The crustal tongue melting model and the origin of massive anorthosites

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ABSTRACT

Recent detailed field studies in several anorthosite complexes have shown that anorthosites are frequently associated with weakness zones in the crust which may have favoured their emplacement at mid-crust levels. Recent experimental data have shown that the parent magma compositions of various anorthosite massifs lie on thermal highs in the relevant phase diagrams at 10–13 kbar, indicating that these magmas cannot be derived by fractionation of peridotitic mantle melts but by melting of gabbronoritic sources in the lower crust at 40–50 km depths. In the Sveconorwegian Province terrane boundaries have been traced in deep seismic profiles to Moho offsets or to tongues of lower crustal material underthrust to depths higher than 40 km. In Southern Norway, we suggest that a lithospheric-scale weakness zone (the Feda transition zone?) has channelled the Rogaland anorthosites through linear delamination, asthenospheric uprise and melting of a mafic lower crustal tongue.

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Introduction

Petrogenetic models for a number of igneous rocks such as anorogenic granites, continental flood basalts or Proterozoic AMCG [anorthosite, mangerite, charnockite (rapakivi-) granites] suites postulate the development of a magma chamber at the base of the crust by underplating or intraplating. Such models are convenient to account for the widespread geochemical and isotopic evidence for interaction between magmas and crustal material. A mantle origin for the ponding magma is generally accepted though its composition, varying from basaltic to picritic, is rather loosely constrained, as is the melting process invoked.

In the case of the AMCG plutonism, the discovery of high-Al orthopyroxene megacrysts (HAOM) in most massif-type anorthosites (Emslie, 1975) was the decisive step in building up a polybaric model for the genesis of anorthosites: a basaltic magma ponds in a deep-seated magma chamber at the base of the crust, where HAOM crystallize before the magma products intrude to mid-crustal levels (Emslie, 1978). The depth of the deep-seated chamber was constrained by experiments on the stability of 7–9% Al_2O_3 HAOM which proved to be a liquidus mineral between 11 and 13 kbar (Fram and Longhi, 1992).

The nature of the parent magma(s) of anorthosite suites has been discussed extensively, particularly in connection with the origin of related mangerites, charnockites and (rapakivi-) granites. Several magma types were invoked, each with convincing arguments: high-alumina noritic basalt, e.g. in the Harp Lake Complex (Emslie, 1978), troctolitic basalt (Nolan and Morse, 1986), and jotunitic (Fe-Ti-P-rich hypersthene monzodioritic) liquid, well represented in Rogaland (Demaiffe *et al.*, 1986; Duchesne, 1990; Vander Auwera and Longhi, 1994). Granitic melts produced by crustal anatexis (e.g. Emslie *et al.*, 1994) and/or by differentiation from intermediate magmas (Duchesne and Wilmart, 1997; Vander Auwera *et al.*, 1998) are voluminous. Contact anatexis around the deep-seated magma chamber and various amounts of contamination between mantle-derived magmas and crustal materials were invoked to explain the large variety of melts. A consensus model eventually emerged (Ashwal, 1993), in which 'plumes' or 'mantle swells' were generating the basaltic magma underplating the crust. The possibility of a mafic lower crustal source is rarely considered, even though radiogenic isotopes cannot differentiate easily between juvenile lower crustal and mantle sources. However, recent petrological experimental work (Longhi *et al.*, 1999) has pointed to a crustal gabbronoritic primary source for the parent magma of the anorthosite suite, thus questioning the classical mantle origin. The 'crustal tongue melting' model proposed here integrates these data as well as deep-seismic profiles recently interpreted by Andersson *et al.* (1996), and major lithospheric discontinuities in South Norway, to provide a constrained melting mechanism for anorthosite genesis.

