Viscous collision in channel explains double domes in metamorphic core complexes

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ABSTRACT

In hot orogens, gneiss domes are a response to upper crustal stretching and lower crustal flow. Two-dimensional thermal-mechanical modeling shows that localization of extension in the upper crust triggers, in the deep crust, oppositely verging horizontal flows that converge beneath the extended region. Upon viscous collision, both flowing regions rotate upward to form two upright domes of foliation (double domes) separated by a steep median high-strain zone. In such systems, horizontal shortening in the infrastructure develops in an overall extensional setting. Dome material follows a complex depth-dependent strain history, from shearing in the deep crustal channel, to contraction upon viscous collision in the median high-strain zone, to extension upon advection into the shallow crust. This depth-dependent strain history is likely a general feature of dome evolution, and is arguably well preserved in double domes such as the Montagne Noire (France) and Naxos (Greece) gneiss domes.

INTRODUCTION

Gneiss domes form in a variety of tectonic settings and are characterized by a core of variably deformed gneiss, migmatite, and granitic rocks surrounded by a metamorphic envelope (e.g., Teyssier and Whitney, 2002; Whitney et al., 2004, and references therein). Deciphering fold-dominated, detachment-dominated, or gravity-dominated domes from their shapes and finite strain fields remains ambiguous (Burg et al., 2004). Physical experiments on diapirism have produced mushroom-shaped domes in which finite strain is characterized by vertical stretch (constriction strain) in the central part of the diapir, subhorizontal flattening in the roof, and near simple shear (plane strain) at the margins (e.g., Cruden, 1990). In the absence of far-field extension, buoyancy alone drives the flow, and the viscous drag at the diapir margins controls the fabric in the diapir. Although many studies have investigated the development of gneiss domes in extensional settings (e.g., Wdowinski and Axen, 1992; Tirel et al., 2004; Wijns et al., 2005, Gessner et al., 2007; Rey et al., 2009a), none of them has documented the finite strain in the domes. In the field, many gneiss domes reveal an internal architecture composed of two subdomes separated by a narrow, steeply dipping, median high-strain zone (MHSZ). Interestingly, double domes structure separated by a MHSZ can be inferred in some numerical experiments (Tirel et al., 2004; Wijns et al., 2005; Gessner et al., 2007). Examples of double domes include the Montagne Noire (southern France), Naxos (Cyclades, Aegean Sea, Greece), Entia (central Australia; Whitney et al., 2004), Aston-Hopitalet (Pyrénées; Denèle et al., 2007), and possibly Thor-Odin (Britich Columbia, Canada; Whitney et al., 2004). In some domes, the steeply dipping MHSZ has been interpreted in terms of diapiric ascent of migmatites and granite (Jansen and Schuiling, 1976; Vanderhaeghe, 2004), horizontal shortening and folding (Jolivet et al., 2004; Malavieille, 2010), or a combination of buoyancy-driven flow in migmatites and regional tectonics (Kruckenberg et al., 2010). Here, we show that in an extensional setting, double domes separated by a MHSZ form as a result of the viscous collision within a weak channel below the site of upper crustal necking. Although we apply a steady far field extension velocity, the tectonic regime is highly variable in space and time, with coeval extension in the upper crust and contraction in the lower crust.

We apply this concept to the evolution of the Montagne Noire and Naxos domes, where the double domes structure is well preserved.

NUMERICAL EXPERIMENTS, CODE, AND MODEL SETUP

A series of two-dimensional (2-D) numerical experiments were carried out to explore the flow of deep crust during extension. These experiments focus on the strain field within and around domes, and build on earlier work that examined the thermal and mechanical implications of the role of partial melting in the development of metamorphic core complexes (Rey et al., 2009a, 2009b). The model is 360 km long and 120 km deep (Fig. 1). It includes a 60-km-thick crust above 45 km of mantle and below a 15-km-thick layer of air. To facilitate comparison among experiments, we include a fault-shaped weak element in the upper crust to force the development of a dome at the center of the model. Without this anomaly, the location of faulting changes from one experiment to the next. Circular passive markers are distributed in the ductile crust below the fault-like anomaly to track strain (Fig. 1). During deformation these markers are transformed into ellipses, which record flattening and stretching directions, as well as strain intensity (through the ratio long/short axes).

