paths through the crust [3, 15].

There are two main categories of petrogenetic models for carbonatites. First, the parental melts to carbonatites form in the upper mantle by low-degree partial melting of carbonate-bearing peridotite or eclogite. Second, carbonatite-forming melts segregate within the crust by liquid exsolution from mantle-derived carbonated silicate magmas such as ultramafic lamprophyres, melilitites and nephelinites or their plutonic equivalents (Fig. 2). Experimental studies at upper mantle pressure demonstrate that the lowest temperature melts of carbonate-bearing peridotite or eclogite have alkali-enriched calcio-dolomitic to dolomitic compositions. Carbonate melts can migrate rapidly through the mantle due to their ability to effectively wet silicate mineral surfaces and because of their low density (Section 3). However, melt transfer from the mantle to the crust may be inhibited by decarbonation reactions in the uppermost mantle [13].

Several experimental studies have explored immiscibility between silicate and carbonate melts at upper mantle and crustal pressures. These studies demonstrated that a two-liquid solvus may exist on the liquidus surface of melts in these systems. Ultramafic lamprophyres, melilitites, nephelinites or evolved phonolites are considered to represent the parental liquids that can undergo immiscibility because they contain a few weight-% of dissolved carbonate when entering the shallow lithosphere. Upon continued magma ascent, fractional crystallisation of silicate and oxide minerals (e.g., olivine and magnetite, respectively) drives the residual liquids onto the solvus, leading to unmixing of a carbonate melt. Immiscibility is most likely to occur at lower-to-mid crustal conditions. Once a carbonate melt has separated from its conjugate silicate melt, it rises more rapidly through the crust towards final emplacement. Field evidence for carbonatite genesis by immiscibility is the spatiotemporal association between magmatic carbonatite occurrences (75% of the global database) and alkaline silicate rocks [13, 15]. Additional evidence comes from fluid and melt inclusions that show liquid unmixing at the micro-scale [16]. In contrast, evidence for carbonatite origins by direct partial melting of the upper mantle is very rare, being largely restricted to the absence of associated silicate rocks including calcite kimberlites (<20%).

Regardless of their ultimate origins, alkali-bearing Ca-rich carbonatites crystallize calcite on the liquidus, joined by apatite. As the $MgCO_3$ and $FeCO_3$ contents of the residual melts increase, dolomite and Fe-rich carbonate minerals may form, along with various oxide and silicate minerals (e.g., magnetite, pyrochlore, perovskite, phlogopite, clinopyroxene). Other anions present in the carbonatite melt (fluoride, phosphate, sulphate, etc.) will lead to fluorite, REE-phosphate and baryte mineralization. Hence, the final carbonatite intrusion emplaced into the Earth's crust is a complex assemblage of accumulated minerals, with no individual carbonatite rock sample directly representing the parental liquid from which it formed.

2.3. Alkaline silicate rocks

Because of their broader definition it is unsurprising that alkaline silicate rocks are more common than kimberlites and carbonatites and that they are more widely distributed (Fig. 1). They have emplacement ages comparable to those of kimberlites and carbonatites, ranging from Neoarchean (2.75 Ga) to the present day, following the dynamic evolution of Earth's crust–mantle system. Similar to their carbonatite cousins, the REE–HFSE mineralization potential of alkaline silicate rocks has changed through time, with major economic deposits formed globally at around 1.2 Ga [17, 18]. Both carbonatite and alkaline silicate magmatic systems are localized by sutures at the margins of continental shields, occurring along rift zones, in post-collisional settings and as intraplate magmatism associated with thermochemical mantle plumes (Fig. 2). On oceanic islands, alkaline silicate systems are typically unmineralized.

Experimental studies and thermodynamic modelling show that the origins of alkaline SiO_2-undersaturated melts are best explained by low-degree partial melting of lithologically heterogeneous upper mantle domains beneath thick continental lithosphere. However, fundamental questions remain concerning the composition and mineralogy of the mantle sources for alkaline magmas, and the impact of source composition on mineralization potential. A few occurrences appear to preserve primitive isotopic signatures from the deep sublithospheric mantle, whereas others record inputs from recycled surface materials such as altered oceanic crust and pelagic sediment [19]. In some cases, the REE–HFSE in mineralized alkaline silicate and carbonatite magmatic systems are sourced from metasomatically enriched lithospheric mantle rocks (inferred to be phlogopite- or amphibole-bearing), without significant contributions from subduction recycled crustal components [20]. A mechanistic understanding of the factors that produce and preserve fertile mantle sources capable of generating world-class REE–HFSE deposits could be used to guide exploration campaigns toward geologically favourable areas (Fig. 2).

3. Magma properties relevant to ascent, emplacement and economic mineralization

There are similarities and differences between kimberlite, carbonatite and alkaline silicate magmas regarding their physicochemical properties and, thus, emplacement modes (Fig. 3). For example, kimberlitic magma is denser (~2.5-3.2 g/cm^3) and has a higher viscosity (>10 Pa s) than carbonatitic melt (~2.2-2.6 g/cm^3; ~0.01 Pa s). Carbonatite melts are ionic liquids, whereas kimberlite and alkaline silicate melts are characterized by interconnected Si–O tetrahedra resulting in a polymerized melt structure. Unlike alkaline silicate melts, kimberlite and carbonatite magmas are unable to create sizeable magma chambers because this would require horizontal expansion, which is difficult to attain in such fast-ascending volatile-rich magmas of limited supply. Kimberlite and carbonatite magmatic bodies are characterized by steep-sided pipe- or plug-like structures, fed by relatively narrow dykes typically <1 meter wide. Shallow-level sill complexes are common in kimberlite volcanic fields, whereas they are rare to absent in carbonatite complexes owing to the much lower viscosities and crystal loads of carbonatite magmas (Fig. 3).

Although the magmas discussed here rise rapidly through the lithosphere, theoretical ascent rates are highest for carbonatites, followed by kimberlites and alkaline silicate magmas. For kimberlite magma, buoyancy is amplified by exsolution and decompression of CO_2-rich fluids in a frothy melt. Melts charged with CO_2 bubbles become increasingly more buoyant upon ascent due to expansion of the bubbles. Subsequent CO_2 degassing is explained by melt decompression and by a drop in melt CO_2 solubility with increasing SiO_2 – for example, a 15% drop in CO_2 solubility from 15 to 30 wt.% SiO_2 in the melt. Dissolution of entrained silicate minerals, such as orthopyroxene xenocrysts, has been discussed as the main process by which the SiO_2 content of kimberlitic melt increases [5]. Carbonatite magma ascent is not directly comparable to kimberlite magma ascent and emplacement, mainly because of the different melt structures. In general, carbonatite melts contain more CO_2 and H_2O (up to 10 wt.% H_2O at 0.1 GPa) than kimberlites [21], and these common volatile species are only limited by melt stoichiometry. Hence, carbonatite melts can exsolve significantly higher volumes of fluid than kimberlite magma, which helps to explain the large alkali metasomatic fenite aureoles that characterize carbonatite complexes in contrast to the lack of geochemical halos around kimberlite intrusions (Section 4).

Fluid inclusion studies show that liquid immiscibility and boiling (effervescence) are important processes in carbonatite magmatic systems. Magma effervescence occurs due to reduction in confining pressure and associated volume expansion – processes accompanied by fracture development in the country rocks. Expansion also leads to fluid overpressure, which – in combination with continuous melt degassing and country rock fracturing – creates the positive feedback loop responsible for rapid carbonatite magma ascent. Cyclic pressure build-up in carbonatitic fluids and catastrophic pressure release are comparable to the working mechanism of a pneumatic jackhammer [16]. Jackhammer-style ascent of carbonatite magma in a vertical conduit can explain the formation of roof breccias during magma propagation and the simultaneous development of fenite aureoles in the sidewalls to an active conduit system. These processes facilitate economic REE–HFSE mineralization because metasomatic reactions are initiated by expelled magmatic–hydrothermal fluids (Fig. 3). Experiments and fluid flow modelling indicate that carbonatitic melts can ascend to Earth's surface at supersonic speed of up to 65 m/s [16], which is twice as fast as kimberlite magma ascent (<30 m/s). If correct, then carbonatite and kimberlite magmas may traverse the continental crust within a few minutes. In contrast to rapidly emplaced kimberlite and carbonatite magmatic bodies, the more gradual ascent and repeated emplacement at a similar structural level in the crust favours development of highly evolved, alkali-enriched silicate magmas (Fig. 3).