Petrological constraints on the nature of the anorthosite source rocks

Experimental data on the liquidus equilibria relevant to the anorthosite petrogenesis in the range of 1 bar to 13 kbar have brought new constraints on the nature of the source rocks of the parental magmas (Longhi *et al.*, 1999). Between 10 and 13 kbar, i.e. the pressure conditions of a deep-seated magma chamber where HAOM coexist with plagioclase, the phase diagram (Fig. 1) shows a thermal barrier between the eutectic liq + plag + l-pyx + sil + cpx and the reaction point liq + plag + gar + opx + cpx (B in Fig. 1). A jotunitic parent magma in the Rogaland anorthosite complex, as represented by the Tjörn jotunite (TJ), sits close to the liq + plag + opx + cpx cotectic, where the curve crosses the plane of plag + opx + cpx compositions and necessarily forms a local thermal maximum on the liquidus surface. Consequently, TJ cannot be derived by fractionation of melt of olivine-dominated mantle because such melts reach plagioclase saturation at the reaction point (B) and then evolve down-temperature in the direction of the arrows away from TJ. TJ can only be produced by melting of a source rock whose composition lies in the plag + opx + cpx plane, a gabbronoritic

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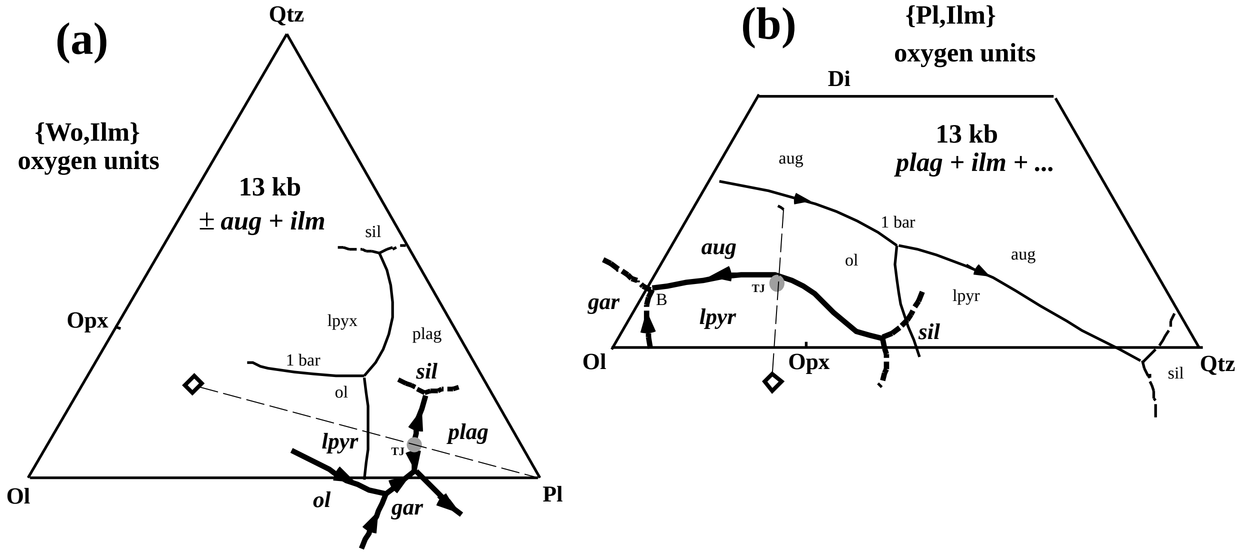

Fig. 1 Liquidus equilibria at 13 kbar projection into the system Ol (olivine)–Pl (plagioclase)–Qtz (quartz)–Wo (wollastonite). (a) projection from the wo (wollastonite), or (orthose) and ilm (ilmenite) components; (b) projection of plagioclase-saturated liquids from the pl (plagioclase), ilm and or components. Light solid lines show augite-saturated and plagioclase-saturated liquid boundaries at 1 bar calculated for TJ-like projection. Heavy lines show 13 kbar liquid boundaries. Arrows indicate direction of lowering temperature. Double arrows show peritectic lines. TJ (ellipse) is the composition of the jotunitic parent magma in the Rogaland province. Diamond is HAOM composition. Dashed lines connect coexisting phases. Abbreviations: lpyx, low-Ca pyroxene; sil, silica phase; plag, plagioclase; ol, olivine; opx, orthopyroxene; aug, clinopyroxene; gar, garnet (simplified from fig. 3 of Longhi *et al.*, 1999).