Crust and mantle have a visco-plastic rheology with a pressure-dependent yielding, which includes strain weakening, and a temperature, stress, strain-rate and melt-dependent viscosity based on quartz-rich rock rheology for the crust and olivine for the mantle (cf. the GSA Data Reposi-
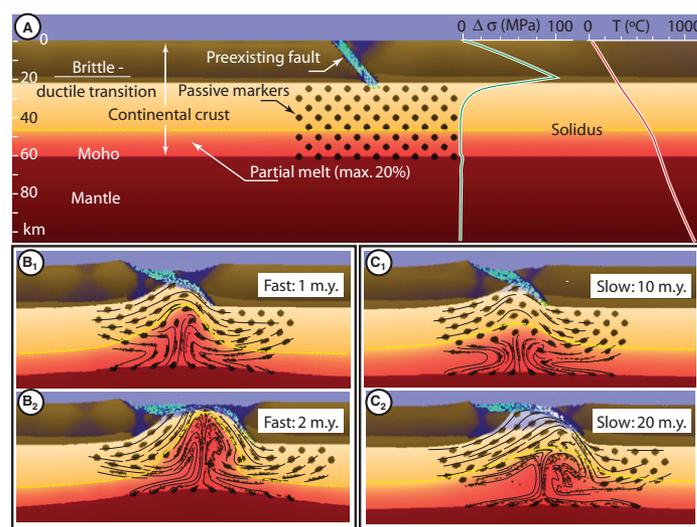
Figure 1. Numerical experiments of extensional gneiss domes. A: Model setup. B_{1-2}: Fast extension, both vertical walls move at 11.3 mm/yr. C_{1-2}: Slow extension, both vertical walls move at 1.13 mm/yr. Experiments B_1 and C_1 have recorded the same amount of extension, same for B_2 and C_2. Solid lines through the strain markers represent foliation trajectories.

[1]GSA Data Repository item 2011124, supplemental information on the model, is available online at www.geosociety.org/pubs/ft2011.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

tory[1]). The geotherm is controlled by radiogenic heat production, basal heat flow, and constant surface temperature. Crustal solidus and liquidus are adjusted to obtain a maximum melt fraction of 20% at the Moho at the onset of extension (see the Data Repository). The viscosity decreases linearly by 3 orders of magnitude when the melt fraction increases from 15% to 25% (critical melt fraction). Rosenberg and Handy (2005) showed that significant weakening occurs at 7% melt fraction. We have verified that our results are not affected by lower critical melt fractions down to 2%–12%. Below the solidus, the density of the continental crust is 2720 kg m^{-3} and is temperature independent. Between the solidus and liquidus the density of the continental crust decreases linearly by 13% (Clemens and Droop, 1998). In the models, there is no segregation of the melt from its source, a reasonable approximation for many migmatite-cored metamorphic core complexes, in which melt and solid fractions move en masse (Teyssier and Whitney, 2002). Boundary conditions involve free slip on all sides of the model. A velocity condition is attached to both vertical walls to drive extension at a prescribed velocity. In experiments exploring fast extension, a velocity of 11.3 mm/yr on both sides of the model drives extension (2 × 10^{-15} s^{-1} average strain rate; about an order of magnitude faster in the localized deformation zone). This velocity is reduced to 1.13 mm/yr in experiments exploring slow extension. We use Ellipsis, a Lagrangian integration point finite-element code to solve the governing equations of momentum, mass, and energy in incompressible flow (Moresi et al., 2003).

RESULTS

Fast Extension

During extension, the upper crust stretches and thins while the fault/detachment system accommodates the upward flow of the ductile crust (Fig. 1B$_{1-2}$). The flow in the lower crust is driven by the horizontal pressure gradients created by the localized thinning in the upper crust. Passive strain markers reveal the formation of an overarching dome, visible in the shallow crust (Fig. 1B$_1$), enveloping two subdomes separated by a MHSZ. This structure is built through the viscous collision and upward rotation of oppositely verging horizontal flows that converge below the region of upper crustal necking. Both the solidus and the brittle-ductile transition are advected upward (geometry akin to metamorphic core complexes) (Fig. 1B$_2$). Strain markers show that the double domes and the overarching dome are domes of foliation that form synchronously during extension.