Numerous intrusive carbonatite complexes exhibit outward zonation, with carbonatite rocks in the core followed by phoscorites, ultramafic rocks (dunite, clinopyroxenite, etc.) and alkaline silicate rocks towards the margins (Fig. 3). The ultramafic rocks, traditionally viewed as cumulates, have recently been interpreted as products of metasomatic replacement of felsic country rocks (e.g., granites) by invasive carbonatite magmas that may get consumed during this process. In cases where the replacement reactions do not go to completion, residual carbonatite magmas and related phoscorites may form economic mineralization such as apatite-hosted phosphorous deposits at Kovdor in Russia and at Palabora in South Africa. During the waning stage in the evolution of such complexly zoned conduits, primitive carbonatite magmas may intrude the cores [22].

A broad variety of intrusion morphologies are catalogued for alkaline silicate magmatic systems. Those that host mineralization typically comprise polyphase layered intrusions of SiO_2-undersaturated alkaline rocks emplaced at subvolcanic levels (1-4 kbar pressure equivalent to 3-12 km depth) (Fig. 3). Up to 40 km wide, these intrusive complexes may preserve an inward decrease in SiO_2 and an increase in alkalis with time, tracking progressive fractional crystallization at a variety of pressures [18]. Economic deposits are also associated with smaller, sometimes SiO_2-saturated peralkaline intrusions with diameters below 6 km (e.g., Strange Lake, NE Canada). Carbonatites associated with alkaline intrusions tend to cross-cut the silicate-dominated rock units, which indicates carbonatite magma emplacement is late, for example at Khibiny in NW Russia. Importantly, magmatic–hydrothermal REE-HFSE deposits associated with alkaline magmatism are characterized by enrichment of the more valuable HREE, with higher HREE/LREE ratios compared to carbonatites. Extensive volcanism appears to be incompatible with economic mineralization because volatile release can dissipate REE–HFSE into the wider environment (Fig. 3). This model is consistent with the limited fenite aureoles typically observed at strongly mineralized alkaline silicate or carbonatite complexes, relative to larger metasomatic halos around barren intrusions of these magmatic suites [23].

For alkaline silicate magmas, retention of volatiles combined with the high melt alkali contents result in low viscosities and depressed solidus temperatures relative to more common metaluminous and peraluminous magmas. These properties of peralkaline magmas impede the crystallization of zircon and other REE–HFSE phases, which permits enrichment of these metals in residual melts during extensive fractional crystallization, inferred to exceed 98%. Such extreme magmatic differentiation creates large volumes of mafic cumulates in sizeable mid-crustal intrusions, as imaged geophysically beneath some of the world's largest REE–HFSE mineralized peralkaline complexes (i.e., Ilimmaasaq in southern Greenland and Lovozero in NW Russia) (Fig. 4).

4. Exploration, mining, and sustainable development

4.1. Kimberlites as primary sources of diamonds

Kimberlites are the most significant source of natural diamonds. Diamond production from kimberlite-hosted primary deposits accounts for 80% of the value of this industry, with the remainder generated by diamond mining from orangeite and lamproite deposits, and by alluvial diamonds originally sourced from kimberlite pipes. Importantly, discovery of new kimberlite occurrences does not guarantee the presence of new diamond resources. Exploration data show that only 10-15% of the approximately 6000 known kimberlite bodies worldwide can be classified as 'diamondiferous' [10], where macro-diamond (≥0.5 mm) grades exceed 0.01 carat per tonne of ore (1 metric carat = 0.2 gram). Approximately 1% of global kimberlites and related rocks contain economic quantities and qualities of diamond, and they host the 60+ active and past-producing diamond mines. Tier-1 diamond deposits contain over 20 billion US$ reserves prior to exhaustion, and only seven kimberlite-hosted diamond mines hold this status, listed in order of decreasing value: Jwaneng (Botswana), Orapa AK1 (Botswana), Udachnaya (Russia), Venetia (South Africa), Cullinan (South Africa), Catoca (Angola), Mirny (Russia). Five out of these seven superlative diamond deposits are located in sub-Saharan Africa, a cratonic region where favourable interplay between deep mantle, lithospheric and surficial processes has created a diamond mineral system that is more prosperous compared to those developed on other continental shields (Fig. 4).

Exploration for and evaluation of economic kimberlites are very expensive and time-consuming operations. In 2007, global diamond exploration expenditure peaked at approximately 1 billion US$, an amount comparable to the annual education budget of South Africa around this time, but has been in decline ever since. Despite record investment, following discovery of the giant Venetia and Catoca kimberlite-hosted diamond deposits in the 1980s, and first economic kimberlites in Canada during the 1990s, no new Tier-1 kimberlites have been found. However, discovery of much smaller diamond deposits has continued, but development from advanced exploration projects to active mining is sluggish, taking on average 20 years – for example, 21 years for the Gahcho Kue mine in Canada (Fig. 4), which opened recently in 2016. But why is diamond exploration and mining so much more complicated in comparison to other magmatic ore systems such as sulfide-hosted Ni-Cu-PGE mineralization?

One of the main reasons is the disconnect between the ultimate diamond sources in the cratonic mantle lithosphere and the formation of kimberlitic carrier magmas in the asthenosphere (Fig. 2). In other words, kimberlites and their entrained diamonds were forced into a non-genetic relationship during joint ascent from >150 km depths. This random mantle sampling process is at the very root of the most vertically extensive economic mineral system. Mineral inclusions in diamonds suggest that several distinct mantle lithologies can be host to diamond formation, with harzburgites and eclogites being most important. Diamonds from sources in the deep convecting mantle down to 800 km depth are increasingly recognized (<10% of global diamond production). Such ultradeep mantle cargo, commonly carrying subduction fingerprints, adds yet another layer of complexity to the ultimate origin of kimberlite magmas (Fig. 2). However, there is growing awareness that such sublithospheric diamonds may have been transported from great depth over million-year timescales via mantle convection or plume activity to the lithosphere–asthenosphere boundary, which is a more viable source region for kimberlite magmatism [11].

Until recently, discovery of kimberlite-hosted diamond deposits was driven by the search for and analysis of kimberlite indicator minerals (e.g., Cr-pyrope, Cr-diopside, Mg-ilmenite). However, this traditional pathfinder tool is slowly losing its edge because of growing awareness that the diamond formation process and favourable mantle sources are much more diverse than previously thought. Indicator mineral techniques, including paleogeotherm estimates to constrain a lithospheric 'diamond window' for a target area, are unable to predict the presence of sublithospheric CLIPPIR diamonds (**C**ullinan-like, **l**arge, **i**nclusion-**p**oor, **p**ure, **i**rregularly shaped, **r**esorbed) [24]. The extraordinary value of these special stones has turned otherwise marginal diamond deposits into lucrative operations (e.g., Letseng in Lesotho, Karowe in Botswana) (Fig. 4). New X-ray technology is being implemented at many mines to identify kimberlite-encased CLIPPIR diamonds prior to rock crushing to mitigate diamond breakage, which causes exponential loss in diamond value.

Once kimberlite magmas and their mantle cargo reach the Earth's surface via discontinuous dyke propagation, a series of eruptions causes maar-diatreme volcanoes to form (Fig. 3). Several diatremes can grow along the same feeder dyke, and numerous sets of shallow-seated precursor dykes may create clusters of kimberlite pipes [6]. For example, the kimberlite cluster within the greater city perimeter of Kimberley in South Africa comprises at least 24 individual kimberlite pipes, of which one is 'The Big Hole' museum of diamond mining (https://thebighole.co.za). Although these subvertical pipes were emplaced only 90 million years ago, they have undergone 1 km of erosion, with present-day surface diameters of less than 500 meters. Small pipe sizes and the fact that kimberlite volcanic features are exceptionally rare provide a great challenge to the exploration geologist. Moreover, kimberlite pipes do not have much of a geochemical halo and their erosional remnants are typically covered by thick overburden (e.g., glacial till, desert sand, flood basalt), or even lakes. Consequently, airborne or ground geophysical surveys have the highest chance of success in finding new kimberlite pipes. Commonly employed methods are gravity and magnetic surveys, which also provide first-order information about the size and shape of kimberlite pipes and, thus, estimates of ore tonnage. However, the multiphase architecture of kimberlite pipes can only be constrained by detailed drill-core logging, which is the most informative guide to effective diamond testing of the different kimberlite facies and phases (e.g., coherent magmatic, pyroclastic and resedimented volcaniclastic kimberlite units). Early-stage drill-core testing aims to recover micro-diamonds to predict a diamond size distribution (the coarser the better), but eventually several tonnes of kimberlite need to be treated for macro-diamond recovery in order to gain more confidence about diamond grade and value (Fig. 4). This is a tedious exercise because even the best-endowed kimberlites hardly ever exceed a diamond concentration of 1 ppm. For instance, the Internationalnaya pipe in Siberia has a record diamond grade of 10 carats per tonne of kimberlite ore, which is equivalent to only 2 ppm by weight (Fig. 4).