rock, which cannot be a mantle rock. Longhi *et al.* (1999) have also demonstrated that some high-alumina basaltic liquids, such as the parent magma of the Harp Lake Complex, show the same character as TJ despite higher Mg contents, and that more troctolitic melts, parental to various massifs such as Kiglapait or Laramie, plot on the liq + opx + cpx + plag cotectic at high pressure, and thus are also consistent with derivation from a mafic source.

These experiments have constrained the composition of the source rocks. Mafic granulites–a major component of the lower crust–have adequate average compositions, although they vary greatly between different tectonic provinces (Rudnick and Fountain, 1995). The best candidates are layered intrusions of basaltic kindred such as the Stillwater Complex which have higher mg-numbers than average lower crust.

Because of the thermal barrier at high pressure, mixing with silica-rich (crustal) material has no effect on the major element chemistry of liquids situated on the olivine side of the barrier. Whatever the amount of mixing, these liquids are inexorably brought back to the ol + opx cotectic by decreasing temperature and forced to follow it through the two reaction points and further (Fig. 1). Contamination by the crust of a basaltic magma in a deep-seated magma chamber is thus possible, especially at the radiogenic isotope level, but fractionation does not lead to silica-rich liquids.

Emplacement mechanism and structural setting

The diapiric uprise through the lower crust of a plagioclase crystal mush was initially proposed by Martignole and Schrijver (1970). It has been successfully simulated using finite-element modelling (Barnichon *et al.*, 1999). The model accounts for the shape of the diapir and its associated rim-syncline; the computed strain is consistent with field measurements, and the emplacement time (2–3 Myr) with geochronological data. The simulation considers an isotropic lower crust and the absence of a regional stress-field, conditions approximating those of anorogenic/within-plate settings.

Several anorthosite provinces, however, coincide with structural weaknesses in the lithosphere, which appear to have controlled and favoured the emplacement of diapirs. Indeed, the Nain anorthosite province straddles the limit between the Nain and Churchill (Rae) Provinces (Wiebe, 1992; Emslie *et al.*, 1994); the Lac St Jean and Havre–St. Pierre anorthosite complexes are associated with long lineaments (Higgins and van Breemen, 1992; van Breemen and Higgins, 1993); in the Laramie anorthosite complex two generations of anorthosites (separated by an interval of some 0.3 Gyr) are linked to a terrane boundary (Scoates and Chamberlain, 1997); the Suwalki anorthosite has emplaced in the Svecofennian platform along an EW lineament (Wiszniewska *et al.*, 1999).

In Southern Norway and Sweden, several major crustal lineaments have been recognized in the field and in geophysical profiles: from east to west, the Dalsland Boundary Thrust (DBT, Fig. 1), the Kristiansand–Porsgrunn shear zone (KPSZ, Fig. 1), which separates the Bamble and Telemark Provinces, and the Mandal–Ustaoset Line (MUL, Fig. 1) between Telemark and Rogaland–Vest Agder. Geophysical data indicate a Moho offset south-east of the Rogaland anorthosite complex (in ILP-11 profile; Andersson *et al.*, 1996). This offset is a major lithospheric structure and could correspond to the transition zone of Falkum (1998) ‘separating the south Rogaland anorthosite Province to the west and the Agder migmatitic terrane to the east’ and marked by the N–S elongated (100 km × 2–5 km) Feda augen gneiss displaying ubiquitous local shear zones (Falkum, 1998). Despite quite important late deformation linked to intrusion of late plutons, this transition zone is characterized by general N–S-striking and E-dipping foliation planes, with common strong subhorizontal N–S-orientated mineral lineations parallel to the fold axis (Falkum, 1998). This Feda weakness zone could be similar to the southern part of the Mandal–Utsaoset Line, which coincides with an augen gneiss and is straddled by numerous late granites (Sigmond, 1985) (Fig. 2). Both Feda and Mandal augen gneisses emplaced during the Sveconorwegian orogeny (1.05–1.03 Ga, Bingen and van Breemen, 1998a); the climax of their metamorphism has been dated by U–Pb on monazite at 1.02–1.00 Ga and a second phase at 0.930–0.925 Ga (Bingen and van Breemen, 1998b) is contemporaneous with Rogaland AMCG emplacement at 0.930–0.920 Ga (Schärer *et al.*, 1996).

