The effects of lower crustal strength and preexisting mid-crustal shear zones on the formation of continental core complexes and low-angle normal faults

Guangliang Wu^{1,2} and Luc L. Lavier^{1,2}

¹ Institute for Geophysics, University of Texas, Austin, TX, United States.

² Department of Geological Sciences, University of Texas, Austin, TX, United States.

Correspondence to: Guangliang Wu (glwu@utexas.edu)

Key points:

- The origin of core complexes is proposed based on numerical models.
- The new models are consistent with regional geology and geophysics in the US Cordillera and the Aegean.
- Tectonic significance and implications of core complexes are discussed in the context of US Cordillera and the Aegean.

This is the author manuscript accepted for publication and has undergone full peer review but has not been through the copyediting, typesetting, pagination and proofreading process, which may lead to differences between this version and the [Version of Record](#). Please cite this article as doi: [10.1002/2016TC004245](https://doi.org/10.1002/2016TC004245)

Abstract

To investigate the formation of core complexes and low-angle normal faults, we devise thermo-mechanical simulations on a simplified wedge-like orogenic hinterland that has initial topography, Moho relief, and a preexisting mid-crustal shear zone that can accommodate shear at very low angles ($< 20^\circ$). We mainly vary the strength of the lower crust and the frictional strength of the preexisting mid-crustal shear zone. We find that the strength of the lower crust, and the existence and strength of a preexisting shear zone significantly affect the formation and evolution of core complexes. With increasing lower crustal strength, we recognize varying extensional features with decreasing exhumation rate: these are characterized by bivergent metamorphic massifs, classic Cordilleran metamorphic core complexes, multiple consecutive core complexes (or boudinage structures), and a flexural core complex underlined by a large subsurface low-angle detachment fault with a small convex curvature. Topographic loading and mantle buoyancy forces, together with divergent boundaries, drive a regional lower crustal flow that leads to the exhumation of the lower crust where intensive upper crustal faulting induces strong unloading. The detachment fault is a decoupling zone that accommodates large displacement and accumulates sustained shear strain at very low angle between upper and lower crust. Though the regional stress is largely Andersonian, we find non-Andersonian stress in regions adjacent to the preexisting shear zone and those with high topographic gradient. Our new models provide a view that is generally consistent with geological and geophysical observations on how core complexes form and evolve.

Index terms and key words:

8109 Continental tectonics: extensional (0905)

8120 Dynamics of lithosphere and mantle: general (1213)

8159 Rheology: crust and lithosphere (8031)

8004 Dynamics and mechanics of faulting (8118)

Core complex; Low-angle normal fault; Detachment fault; Crustal flow; Shear zone; US

Cordillera

1. Introduction

Continental core complexes have drawn great interest since their first discovery about 40 years ago. They 1) have a distinctive structural style [e.g., *Platt et al.*, 2014], 2) are generally found in orogens such as the US Cordillera and the Aegean [e.g., *Whitney et al.*, 2013], and 3) have important implications for the mechanical properties of crust, stress state, and post-orogenic extension [e.g., *Platt et al.*, 2014]. They typically share three key characteristics (Figure 1): 1) a metamorphic core that exposes deep crust from a depth of up to 15-20 km [e.g., *Lister and Davis*, 1989]; 2) a low-angle detachment fault as a weak layer that contains mylonite, blastomylonite, and microbreccia shattered by closely spaced fractures [*Davis and Coney*, 1979]; and 3) mid-to-high angle hanging wall normal faults lying above the detachment faults [*Davis and Coney*, 1979]. Finally, a nearly flat Moho is typically observed beneath regions where core complexes are found [*Klemperer et al.*, 1986; *McCarthy et al.*, 1991; *Sodoudi et al.*, 2006]. These features provide end-member observations that allow us to explore the ways by which faults slip, strain localizes and partitions. They also shed light on the evolution of the thermal structure and rheology of the crust.

Several models have been proposed to explain the origin of core complexes and/or detachment faults. Stress rotation induced by basal shear or basal normal forces [e.g., *Yin*, 1989; *Spencer and Chase*, 1989; *Melosh*, 1990] has been analytically shown to allow the formation of low-angle normal faults (LANFs), although debate persists on whether the resulting stress field allows slip on a low-angle detachment fault [*Wills and Buck*, 1997]. The presence of an initial weak zone has been shown not only to rotate the regional stress but also reduce the minimum dip angle for slip on a fault surface [*Faulkner et al.*, 2006; *Collettini et al.*, 2009]. Finally, crustal flow [*Block and Royden*, 1990; *McKenzie et al.*, 2000] and magmatic activity [*Gans and Bohrson*, 1998] have also been shown to play a significant role in the formation of some core complexes. Each of these models contribute to explain certain aspects of the evolution of core complexes, but none simultaneously explains the three main characteristics described above or the presence of a sub-horizontal Moho. For instance, the current most widely accepted rolling-hinge model [*Spencer*, 1984; *Wernicke and Axen*, 1988; *Buck*, 1988; *Lavier et al.*, 1999; *Choi et al.*, 2013] successfully predicts the geometry of many core complexes, however, it gives little information on the mechanisms controlling lower crustal flow, or the exhumation of a metamorphic core and the presence of normal fault earthquakes slipping at very low angle [e.g., *Wernicke*, 1995]. Moreover, it predicts that the detachment fault is of high-angle origin, a point that contradicts the fact that some detachment faults exhibit characteristics that are only explainable when invoking slip on a normal fault at very low angle [e.g., *Collettini*, 2011 and references therein]. More specifically, in places like the Whipple Mountains [e.g., *Lister and Davis*, 1989] or the Mikonos detachment system [*Lecomte et al.*, 2010], the intersection angles between syn-tectonic depositional strata and the detachment fault indicate slip at low angle. In other places like the