At an advanced project stage, information gleaned from kimberlite petrology and geophysical modelling is combined into a geological pipe model that is used in conjunction with the diamond data to define a mineral resource. Unlike most other commodities, rough diamond does not have a trading price dictated by a stock exchange. Instead, diamonds are combined into parcels guided by consumer preference and assigned an arbitrary value that somehow reflects stone size, shape, colour and clarity. Special stones larger than 10 carats, such as CLIPPIR diamonds, are sold through auctions at prices that may reach several tens of millions of US$. No resource model can in any reliable way forecast this level of extravagance when mining a volcano for diamonds.

From the above, it becomes clear that a successful diamond exploration team must be multi-facetted, embracing expert knowledge across the entire spectrum of the geosciences (Precambrian geology, tectonics, petrology, mineralogy, geochemistry, volcanology, sedimentology, geometallurgy, geophysics, statistics, geospatial data management, etc.). The diamond industry has also recognized that its future success will rely on accountability regarding contributions to sustainable development. But how can mining of a luxury good such as diamond make a positive contribution to sustainable development? One of the industry leaders aims for carbon-neutral mining by 2030 via atmospheric CO_2 capture in kimberlite tailings (https://www.debeersgroup.com/reports). However, the latest research on carbon mineralization from the Venetia kimberlite-hosted diamond deposit indicates that mine site CO_2 offset does not exceed 20% [2].

4.2. Carbonatites as sources of critical raw materials

Carbonatites are highly prospective exploration targets for a wide variety of metal commodities including base-metals (e.g., Cu-Fe at Palabora, South Africa), precious-metals (e.g., Au at Lac Short, Canada), technology-metals (e.g., Zr at Kovdor in Russia and Nb at Araxa in Brazil), LREE (e.g., Mt Weld, Australia), and HREE (e.g., Lofdal, Namibia) (Box 2). Carbonatites may also host Th-U reserves of strategic importance as nuclear fuels (e.g., Th at Fen in Norway and U at Sillai Patti in Pakistan), as well as economic resources of F and P (e.g., fluorite at Okuruso in Namibia and apatite at Sokli in Finland). Approximately 60 out of the 600 known carbonatite occurrences host economic reserves of the types reported above (Fig. 4), often in multi-commodity style. This makes 'carbonatite' one of the best-endowed mineral systems in the world, with a mine-to-occurrence ratio of 1:10 (compare with kimberlite-hosted diamond mines at 1:100 and orogenic gold systems at 1:1000). However, exploration for carbonatite-hosted mineral deposits has its challenges because of the relatively small sizes and multi-commodity nature, which limits the utility of simple exploration models.

In general, carbonatites need to pass through several magmatic and hydrothermal stages before economic grade and tonnage for a specific commodity are reached. These stages include: *(i)* fractional crystallisation, *(ii)* multiple liquid immiscibility events, and *(iii)* evolution towards a brine melt [7]. In addition, autometasomatic processes are highly beneficial for reaching economic levels of enrichment, which may be achieved in 'unerupted' or 'failed' carbonatite complexes (Fig. 3). For many world-class critical metal deposits hosted by carbonatites, a major phase of supergene upgrading has occurred (see Part 8 Glossary), typically associated with tropical laterization (Box 2). Laterites developed on top of carbonatite complexes may reach >100 meters thickness, in particular in stable cratonic regions where old weathering profiles can be preserved (e.g., Bingo in the D.R. Congo, Araxa in Brazil, Mt Weld in Australia). Laterite formation can increase metal concentrations by an order of magnitude (Fig. 4). Besides supergene enrichment, erosion of carbonatites can contribute to secondary critical metal deposits in river and coastal environments such as monazite sands.

Mining of carbonatite-hosted ore deposits faces two main geometallurgical challenges, namely the complexity of the ore mineralogy and the presence of abundant radioactive nuclides such as Th. Other deposit types, such as orogenic gold and porphyry copper systems, typically host the metal of interest in a single dominant mineral phase. For these mineralogically simple systems, grade and tonnage are the main constraints on economic viability of mining. In contrast, processing of the complex ore mineralogy of a typical carbonatite presents a great challenge for mining companies, and processability is often the main reason for project failure. For example, the rare-earth elements are hosted by numerous mineral phases in a carbonatite rock, such as pyrochlore, monazite, xenotime, alkali-REE-carbonates (burbankite, carbocernaite, etc.), REE-F-carbonates (e.g., bastnäsite), and even by 'gangue' calcite. Thus, improved geometallurgical engineering to cope with these mineralogically complex rocks is a prerequisite for successful mining of critical metals from carbonatites in the future. Additional complexity encountered during carbonatite mining is the high abundance of radioactive Th and U, which requires special consideration. These 'penalty elements' are commonly hosted by the REE–HFSE ore minerals themselves and, thus, will not be removed until late during ore processing. This may result in mine waste with elevated radioactivity; however, the associated environmental risk can be mitigated if Th and U are extracted and sold as a by-product.

New exploration for carbonatite-hosted ore deposits occurs mainly in areas with past-producing mines in so-called 'brownfields' operations. Also, many known occurrences are being re-evaluated because of the steadily increasing prices of critical metals. Modern carbonatite exploration applies a range of geophysical and geochemical methods scaled to targets at province, district and prospect levels. As for kimberlites, carbonatite bodies are relatively small targets with diameters of intrusive complexes and thicknesses of dykes rarely exceeding 1 km and 10 m, respectively (Fig. 3). This necessitates fine line spacing during exploration programs to discover economic mineralization that is typically confined to even smaller compartments within a carbonatite body. The mineral systems approach applied to carbonatite exploration considers the following processes sequentially: *(i)* mantle source fertilization such as during subduction events, *(ii)* establishment of lithospheric pathways (rifts, suture zones, lineaments, faults), *(iii)* tectonic trigger events to cause carbonatite magma formation, ascent and emplacement, and *(iv)* uplift and erosion to create optimal exposure levels (Fig. 2).

BOX-2 SOMEWHERE HERE *[text+figure, see the end of manuscript]*

4.3. Alkaline silicate rocks as sources of REE–HFSE

Alkaline magmatic complexes host some of the world's largest REE–HFSE deposits [18], with comparable metal tenor to carbonatites (Fig. 4). However, a major attraction of mineral deposits associated with alkaline magmatism is their much higher concentration of the more valuable HREE compared to LREE-enriched carbonatites. Mineralization in alkaline silicate magmatic systems is divided into: *(i)* orthomagmatic deposits associated with layered agpaitic syenite intrusions, *(ii)* silicate roof-zone deposits at the contacts between alkaline intrusions and country rocks, and *(iii)* mineral deposits linked to peralkaline granites and pegmatites (Box 3). The former include deposits exceeding 10 Mt contained TREO (e.g., Lovozero in NW Russia, Ilimmaasaq in southern Greenland), and the latter host much smaller deposits not exceeding 6 Mt contained TREO (e.g., Strange Lake in NE Canada) (Fig. 4).