Further east a number of postcollisional granites such as the Bohus and Flå granites (Fig. 2) have been related to major terrane boundaries (Andersson *et al.*, 1996). Although more accurate age determinations are needed to constrain the timing of this outburst of late plutonism, it appears that a major plate-scale structure channels granite emplacement. We suggest that large discontinuities, such as terrane boundaries or the Feda transition zone, also control the anorthosite emplacement during the same postcollisional event.

Link with major lithospheric structures: the crustal tongue melting model

Deep seismic profiles in the Skagerrak have been interpreted as reflecting Sveconorwegian structures by Andersson *et al.* (1996) (Fig. 3). At depth the Telemark–Bamble terrane boundary, i.e. the KPSZ, becomes a surface along which underthrusting has taken place. A slab of lower crust–the Telemark Craton Tongue–decoupled from the upper and mid crusts, can be identified at depths between 30 and 50 km in the downward prolongation of the KPSZ. Further east another terrane boundary, along which the Bohus and Flå granites are elongated, corresponds at depth with a Moho offset. Andersson *et al.* (1996) have suggested that the Bohus, Flå and other postcollisional granites in South Norway were produced by ana-

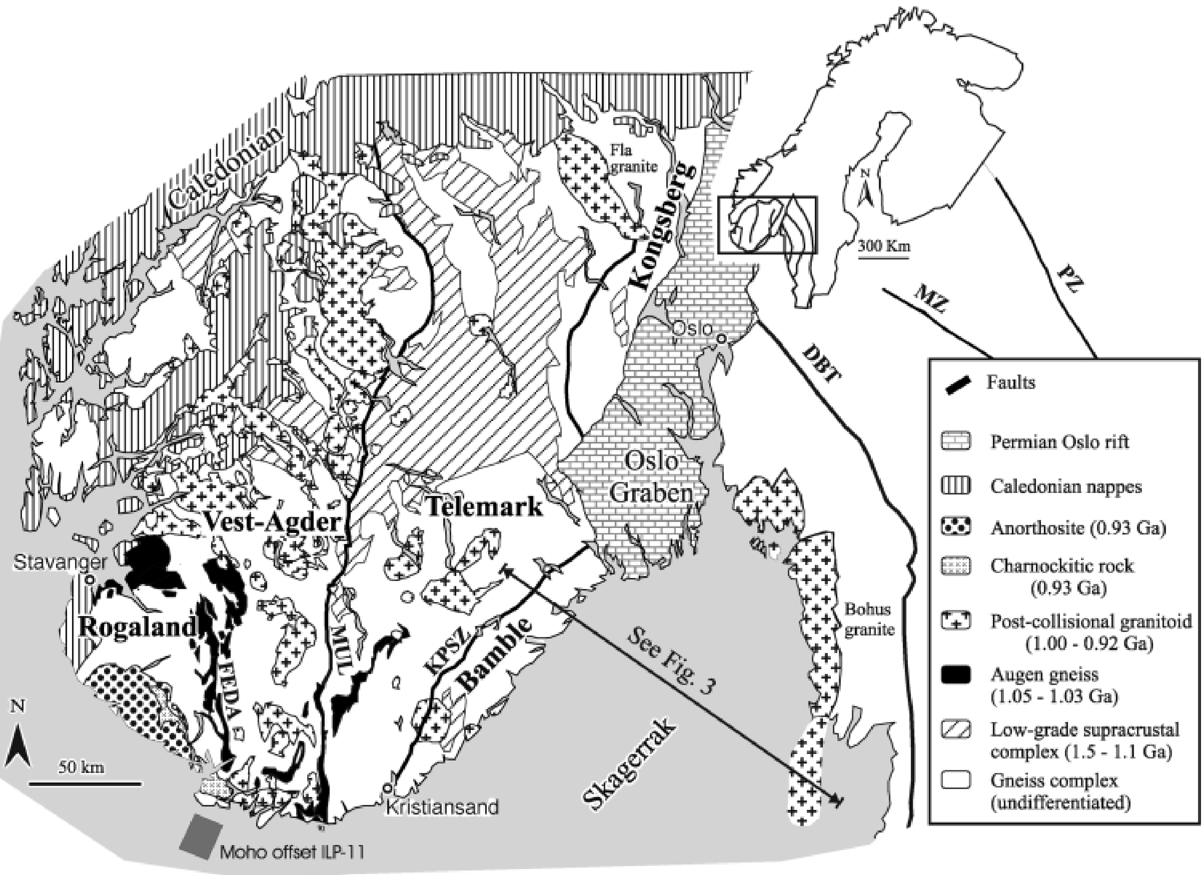

Fig. 2 The Sveconorwegian Province with major terrane boundaries: MUL, Mandal-Utsaoset line; KPSZ, Kristiansand-Porsgrunn shear zone; DBT, Dalsland boundary thrust; MZ, Mylonite zone; PZ, Protogine zone. The seaward extension of the Bohus granite and the Moho offset on deep seismic profile ILP-11 after Andersson *et al.*, 1996. Location of the profile across the Skagerrak shown on Fig. 3 is indicated by a heavy line.

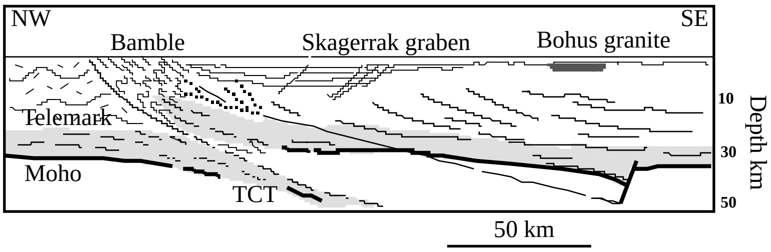

Fig. 3 Structural interpretation of the deep seismic profile under the Skagerrak and onshore extension. See Fig. 2 for location of the profile (after fig. 5 of Andersson *et al.*, 1996).

texis of underthrust mid-crustal material due to tectonic burial and consequent temperature increase.

