



# The volcano–pluton interface; The Longonot (Kenya) and Kûngnât (Greenland) peralkaline complexes

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## ABSTRACT

An important step in studying volcanic processes is to consider the coeval processes in the subjacent magma reservoir(s). The trachytic caldera volcano Longonot (Kenya Rift Valley) and the Kûngnât syenite complex (Gardar province, Greenland) are taken to represent complementary magmatic systems, providing evidence of the volcanic and plutonic stages of evolution, respectively. The systems show many features in common; they have similar sizes, experienced two periods of caldera collapse, and were dominated by trachytic magmas, with smaller volumes of basic magma. Magmatic differentiation was dominantly by fractional crystallization of basaltic parents, with minor episodes of magma mixing and, at Kûngnât, some crustal contamination of parental basalts. A model is presented of a single, hypothetical trachytic centre, showing how evidence from one complex can be used to infer processes at the other. For example, an active convective system, with formation of wall and floor syenitic cumulates, can be inferred to exist in the Longonot magma chamber. At Kûngnât, the intermittent development of compositionally zoned caps to the magma chamber is postulated and the nature of syn-caldera eruptive activity is outlined.

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## 1. Introduction

Recent studies of salic magmatic systems have led to significant increases in our understanding of how such systems are initiated, evolve and die (Bachmann and Bergantz, 2004, 2008; Bachmann et al., 2002, 2007; Charlier et al., 2005; Hildreth, 2004; Hildreth and Wilson, 2007; Lipman, 1984, 2007; Metcalf, 2004). Models of magma chambers are available at all scales, from rhyolite dome fields such as the Coso Volcanic Field, California (Bacon et al., 1981; Manley and Bacon, 2000), the Mono Craters, California (Bursik and Sieh, 1989; Kelleher and Cameron, 1990) and the Olkaria complex, Kenya (Macdonald et al., 2008; Marshall et al., 2009), up to those that feed supervolcanoes (Cooper et al., 2012; Miller and Wark, 2008; Pamukcu et al., 2013). The complexity of petrogenetic processes producing compositional variation in the eruptive products is being clarified; different magmatic systems show different combinations of crystal fractionation, magma mixing, crustal assimilation and addition of new magma batches as major differentiation mechanisms. The majority of general models have been based on subalkaline silicic rocks, perhaps unsurprisingly since they constitute the largest proportion of silicic magmas and have erupted the largest volumes of materials. Peralkaline systems, in contrast, tend to be smaller and less

voluminous (Mahood, 1984). However, their peralkalinity and volatile-rich nature lead to petrogenetic processes which are arguably more complex than those in subalkaline systems (Macdonald, 2012). For example, Bachmann et al. (2007) suggested that in viscous magmas crystal–liquid separation is slow and less efficient than in mafic magmas, leaving textures that are not obviously cumulative, prompting Deering and Bachmann (2010) to raise the question, Do intermediate-silicic cumulate units actually exist or are they just difficult to identify? A major feature of the alkaline–peralkaline salic complexes of the Mesoproterozoic Gardar Province, SW Greenland, is spectacular layering whose formation has been related to low magma viscosities (Upton et al., 1996, 2013).

Peralkaline silicic systems seem almost invariably to be driven by input of basalt magma. Some at least are underlain by mafic magma chambers and/or concentrations of mafic dykes. During the volcanic stage of their evolution, silicic systems approach a steady state in which magma input is balanced by magma output over time-scales longer than those required for crystallization. A decline in heat input at some stage of the system's evolution might be related to the intensity of dyke intrusion beneath the complex, itself a function of decreasing rates of local extension (Bohrson and Reid, 1997; Macdonald, 2012). When the heat input to the system is lower than cooling-induced crystallization, the plutonic stage starts. Although the two stages are in many ways mutually dependent, crystallized plutons may contain no obvious record of a volcanic evolutionary phase.

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One approach to understanding the sub-volcanic processes in active systems is to study what can be inferred to be “fossil” magma chambers, i.e. high-level plutons. In this paper we focus on two continental rift-related complexes which show many petrographic, compositional and evolutionary features in common and which can be considered to be a complementary volcano–pluton pair. Although the histories of the two systems are not a perfect match, we draw particular attention to aspects of each complex which allow us a clearer picture of the structural development and petrogenesis of the other.