Although partial melting of metasomatized lithospheric mantle sources and subsequent focussed magma ascent, including extensive fractional crystallization, are prerequisites to REE–HFSE deposit formation in alkaline silicate magmatic systems (Sections 2 and 3), a final mechanism to further concentrate and trap the metals is required. For example, cumulate layers containing eudialyte-group minerals form in response to cooling of SiO_2-undersaturated agpaitic magmas (Box 3). Autometasomatism of such orthomagmatic mineralization is common, typically driven by residual Na-rich melts and fluids, which are flushed through the cumulate rocks. This internally induced mechanism can initiate the recrystallization of primary mineralization leading to fine-grained assemblages of secondary REE–HFSE minerals that are more difficult to process. In roof-zone deposits, REE–HFSE mineralization is trapped via reactions between metal-rich residual melts/fluids and the roof rocks of alkaline intrusions, with the Motzfeldt syenite of the ca. 1.2 Ga Gardar rift in southern Greenland as a prime example (Fig. 4). Granite- and pegmatite-hosted REE–HFSE deposits form near the roofs of plutons (e.g., Strange Lake in NE Canada), but also in plugs or dykes external to a major SiO_2-undersaturated magmatic center (e.g., the T-zone at Nechalacho in NW Canada) (Box 3). Such mineralization may result from hydrothermal mobilization of metals from primary agpaitic mineral assemblages. The REE–HFSE may also become enriched via immiscible fluoride melts, as evident from fluorite-rich breccias and melt blebs in some peralkaline granites (Fig. 3).

Mining of REE–HFSE deposits hosted by alkaline magmatic complexes has thus far been hampered by their mineralogical diversity, which is far greater than in typical base- or precious-metal deposits, and only rivalled by carbonatite-hosted REE–HFSE deposits (Section 4.2). To date only a few alkaline complexes have been commercially mined: i.e., Lovozero in NW Russia for loparite-hosted REE–Nb, Khibiny in NW Russia for apatite- and titanite-hosted REE–HFSE, Nechalacho in NW Canada for bastnäsite-hosted REE, and Weishan in China for bastnäsite/parisite-hosted REE (Box 3). Currently, numerous mineral exploration projects are targeting alkaline provinces worldwide (Fig. 1). Development of new geometallurgical procedures for the extraction of REE–HFSE from many different host minerals is at pilot stage. Unfortunately, classic grade–tonnage diagrams contain only limited information about project economics because the REE–HFSE in alkaline silicate rocks are distributed among several mineral phases that are not equally processable. As for carbonatite-hosted REE–HFSE deposits, the presence of larger quantities of U and Th requires special consideration. In 2021, geological work on the world-class Kvanefjeld U-REE-Zn deposit in the Ilimmaasaq alkaline complex of southern Greenland came abruptly to an end, because new laws are banning exploration for and exploitation of uranium-enriched minerals. Consequently, ongoing research seeks to identify magmatic–hydrothermal processes by which deleterious radionuclides are naturally separated from the REE–HFSE commodities of interest.

As with carbonatite-related REE–HFSE deposits, supergene processes during weathering can upgrade primary mineralization hosted by alkaline silicate rocks (e.g., the Tantalus project in Madagascar; Figs. 1 and 4), which may reduce costs associated with ore processing. For example, the majority of global HREE production comes from supergene ion adsorption clay deposits in Southeast Asia. These important deposits formed via weathering of granites (including peralkaline varieties) and subsequent adsorption of REE onto the surfaces of clay minerals within the regolith. The two-dimensional nature of these only tens of meters thick regolith-hosted REE deposits, combined with their low cut-off grades, creates a greater risk of environmental damage compared with mining of spatially confined high-grade REE deposits hosted by alkaline silicate or carbonatite magmatic systems (Fig. 4).

Mineral exploration in alkaline silicate magmatic systems is relatively immature because demand for their contained REE–HFSE has historically been low. Many advanced resource projects investigate mineralized occurrences at surface, typically identified by governmental geological surveys. There are numerous large REE–HFSE mineral deposits at or near the Earth's surface, which differs from more mature commodities such as Cu, where easily-located large deposits are nearing depletion, and exploration is being pushed into more challenging environments such as under km-thick sedimentary cover. Future success in developing economic resources hosted by alkaline silicate rocks relies on an understanding of both the geological distribution and metallurgical processability of REE–HFSE mineralization under scrutiny. A mineral systems approach to exploration is recommended for this complex task, because the identified critical processes operating in nature can be linked to features mappable by geophysical and geochemical techniques, which helps to reduce the scale of investigation [18].

BOX-3 SOMEWHERE HERE *[text+figure, see the end of manuscript]*

5. Geopolitical impact

The mineral resources discussed in Section 4 are often at the root of geopolitical conflict.
For example, diamonds held strategic value over the course of several centuries, being closely tied to European colonialism and civil wars in Africa (e.g., 'blood' or 'conflict' diamonds). Arguably, many of us are stunned by the timeless beauty of a rough diamond; however, kimberlite-hosted diamond mining may phase out because synthetic diamond production has become more economic and leaves a smaller environmental footprint.

Given that the global transition to green energy technologies will be mineral-intensive when compared to a fossil fuel economy, an uninterrupted supply of critical metals – such as the rare-earth elements – is of paramount importance for sustainable development. It is therefore argued that the supply of critical raw materials should be governed transparently at the international level to coordinate the interests of all parties involved: from mining countries in the developing world through to manufacturers of zero-emission technologies in industrialized countries [25].

Many metals enriched by carbonatite and alkaline silicate magmas are critical to technologies that facilitate the green-energy-shift, such as neodymium magnets used in wind turbines and e-mobility. The assessment of criticality differs among countries and continents, and it represents a simplified snapshot of a highly dynamic socio-economic system, potentially with only limited predictive value. For example, Canada expanded its list of critical raw materials in 2024 now also including phosphorous (https://www.canada.ca/en/campaign/critical-minerals-in-canada.html), a food production essential commodity that is hosted by several large carbonatite complexes worldwide such as Sokli in Finland and Palabora in South Africa (Box 2). The European Commission included four new metals (As, Mn, Ni, Cu) into their 2023 list of critical raw materials (https://data.europa.eu/doi/10.2873/725585), while indium was removed from that list and is no longer considered to be 'critical'. This shows how over the course of only three years the supply risk or economic importance of some commodities have changed from the perspective of the European Union and its member states. This development impacts the exploration efforts to find new carbonatite complexes because Cu can be an integral part of their multi-commodity style mineralization (Section 4.2).

Scarcity of natural resources and resource nationalism are a recurring theme in the history of the industrialized world. A new development is, however, that the capacity to process and refine certain metals has been concentrated in only a few countries, which presently assert control over global markets. The global transition to a low-carbon renewable energy economy will be facilitated by the uninterrupted supply of critical metals. Mining of critical raw materials can be a positive contribution to sustainable development, but only where the entire resource supply chain adheres to the highest ethical and environmental standards. Unfortunately, critical metals mining and refinement are in many places responsible for environmental degradation and social inequality.

6. Conclusions

Low-degree partial melting of asthenospheric and lithospheric mantle sources in the presence of volatiles produces kimberlite, carbonatite and alkaline silicate magmas, mainly in continental settings. The rocks formed from these exotic magmas are among the most mineralogically diverse, which requires special definitions that deviate from IUGS classification schemes, as demonstrated in this Chapter (Box 1).

Whereas clear-cut genetic models can be formulated for kimberlites and alkaline silicate rocks, carbonatite magmas have highly diverse sources and follow complicated evolutionary paths, which hampers development of a unifying petrogenetic model. This situation is exemplified by the fact that carbonatites can have genetic relationships to both kimberlites in cratonic settings and alkaline silicate rocks in continental rifts and orogenic belts (Figs. 1 and 2).

Regardless of the exact origins of these deep-sourced magmatic systems, their high contained economic value is vastly out of proportion to their low eruption frequency and magma volumes (i.e., rare low-volume magmatism with large commercial benefits) (Fig. 4; Table 1). Kimberlites are the main source of natural diamonds but there is no direct genetic link to their precious mantle cargo. In contrast, the large resources of rare-earth elements, Nb and other commodities associated with carbonatite and alkaline silicate magmatism formed via prolonged igneous differentiation and hydrothermal overprinting (Box 2 and 3). At current global market conditions, many of the carbonatite-hosted deposits of critical metals (e.g., REE–Nb) are economic only because of upgrading via supergene processes in lateritic weathering profiles. Although exploration for carbonatite-hosted ore deposits is challenging, the excellent mine-to-occurrence ratio and the ever-growing demand for technology metals highlight the important role carbonatites will play during the green energy transition. A world with more zero-emission technology will be a world with increasing mining of critical raw materials from carbonatites and alkaline silicate rocks.