We propose that a modified version of Andersson *et al.*'s model also accounts for the formation of anorthosites. Crustal tongues of lower mafic crust reach sufficiently high pressure by underthrusting to account for the stability pressure of HAOM (11–13 kbar corresponding to 40–50 km). After a thermal relaxation time of several tens of Myr (Platt and England, 1994) and linear lithospheric delamination along the shear zone allowing an asthenospheric uprise (Black and Liégeois, 1993), as is the case, e.g. under the San Andreas shear zone where the asthenosphere is close to Moho (Lachenbruch *et al.*, 1985), the temperature rises to produce jotunitic melts at *c.* 1200°C (Vander Auwera and Longhi, 1994) and high-alumina basalts at *c.* 1300°C (Fram and Longhi, 1992). By melting, the tongue becomes a magma chamber (Fig. 4) which evolves in a similar way as in the classical model, but with the variant that magmas and crystal mushes are channelled by crustal structures instead of rising vertically. Acidic rocks can be generated during the whole process both by differentiation of the mafic magma and/or by anatexis of the lower crust. Initial melting of a mafic lower crustal tongue should produce relatively small volumes of granitic melt, which subsequently may intrude to higher levels in the crust than the anorthosites or may be overtaken and passed by larger anorthositic diapirs with greater buoyant velocities. Granulitic residues of anatexis become gradually involved in the melting process and would give rise to some A-type or rapakivi-granites (Clemens *et al.*, 1986), although asthenosphere could also produce alkaline magmas (Liégeois *et al.*, 1998). In short, the proposed model accounts for the production of a large variety of magmatic material which can intrude over a wider time interval than the anorthosites themselves. The preservation of the Telemark craton tongue might possibly be due to a more refractory composition or to a thicker lithospheric mantle, which buffered the temperature rise and prevented its melting (Black and Liégeois, 1993). When the tongue melts, only a Moho offset is left.