Changing tectonic regime with depth is prominent in our experiments. In the deep crust, intense horizontal contraction owing to viscous collision results in steeply dipping fabrics. In contrast in the overarching dome, shallowly dipping fabrics and horizontal extension dominate. As extension proceeds, material from the deep crust is progressively advected through the contraction/extension transition. In fast extension experiments, this transition moves progressively closer to the surface (Fig. 1B$_{1-2}$).

Slow Extension

Reducing the velocity by one order of magnitude impedes the exhumation of the solidus and slows the formation of subdomes (Fig. 1C$_1$). As extension proceeds, the upward flow velocity in the partially molten crust is faster than the upward flow of the subsolidus region above. This velocity gradient is accommodated in the region above and below the solidus by horizontal flattening and folding of the vertical fabric developed in the MHSZ and in the core of the subdomes (Fig. 1C$_2$). Above this region, the material moves progressively into the shallow crust where horizontal extension dominates. In the partially melted region, the subdomes are dragged progressively into an asymmetric convective motion (Fig. 1C$_2$). This asymmetry is forced by the pressure gradient related to the detachment in the brittle crust. The main differences compared to faster extension are: 1) the two subdomes are deeper and remain buried under the overarching dome, 2) the steep fabrics in subdomes and in the MHSZ are

refolded, 3) the extension-contraction transition zone is deeper, and 4) the strain field in the subdomes grows progressively more asymmetric.

Role of Buoyancy Versus Isostasy

To assess the role of buoyancy, we compared a suite of slow extension experiments with enhanced melting (37% melt fraction at the onset of extension) and enhanced melt buoyancy (density of melt 16% lower than non-melted rock, Fig. 2A), to a suite of fast extension experiments in which melt has no buoyancy (Fig. 2B). In addition, we ran a suite of fast extension experiments in which extension proceeds until a dome is initiated (Fig. 2C$_1$) and is then allowed to evolve under fixed boundary conditions (Fig. 2C$_2$). In experiments with enhanced melting and melt buoyancy, buoyancy-driven convection controls the flow in the partially molten crust and a single dome develops, the core of which shows no organized fabric (Fig. 2A; see the Data Repository for a Rayleigh Number analysis). The double domes structure is best developed in experiments in which the melt has no buoyancy and therefore in which the potential for convection is nil (Fig. 2B). In this case, pressure gradients alone drive upward flow (Wdowinski and Axen, 1992). When buoyancy is involved (Fig. 1B$_2$), the solidus rises slightly faster and strain is localized more strongly compared to the experiment in which buoyancy is disregarded (Fig. 2B). This is because there is a positive feedback between exhumation of the partially molten region (driven by extension) and upward heat advection and softening of the overlying rocks, which favor strain localization (Teyssier and Whitney, 2002). When extension stops, buoyancy forces have limited effect on the structure of the upper section of the double domes (Fig. 2C$_1$) because cooling in the crust freezes the partially molten region faster than it can rise and deform under the buoyancy push. In contrast, in the deeper section of the domes, buoyancy forces drive convective motion, which progressively erases the internal structure in double domes (Fig. 2C$_2$). Overall, buoyancy forces have only a significant impact on domes development when extensional strain rates are low and/or when melt buoyancy is large. In all experiments preserving a double domes structure, where buoyancy forces play a minor role, extension rates control the dome exhumation rates.