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References

1. Gill, J.C., *Geology and the Sustainable Development Goals.* Episodes, 2017. **40**(1): p. 70-76.
2. Jones, T.R., et al., *Accelerated carbonate biomineralisation of Venetia diamond mine coarse residue deposit (CRD) material: a field trial study.* Science of the Total Environment, 2023. **893**: p. 164853.
3. Yaxley, G.M., et al., *Carbonatites: classification, sources, evolution, and emplacement.* Annual Review of Earth and Planetary Sciences, 2022. **50**: p. 261-293.
4. Marks, M.A.W. and G. Markl, *A global review on agpaitic rocks.* Earth-Science Reviews, 2017. **173**: p. 229-258.
5. Russell, J.K., R.S.J. Sparks, and J.L. Kavanagh, *Kimberlite volcanology: transport, ascent, and eruption.* Elements, 2019. **15**(6): p. 405-410.
6. Lorenz, V. and S. Kurszlaukis, *Root zone processes in the phreatomagmatic pipe emplacement model and consequences for the evolution of maar-diatreme volcanoes.* Journal of Volcanology and Geothermal Research, 2007. **159**(1-3): p. 4-32.
7. Anenburg, M., S. Broom-Fendley, and W. Chen, *Formation of rare earth deposits in carbonatites.* Elements, 2021. **17**(5): p. 327-332.
8. Foley, S.F., G.M. Yaxley, and B.A. Kjarsgaard, *Kimberlites from source to surface: insights from experiments.* Elements, 2019. **15**(6): p. 393-398.
9. Giuliani, A., et al., *Kimberlite genesis from a common carbonate-rich primary melt modified by lithospheric mantle assimilation.* Science Advances, 2020. **6**(17): p. eaaz0424.
10. Tappe, S., et al., *Geodynamics of kimberlites on a cooling Earth: Clues to plate tectonic evolution and deep volatile cycles.* Earth and Planetary Science Letters, 2018. **484**: p. 1-14.
11. Tappe, S., et al., *The tungsten-182 record of kimberlites above the African superplume: Exploring links to the core-mantle boundary.* Earth and Planetary Science Letters, 2020. **547**: p. 116473.
12. Giuliani, A., et al., *Primordial neon and the deep mantle origin of kimberlites.* preprint, under review at Nature Portfolio, 10.21203/rs.3.rs-5046180/v1, 2024.
13. Schmidt, M.W., A. Giuliani, and S. Poli, *The origin of carbonatites: combining the rock record with available experimental constraints.* Journal of Petrology, 2024. **65**(10): p. egae105.
14. Pintér, Z., S.F. Foley, and G.M. Yaxley, *Diamonds, dunites, and metasomatic rocks formed by melt/rock reaction in craton roots.* Commun Earth Environ, 2022. **3**(296): p. 1-8.

15. Woolley, A.R. and B.A. Kjarsgaard, *Paragenetic types of carbonatite as indicated by the diversity and relative abundances of associated silicate rocks: evidence from a global database.* The Canadian Mineralogist, 2008. **46**: p. 741-752.
16. Walter, B.F., et al., *Fluids associated with carbonatitic magmatism: a critical review and implications for carbonatite magma ascent.* Earth-Science Reviews, 2021. **215**: p. 103509.
17. Smith, M.P., et al., *From mantle to critical zone: a review of large and giant sized deposits of the rare earth elements.* Geoscience Frontiers, 2016. **7**(3): p. 315-334.
18. Beard, C.D., et al., *Alkaline-silicate REE-HFSE systems.* Economic Geology, 2023. **118**(1): p. 177-208.
19. Hutchison, W., et al., *Mantle sources and magma evolution in Europe's largest rare earth element belt (Gardar Province, SW Greenland): new insights from sulfur isotopes.* Earth and Planetary Science Letters, 2021. **568**: p. 117034.
20. Anenburg, M. and Y. Liu, *A global marine sediment compilation and a cerium anomaly perspective on metasomatized mantle sources for REE-mineralized carbonatites.* Journal of Geophysical Research, 2024. **129**: p. e2023JB028546.
21. Keppler, H., *Water solubility in carbonatite melts.* American Mineralogist, 2003. **88**(11-12): p. 1822-1824.
22. Vasyukova, O.V. and A.E. Williams-Jones, *Carbonatite metasomatism, the key to unlocking the carbonatite-phoscorite-ultramafic rock paradox.* Chemical Geology, 2022. **602**: p. 120888.
23. Sokól, K., et al., *Quantifying metasomatic high-field-strength and rare-earth element transport from alkaline magmas.* Geology, 2022. **50**(3): p. 305-310.
24. Smith, E.M., et al., *Large gem diamonds from metallic liquid in Earth's deep mantle.* Science, 2016. **354**(6318): p. 1403-1405.
25. Ali, S.H., et al., *Mineral supply for sustainable development requires resource governance.* Nature, 2017. **543**(7645): p. 367-372.