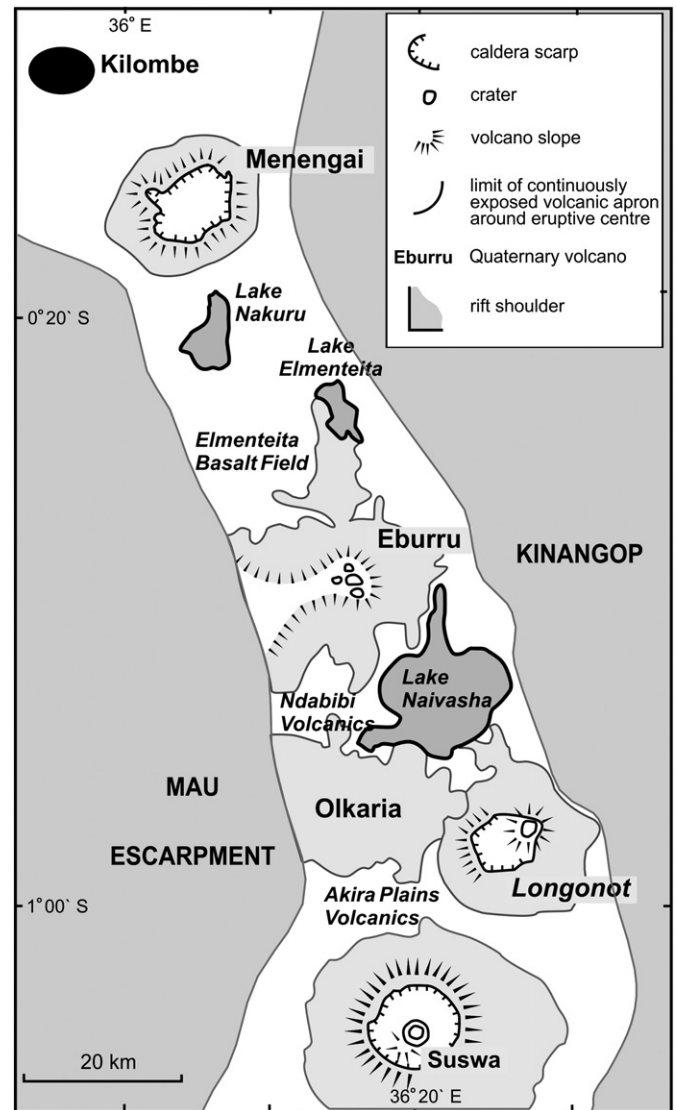
## 2. Geological setting

Peralkaline silicic complexes are located dominantly in extensional zones, especially in continental rift valleys, such as the East African Rift System (EARS). The youthfulness of the EARS, and the consequent lack of deep dissection, mean that little direct information is available on deeper structures. We expand here on suggestions by [Blundell \(1978\)](#), [Macdonald and Upton \(1993\)](#), [Upton \(1960, 2013\)](#), [Upton and Blundell \(1978\)](#) and [Upton et al. \(1990\)](#) that some features of the Mesoproterozoic Gardar rift zone of south Greenland may provide analogies with the EARS. Specifically, we compare the evolutionary histories, petrology and geochemistry of the Longonot trachytic caldera complex in the EARS and the Precambrian Kūngnāt syenite complex of Gardar age.