FIELD EXAMPLES

The Montagne Noire (French Massif Central) and the Naxos gneiss domes (Cyclades Islands, Greece) (Fig. 3) have been variously interpreted in terms of diapirism (e.g., Gèze, 1949; Jansen and Schuiling, 1976; Beaud, 1985; Soula et al., 2001; Vanderhaeghe, 2004), contractional tectonics (Arthaud et al., 1966; Urai et al., 1990; Buick, 1991; Mattauer
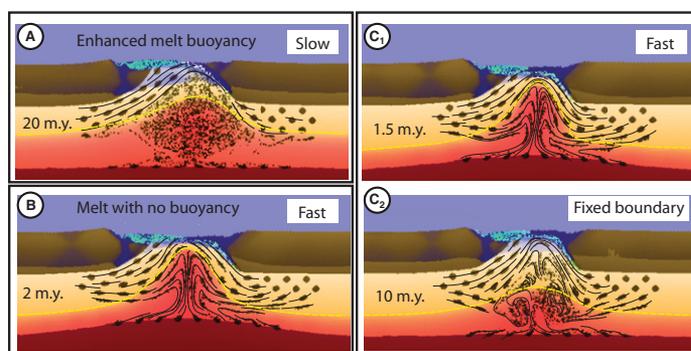
Figure 2. Numerical experiments documenting the role of buoyancy forces on the development of double domes. A: Experiment with enhanced melt fraction and melt buoyancy. B: Experiment involving melt with no buoyancy. C$_{1-2}$: Experiment in which extension lasted for 1.5 m.y., driven by a velocity of 11.3 mm/yr on both vertical walls. C$_1$: Snapshot at the end of extension period. C$_2$: Snapshot after 10 m.y. of evolution under fixed boundary conditions.
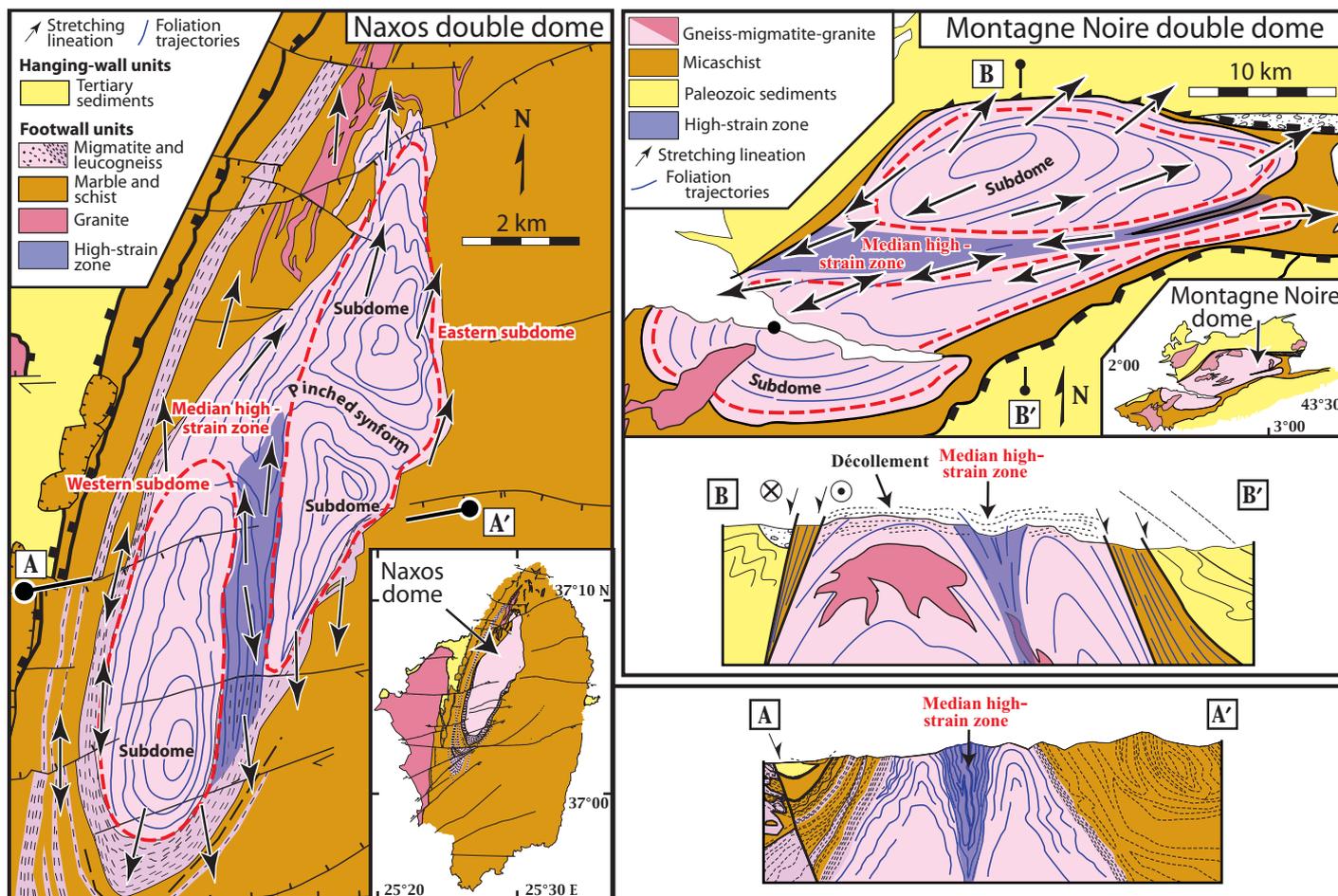
Figure 3. Simplified geological maps and cross-sections of the Montagne Noire dome and the Naxos dome showing their double domes internal structures.