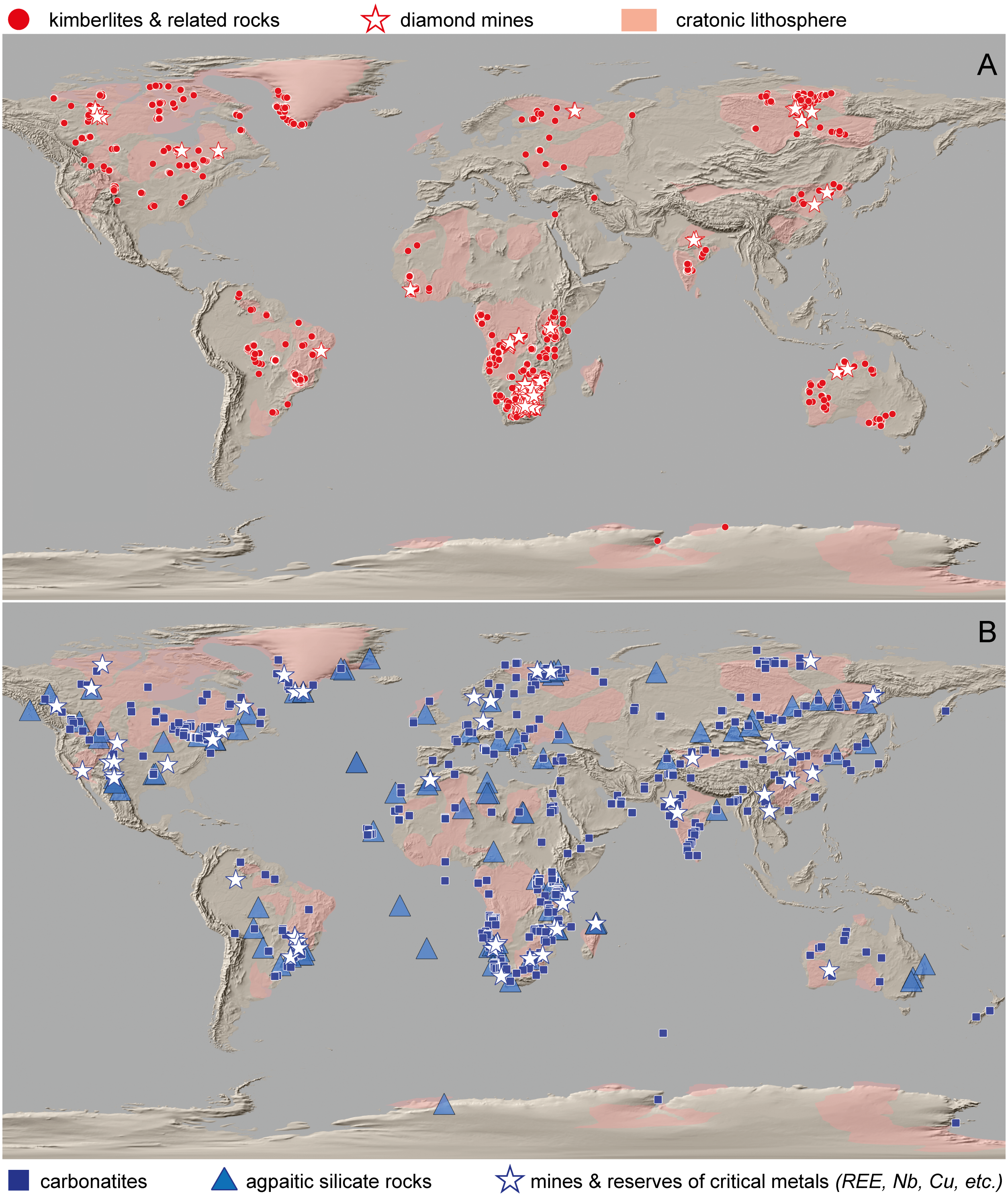


Figure 1: Global distribution of kimberlites, carbonatites and agpaitic silicate rocks superimposed on the ETOPO22 digital elevation model. Regions underlain by Archean cratonic lithosphere are coloured in red. **Panel A** shows kimberlites and related rocks (orangeite, aillikite, lamproite, etc.), plus active and past-producing diamond mines hosted by these volcanic rocks. **Panel B** shows carbonatites and agpaitic silicate rocks. Also shown are operating mines and significant deposits of rare-earth elements and niobium hosted by these exotic magmatic rocks. The dataset used to create these maps is provided in the Supplementary Information (10.17632/j42kxd4hnr.1).

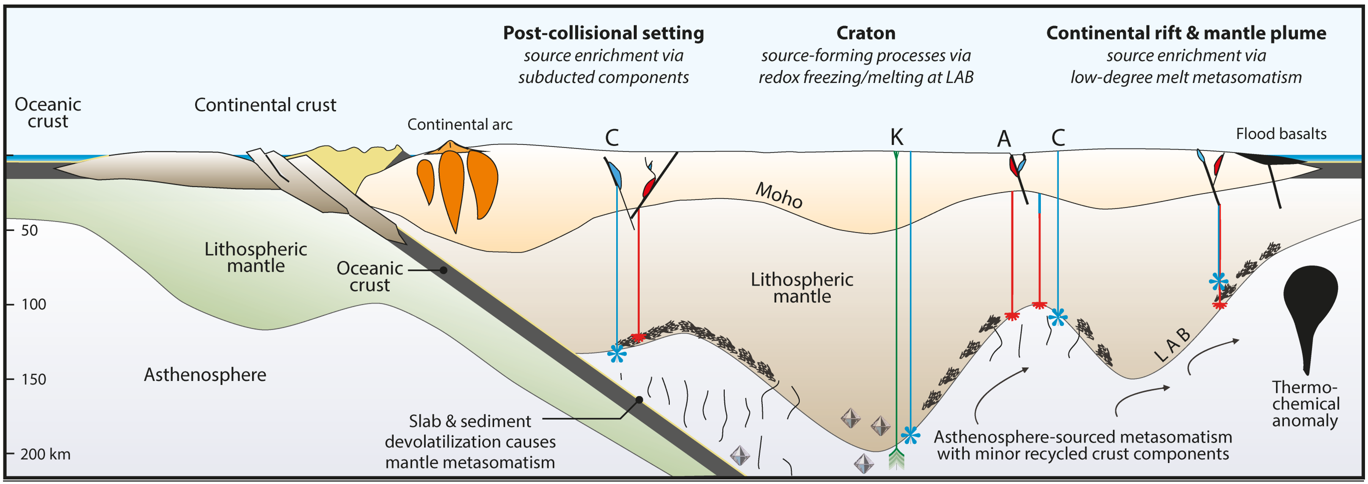


Figure 2: Continental scale cross-section summarizing the most common tectonic settings and mantle sources of kimberlite *(green)*, carbonatite *(blue)* and alkaline *(red)* magmatic systems. Moho – seismic Mohorovičić discontinuity equivalent to the crust–mantle boundary; LAB – lithosphere–asthenosphere boundary (modified from [18]).