### 2.1. Longonot

Longonot lies in the axial part of the Kenya (Gregory) Rift and is part of a suite of five Quaternary–Recent, alkaline–peralkaline silicic volcanic complexes termed the Central Kenya peralkaline province (CKPP; [Fig. 1](#)) by [Macdonald and Scaillet \(2006\)](#). It lies close to the boundary between the Archaean Tanzanian craton and the Mozambique mobile belt, formed during the Pan-African orogenic event in the late Proterozoic ([Smith and Mosley, 1993](#)). The CKPP complexes lie in zones of crustal attenuation; extension in this part of the rift started at around 10 Ma and has stretched the crust and lithospheric mantle by ca. 10 km ( $\beta$ -factor of 1.1–1.2; [Mechie et al., 1997](#)), to its present thickness of ca. 20 km. Superimposed on a ca. –500 g.u. regional Bouguer anomaly over the EARS is a +190 g.u. gravity high centred on the rift axis, the maximum anomalies coinciding with the young (<0.5 Ma) central complexes. [Swain \(1992\)](#) modelled the axial high as the result of pervasive mafic dyke injection in the central 40 km of the rift, in which about 22–26% (to a width of 9–10 km) of the present-day crust down to a depth of 22 km consists of intruded material. [Dunkley et al. \(1993\)](#) and [Macdonald \(2012\)](#) have presented models of the axial central complexes overlying nodes in the dyke swarm.

The geology of Longonot has been described in detail by [Clarke et al. \(1990\)](#) and [Scott \(1980\)](#) and summarised by [Macdonald and Scaillet \(2006\)](#), [Rogers et al. \(2004\)](#) and [Scott and Skilling \(1999\)](#) ([Fig. 2](#)). Construction of a 25 km diameter composite volcano commenced <0.4 Ma ago, with the eruption of peralkaline trachyte lavas and pyroclastics ( $Lt^1$ ). The cone-building phase was terminated by incremental collapse to form a 7.5 km diameter caldera, off-set to the east of the original volcanic summit, at ~21,000 years BP. Caldera formation was accompanied by the eruption of ash and pumice falls and at least five trachytic ignimbrites ( $Lp^1$ – $Lp^4$ ) and promoted the interaction of the magmatic system with groundwater, generating a series of phreatomagmatic deposits. Post-caldera activity continued with a series of sporadic Plinian pumice lapilli beds ( $Lp^5$ ), the oldest of which has been dated at  $9150 \pm 150$  years BP by  $^{14}C$  ([Clarke et al., 1990](#)). Whilst they are found in several eruptive units, syenitic xenoliths occur most frequently in  $Lp^5$  ([Clarke et al., 1990](#); [Rogers et al., 2004](#); [Scott, 1980](#)). This pyroclastic phase ended by  $5650 \pm 120$  years BP and was followed by an abrupt shift to predominantly effusive eruptions, building a lava pile within the early caldera. The earliest lavas are trachyte–hawaiite mixed magma flows ( $Lmx^1$ ), followed by trachytes that formed the present Longonot cone



**Fig. 1.** Locality map of the five young (<0.5 Ma) complexes of the Central Kenya peralkaline province. Menengai, Longonot and Suswa, are caldera volcanoes, whereas Olkaria (the Greater Olkaria Volcanic Complex) and Eburru are rhyolitic dome fields. Kilombe is an older (2 Ma) trachytic caldera volcano. Modified from [Macdonald and Bagiński \(2009\)](#).

( $Lt^2$ ). Growth of the cone was terminated by the eruption of an ash fall deposit ( $Lp^8$ ), carbon dated at  $3280 \pm 120$  years BP ([Clarke et al., 1990](#)), followed by summit collapse to form a 2 km pit crater. Subsequent activity has included trachytic eruptions of trachyte–hawaiite mixed magma flows in the pit crater ( $Lmx^2$ ) and flank eruptions of trachyte ( $Lt^3$ ), the youngest of which may be <1000 years.

Longonot is undoubtedly an active centre. In addition to the youthfulness of the most recent eruptions, [Biggs et al. \(2009\)](#) detected geodetic activity there over the period 2004–2006 using Interferometric Synthetic Aperture Radar. The deformation was episodic and resulted in ~9 cm of uplift over a radius of 6.2 km. They modelled each episode as inflation of a horizontal lensoid at 4.1 km depth, which may represent activity in the volatile-rich cap to a magma chamber.

### 2.2. Kūngnāt

Kūngnāt is the westernmost and most northerly of the alkaline intrusive complexes forming the Mesoproterozoic Gardar province ([Fig. 3](#)). The overall setting is comparable to that of Longonot, in that it lies in the Border Zone between an Archaean craton and the

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