et al., 1996; Aerden and Malavieille, 1999; Avigad et al., 2001; Malavieille, 2010), emplacement in a pull-apart setting (e.g., Nicolas et al., 1977; Franke et al., 2010), or as extensional metamorphic core complexes (e.g., Echtler and Malavieille, 1990; Van den Driessche and Brun, 1992; Gautier et al., 1993). Both domes are characterized by double domes of foliation (Fig. 3) separated by steeply dipping MHSZ (Bouchardon et al., 1979; Kruckenberg et al., 2010). In 2-D experiments (i.e., plane strain deformation), deep crustal flow and extension are bound to have the same direction, but not necessarily in 3-D. In nature, the 3-D geometry of flow necessarily impacts on finite strain fields in gneiss domes. Indeed in both Naxos and the Montagne Noire, the MHSZ is subparallel to the direction of extension not perpendicular as suggested in our 2-D experiments. This is because the MHSZ forms perpendicularly to the direction of flow in the lower crust (i.e., perpendicular to the direction of decreasing pressure) irrespective of the direction of extension. In our 2-D experiments, syn-melting vertical flattening develops in the subdomes and the MHSZ, whereas subhorizontal flattening forms in the region above. In 3-D, however, horizontal contraction in the partially molten subdomes and the MHSZ can led to out-of-plane lateral flow and subhorizontal stretching. In the Montagne Noire and Naxos, such a 3-D flow may explain the large variability—from steeply dipping to subhorizontal—of the plunge of the lineation in the MHSZ, as well as the large variability of the plunge of the lineation in the subdomes (Charles et al., 2009; Kruckenberg et al., 2010). In the subdomes of the Montagne Noire, syn-melting constriction with stretching lineations and the axes of folds subparallel to the subdomes' long axis (Brunel and Lansigu, 1997; Charles et al., 2009) may reflect out-

of-plane lateral flow in response to viscous collision in the deep channel. In 2-D experiments where melt buoyancy is important, folding of pre-existing vertical fabrics can be observed as shown in Figure $1C_2$. Infolding in Naxos eastern subdome (Kruckenberg et al., 2010) may reflect local variation of melt fraction and melt buoyancy, which promote intra-sub-dome folding such as those observed in Figure $1C_2$. The roof of both the Montagne Noire and Naxos gneiss domes is sheared by a gently dipping detachment that carries a sillimanite-bearing stretching lineation parallel to the direction of motion of the upper crust (Van den Driessche and Brun, 1992; Franke et al., 2010; Avigad et al., 2001). This is consistent with our 2-D experiments, in which deeper, steeply dipping fabrics are progressively rotated into horizontal as they are advected across the contraction/extension transition (Figs. 1B−1C). Hence, as our experiments suggest, the tectonic regime is strongly depth-dependent.

CONCLUSIONS

Double domes and median high-strain zones may result from convergence, viscous collision, and upward rotation of horizontal deep crustal flows during crustal extension. Foliation trajectories rotate from subhorizontal in the channel, to vertical upward in the core of the subdomes, and back to horizontal in the upper part of the domes. Horizontal flow in the deep weak channel is driven by the pressure gradient that results from localized thinning of the upper brittle crust. Upward flow in the subdomes is the result of a combination of buoyancy and pressure gradient, which dominate at low melt fraction and/or low melt buoyancy. Horizontal flow in the upper section of the domes is driven partly by the dragging effect

of the upper crust, and partly by the upward motion of the dome itself. The orientation of the median high-strain zone between the double domes could be used to infer the direction of convergent flow in the lower crust. In the domes, rocks are advected through two major rheological transitions (the solidus and the ductile to brittle transition) and a major tectonic regime transition (from contraction to extension).

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