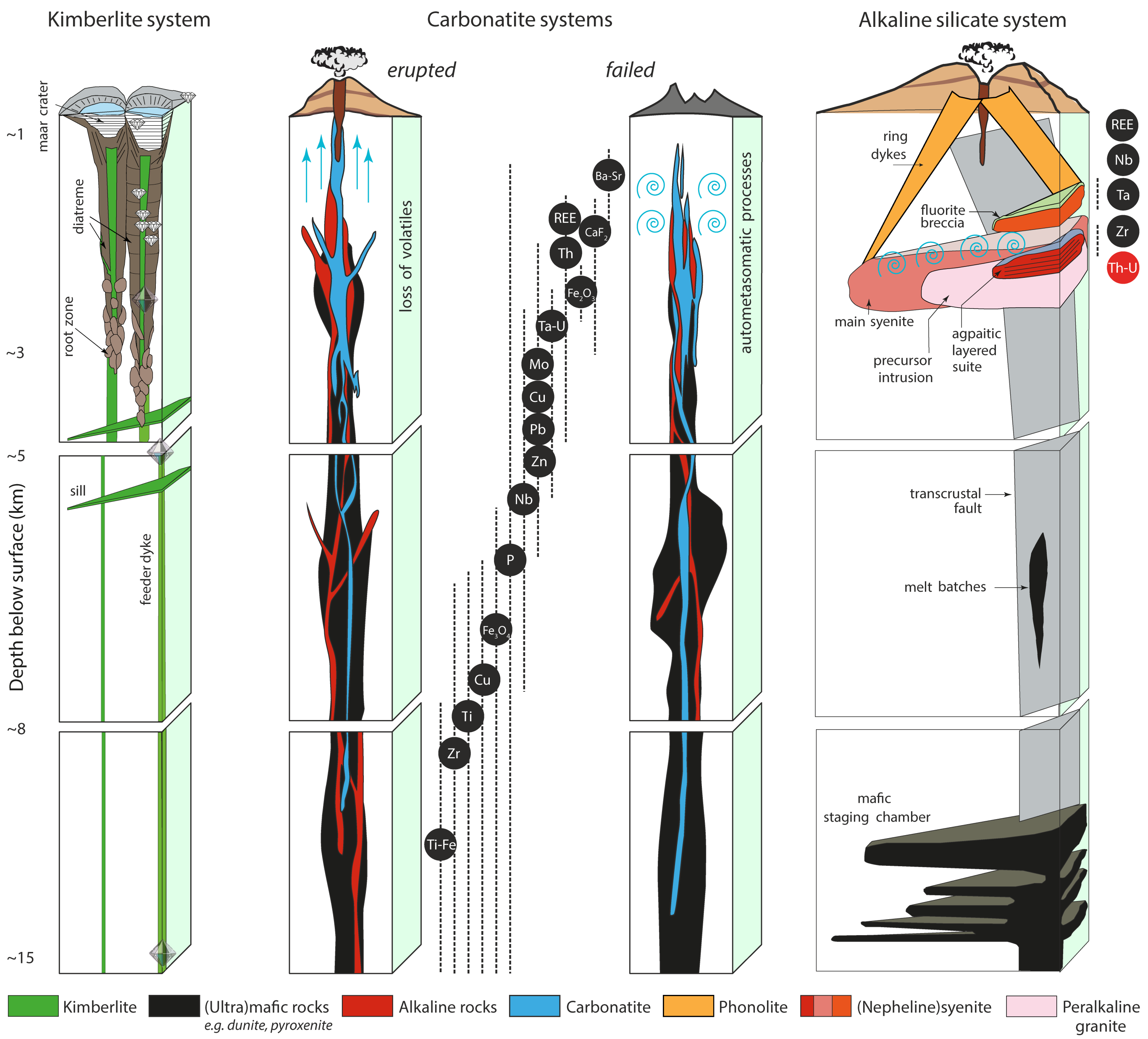
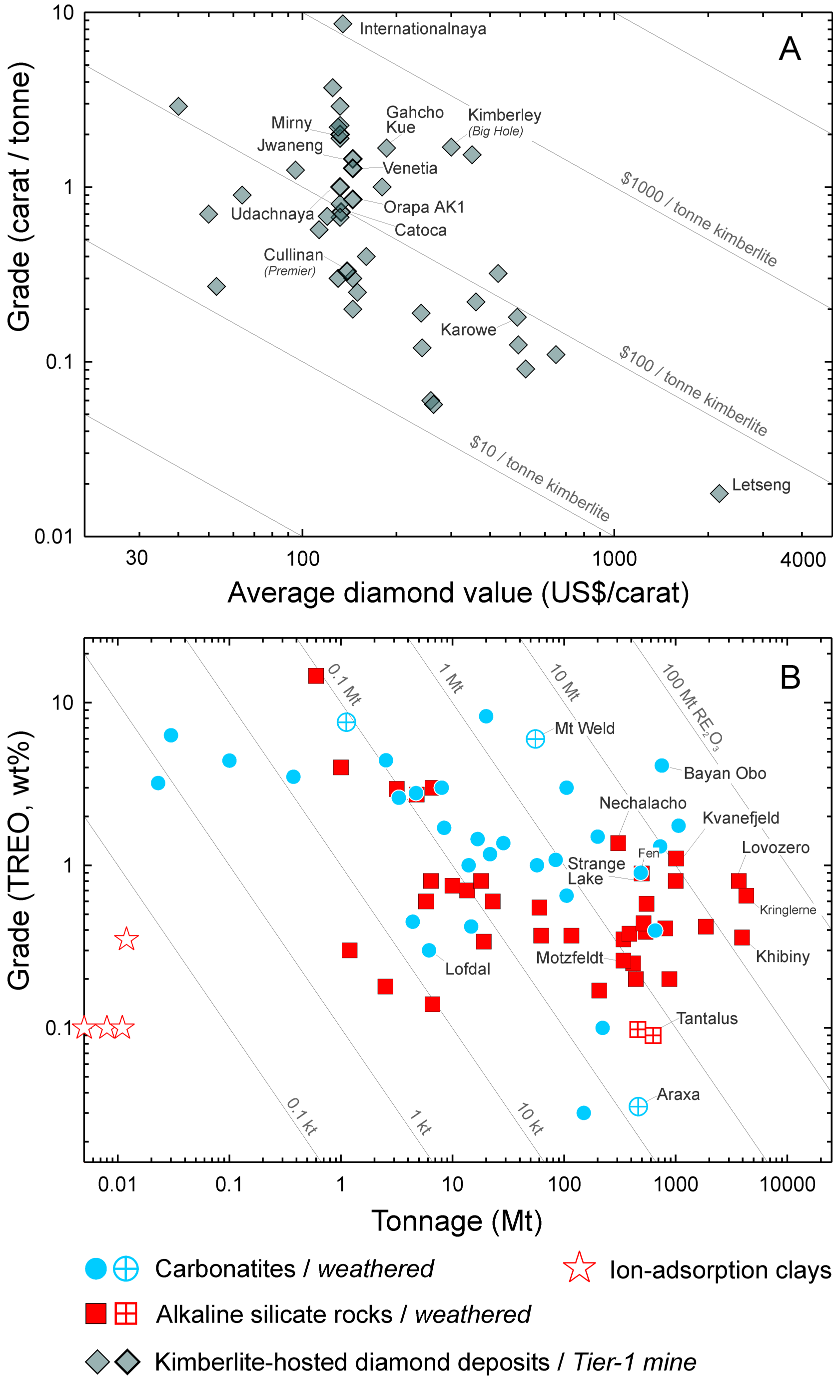
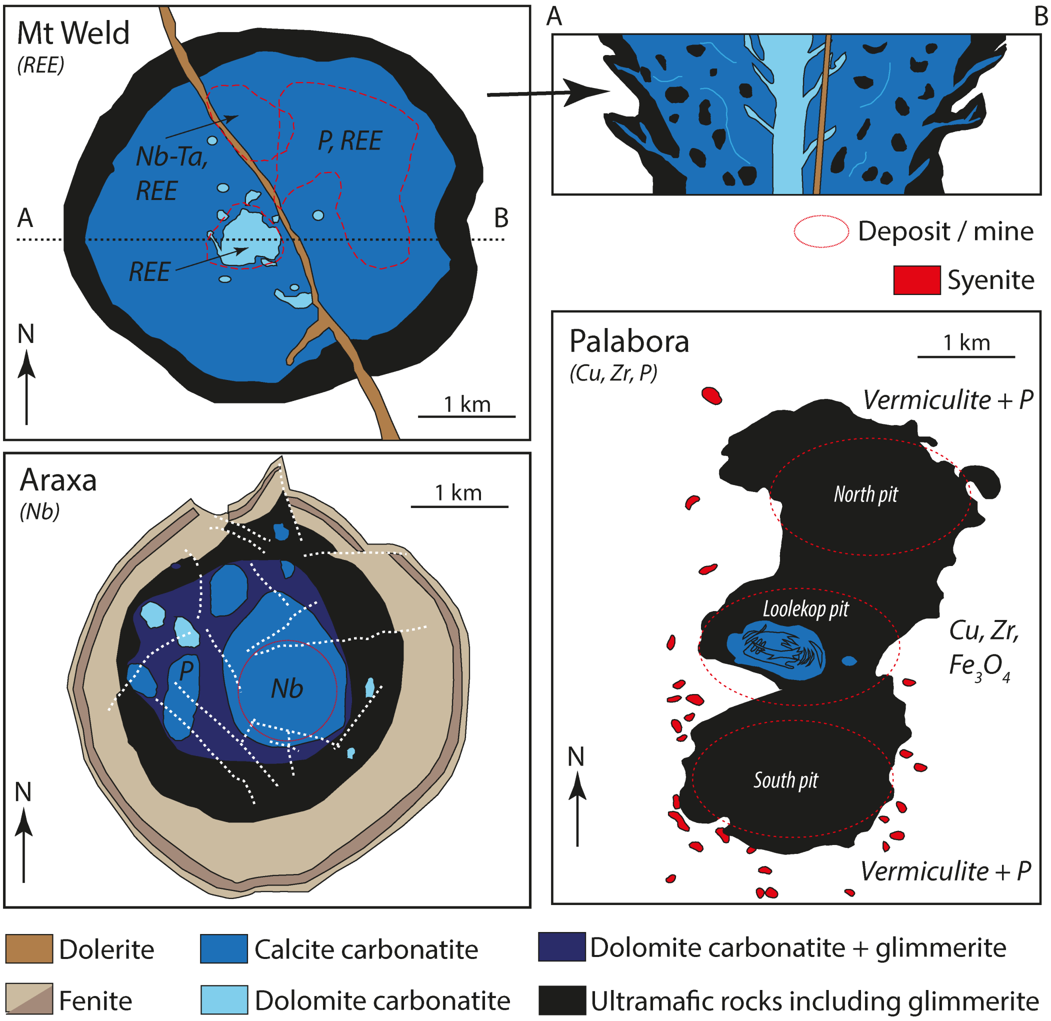