Anorthosite formation can thus be directly linked to reactivation of large lithospheric discontinuities during the post-collisional period, as defined by Liégeois (1998). In a number of cases, however, the weakness zone is much older than the anorthosite emplacement, which suggests its reactivation. In Laramie, two generations of anorthosites were produced along the same limit: the first generation (1.76 Ga) is late-orogenic (post-collisional) and the second (1.43 Ga) anorogenic (Scoates and Chamberlain, 1997). In the Nain province, the suture seems to be some 0.4–0.5 Gyr older than the anorthosites produced 1.3 Ga (Emslie *et al.*, 1994) and another generation of anorthosites at ≈ 2.1 Ga has been identified along the boundary (Hamilton *et al.*, 1998). In Aïr (Niger, Sahara), the emplacement of the ≈ 0.41 Ga subvolcanic anorthosite (similar to massif-type) and associated alkaline ring-complexes (Moreau *et al.*, 1994) is linked to a reactivation of a ≈ 0.6 Ga shear zone separating two contrasted terranes (Liégeois *et al.*, 1994). Reactivation of a crustal underthrusting can trigger a new plutonism episode if another fragment of the crust is forced into the mantle or if an additional heat source develops as, e.g. through linear delamination along the shear zone.

The crustal underthrusting is also an elegant mechanism to transfer lower crustal mafic material to the mantle as a variant of the crustal foundering process (Arndt and Goldstein, 1989).

Concluding remarks

The crustal tongue melting model appears as an alternative to the plume model and provides a plausible mechanism to account for the crustal origin of anorthosite parent magmas. It emphasizes the role of deep lithospheric tectonic structures in producing a variety of magmas from granitic to high-alumina basaltic and in controlling their emplacement.

When produced as a direct consequence of collision, anorthosites can be classified as post-collisional because (i) the magmatism postdated only by some tens of Myr the collision climax, and (ii) the emplacement is controlled by structures directly resulting from the collision; when linked to a reactivation of a major structure, they belong to the within-plate/anorogenic magmatism but with the same geochemical signature.

The thermal and mechanical viability of the model now remains to be tested. We believe that the petrological and structural parameters are now sufficiently compelling to make these further investigations worthwhile.