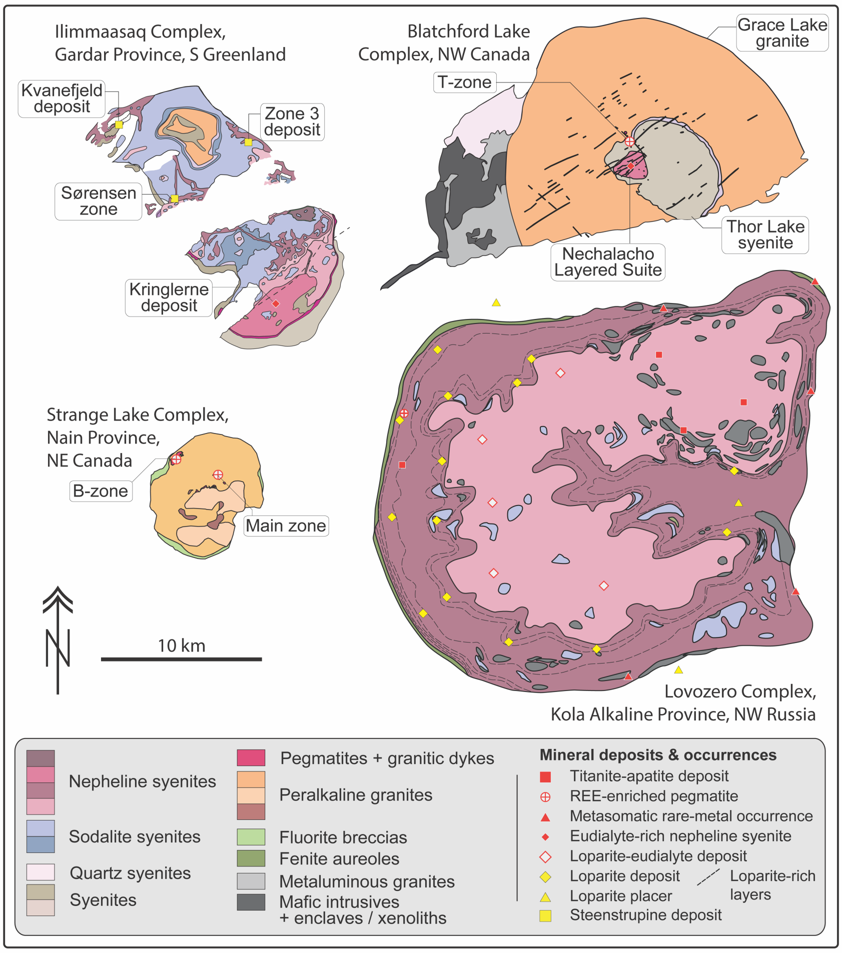
Figure 3: Comparison of the architectures of kimberlite, carbonatite and alkaline magmatic systems at crustal emplacement level. For kimberlites, a diamondiferous and a barren system are shown to occur in the same volcanic field. For carbonatites, an erupted and a failed system are shown, with implications for the type and style of economic mineralization. For the alkaline silicate magmatic system, the internal complexity is indicated by the wide variety of alkaline and peralkaline rock types, with localization of critical metal mineralization in the subvolcanic roof zone. The operation of autometasomatic processes leading to economic mineralization is indicated by the blue spiral symbol.



<u>**Figure 4:**</u> Selected aspects of the economics of kimberlite, carbonatite and alkaline magmatic systems. **Panel A** shows the diamond grade *versus* value for active and past-producing kimberlite-hosted diamond mines. The seven currently operating Tier-1 diamond mines are highlighted in bold. Their superlative character is clearly not defined by diamond grade and value, but rather by ore tonnage (not shown). The grey lines denote constant kimberlite ore value. **Panel B** shows the total rare-earth oxide grade (TREO) *versus* ore tonnage of global REE deposits hosted by carbonatite and alkaline silicate magmatic complexes [18]. Magmatic–hydrothermal deposits upgraded by weathering-related supergene processes as well as ion-adsorption clay deposits are shown for comparison. Grey lines denote the REE tenor, and localities labelled are discussed in this Chapter. The dataset used to create these diagrams is provided in the Supplementary Information (10.17632/j42kxd4hnr.1).



Box 2: Mt Weld in Australia, Araxa in Brazil and Palabora in South Africa are economically important carbonatite complexes, mined for REE *(Mt Weld)*, Nb *(Araxa)*, the technology-metal Zr, the base-metal Cu, P from apatite and Fe from magnetite *(Palabora)*. The three igneous complexes were emplaced at >5 km depth within the upper continental crust, and they comprise a wide variety of carbonatite types, alkaline silicate and ultramafic rocks such as glimmerites, clinopyroxenites, dunites and phoscorites. Open-pit mining of the critical raw materials takes place mainly in the supergene-enriched laterite horizons that developed on the primary mineralization. At Palabora an underground mine also operates. Mining of carbonatites and associated silicate rocks is challenging because the commodities of interest are irregularly distributed and typically confined to a single geological unit. For example, the new south pit at Palabora exploits vermiculite at surface and will eventually reach the base of laterization, where mining will continue to target Cu sulfides, apatite (P) and magnetite (Fe), similar to the currently operating Loolekop pit. It is important to note that, at current market conditions, mining of these three world-class deposits of critical raw materials would be uneconomic without the supergene enrichment (laterization) that is developed on top of the respective magmatic–hydrothermal ore bodies. A further benefit of laterization is the partial removal of radioactive Th and U, with positive implications for ore handling and processability.



Box 3: The Ilimmaasaq (S Greenland), Blatchford Lake (NW Canada), Strange Lake (NE Canada) and Lovozero (NW Russia) igneous complexes are notable examples of composite agpaitic silicate intrusions hosting significant resources of REE, Zr, Nb, Ta, U, Th and P. These alkaline intrusive bodies are characterized by world-class igneous layering, and ubiquitous alkali-metasomatism led to an even more diverse secondary REE–HFSE mineralogy.

The **Ilimmaasaq Complex** consists mainly of eudialyte nepheline syenites with two major mineral deposits. The Kringlerne Ta-Nb-REE-Zr deposit consists of a >250 m thick series of rhythmically layered floor cumulates comprising eudialyte-rich nepheline syenites (local name: *kakortokite*). The second deposit is a steenstrupine-sphalerite (U-REE-P-Zn) mineralization hosted by more evolved nepheline syenites (local name: *lujavrite*) that intruded the roof cumulates (Kvanefjeld, Sørensen zone and Zone 3). The floor cumulates at Ilimmaasaq resemble the **Nechalacho Layered Suite** (Tardiff deposit) within the Thor Lake syenite of the Blatchford Lake Complex, where REE–HFSE mineralization is hosted by magmatic eudialyte that has been hydrothermally replaced by zircon, bastnäsite, monazite, fergusonite and xenotime. A separate mineral deposit (T-zone) just north of Nechalacho consists of a bastnäsite zone underlain by a xenotime zone. Mining of the mineral deposits at Nechalacho began in 2021. The **Strange Lake peralkaline granite** hosts two Zr-REE-Nb-Be mineralized zones (Main- and B-zone), which are enriched in gadolinite, bastnäsite, allanite, kainosite, zircon and gittinsite. The layered **Lovozero Complex** is a giant peralkaline nepheline syenite intrusion that is actively exploited for Ti-Nb-LREE hosted by loparite (a perovskite group mineral) and HREE hosted by eudialyte.

Table 1: Comparisons of kimberlites, carbonatites and alkaline silicate rocks regarding their geology, petrology, volcanology, and economic potential.

	Kimberlites	**Carbonatites**	**Alkaline Silicate Rocks**
Tectonic Setting/s	cratons, craton margins	cratons, craton margins, continental rifts, orogens, oceanic plateaus	cratons, craton margins, continental rifts, orogens, oceanic plateaus
Pressure & Depth of Origin	5-10 GPa (165-330 km)	2-8 GPa (65-265 km)	2.5-6 GPa (80-200 km)
Temperature of Primary Melts	1100-1400°C	980-1100°C	1100-1200°C
Main Source Lithologies	peridotite	peridotite, eclogite	peridotite, pyroxenite, mica-amphibole-ilmenite metasomes
***Mantle Source Redox (fO_2)**	reduced or oxidized (IW-1 to IW+3)	oxidized (IW+3 to IW+10)	reduced or oxidized (IW+1 to IW+8)
Mantle Source Volatiles	H_2O, CO_2, CH_4, F, Cl, S	H_2O, CO_2, F, Cl, S	H_2O, CO_2, CH_4, F, Cl, S
Associated Rocks	aillikites, carbonatites	kimberlites, aillikites, melilitites, nephelinites, phonolites, ultramafic rocks, phoscorites	carbonatites, ultramafic rocks (e.g., pyroxenites), phoscorites, kimberlites, granites
Number of Occurrences	>5600	>600	>100 agpaitic silicate rock occurrences alone
Ages (oldest / youngest)	2.85 Ga / 12 ka	3 Ga / recent	2.75 Ga / recent
Emplacement Mode	dykes, sills, maar-diatremes (pipes)	dykes, stocks, ring complexes	dykes, ring complexes, layered intrusions
Exposed Surface Area / Diameter	0.05 to 3 km	0.05 to 5 km	0.1 to 40 km
Magma Volume (per incident)	ca. 0.2 km^3	>1 km^3	up to 2000 km^3 (i.e., LIP-related alkaline intrusions)
Commodities	diamond (gem & industrial)	REE, Nb, Ta, Cu, Fe, U, Th, P, F	REE, Zr, Hf, Nb, Ta, P, F, U, Be, Li
Number of Mines (active & past-producing)	>60	>40	>10

*IW+1 = log unit deviation from the iron-wüstite fO_2 buffering reaction.