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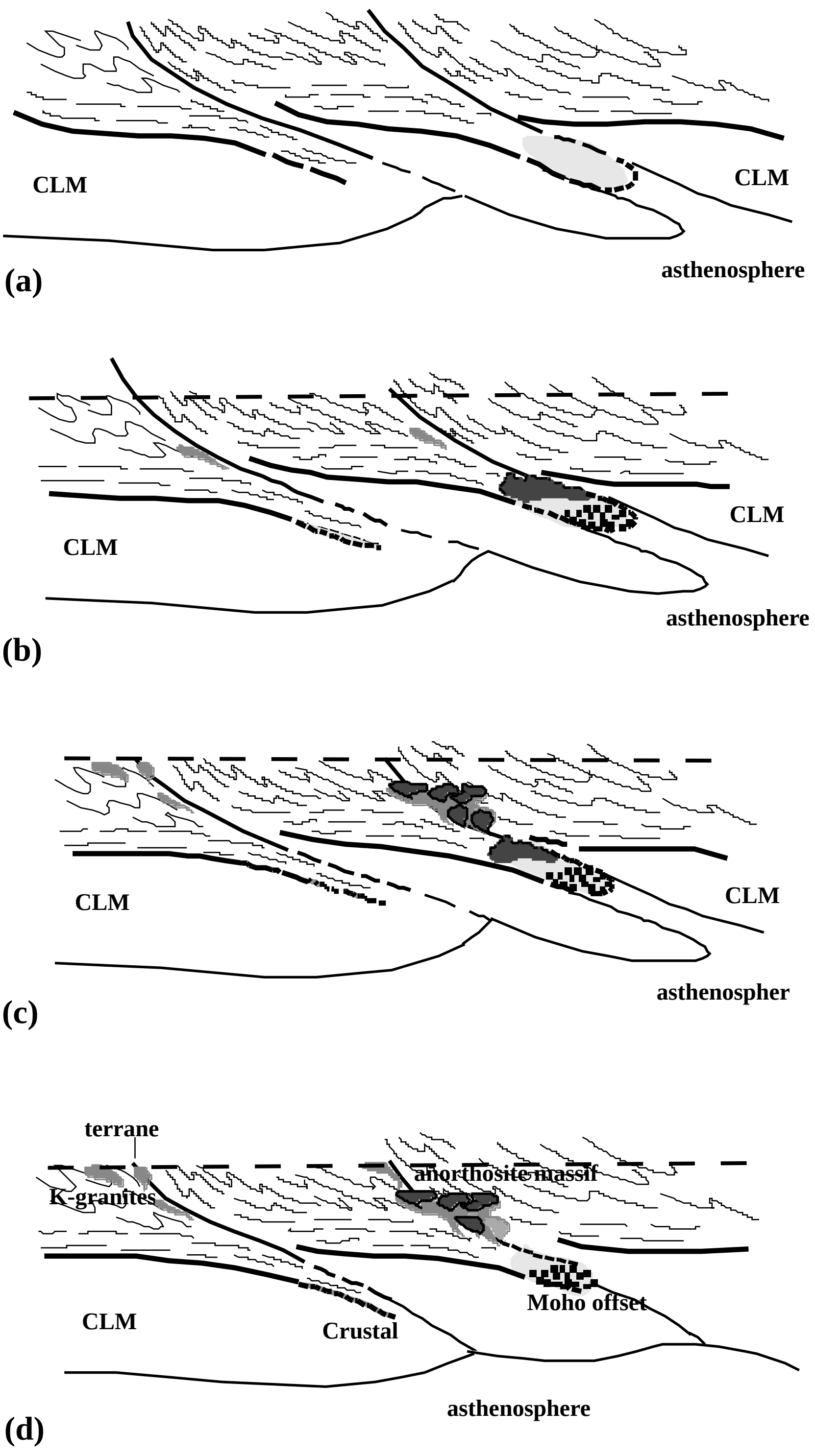

Fig. 4 (left) The crustal tongue melting model. (a) collision stacking of terranes produces underthrust lower crust tongues; (b) granitic liquids produced by anatexis of mid-crustal material intruded at higher levels along the terrane boundaries; due to linear delamination along the weakness zone, the rise in temperature melts a tongue of suitable composition some tens of Myr later, and a deep-seated magma chamber develops in which plagioclase floats to accumulate at the roof; a mafic cumulate sinks; (c) anorthosite diapirs rise through the crust, channelled by the weakness zone, and coalesce higher at mid-crust level to constitute a province of anorthosite massifs; granitic melts are also produced and follow the same path; (d) the mafic cumulate, left behind, becomes indistinguishable from the mantle; a Moho offset is the only scar of the former magma chamber. Abbreviation: CLM, continental lithospheric mantle.

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