Black Mountains, hanging wall high-angle normal faults soling into gently dipping detachment faults that cut subhorizontal sediment layers [Hayman *et al.*, 2003] demonstrates slip at low angle.

Multiple models have been proposed to explain slip at low-angle normal faults (LANFs) [e.g., Yin, 1989; Spencer and Chase, 1989; Axen, 1992; Melosh, 1990; Forsyth, 1992; Westaway, 1999] and more recently Hayman *et al.* [2003]. One key aspect of both these models and observations of active LANFs is the presence of weak frictional phyllosilicates generated by fluid rock interaction [e.g., Colletini *et al.*, 2011] or potentially near lithostatic fluid pressures [Hayman *et al.*, 2003]. In addition, geological observations sometime associate the formation of LANFs to pre-existing ductile shear zones [e.g., Davis, 1983; Colletini *et al.*, 2009] or sometimes reactivated thrusts as in the case in the North-Cycladic detachment system [Jolivet *et al.*, 2010]. While not ubiquitous, these observations suggest that in many cases the presence of a weak ductile shear zone near or below the brittle ductile transition is a necessary component of metamorphic core complexes.

The clustered distribution and striking similarities of continental core complexes in orogenic belts such as US Cordillera and the Aegean Sea motivates us to investigate their origin in a context akin to post-orogenic extension. The US Cordillera and the Aegean are young Mesozoic/Cenozoic orogenic belts that have undergone intense Cenozoic post-orogenic extension mainly due to slab retreat and collapse [e.g., Wernicke, 1992; Jolivet and Brun, 2010; Jolivet *et al.*, 2013]. Both the Cordilleran and Aegean Sea can be categorized as Andean-type orogeny [Dewey and Bird, 1970], consisting of forelands, hinterlands, and a plateau. Orogenic hinterlands have wedge-shaped geometry with topography and Moho relief. The lithosphere is hot and their formation is typically accompanied by magmatism that is consistent with a weak rheological structure [Whitney *et al.*, 2004]. The geometry, rheology and boundary conditions of orogenic hinterlands might significantly affect subsequent extension. It is probable that the extension of an orogenic hinterland due to far-field boundary relaxations naturally leads to formation of core complexes [e.g., Rey *et al.*, 2009].

Here we hypothesize that the presence of a rheologically weak crustal wedge [e.g., Xiao *et al.*, 1991; Platt *et al.*, 2013] including a preexisting crustal-scale ductile shear zones provide the necessary elements to consistently predict the formation of some types of metamorphic core complexes along a preexisting detachment system with mid- to high-angle hanging wall normal faults rooted in the detachment fault. The dip angle of the detachment system can be small, i.e., $< 15^\circ$ and it varies with depth and across different models. We assume that the décollement is low-angle, weak, foliated, and was formed during the previous phase of Andean-type orogenesis.

To test our hypothesis, we first provide a numerical analysis of lower crustal flow driven by topographic loading and mantle buoyancy. We devise numerical simulations of the extension

of a crustal wedge that resembles an orogenic hinterland in a short-lived orogenic belt [see *Wu et al.*, 2015]. We do not assume any orientation of our crustal wedge in an orogen. We simply take a cross section from a region with high topography (termed highland) to a region with low topography (termed lowland) in an orogenic hinterland. Here, we focus on the mechanics of detachment faulting, the associated upper-crustal normal faults and the dynamic evolution of core complexes. We present a series of numerical models for the evolution of core complexes, compare them with the major core complexes in the US Cordillera and the Aegean, and discuss their characteristics. We use physical and rheological properties that are measured in laboratory experiments or deduced from field observations [e.g., *Collettini et al.*, 2009; *Behr and Platt*, 2011] as well as the traditional Mohr-Coulomb elastoplastic model to simulate the formation and slip of normal faults [*Lavier et al.*, 2000]. We also use classical creep law [e.g., *Burgmann and Dresen*, 2008] to simulate the flow of ductile material in the lithosphere. A detailed discussion and comparison of the model setup to previous models of continental extension in orogens is provided in *Wu et al.* [2015].

2. Methods

2.1 Analysis of lower crustal flow

Lower crustal flow as channel flow has been analyzed in both compressional and extensional settings [*Block and Royden*, 1990; *Bird*, 1991; *Wdowinski and Axen*, 1992; *Nagel and Buck*, 2006], however, in the case of a crustal wedge including both topography and Moho relief and two viscous layers, it has not been analyzed in detail. Here, before proceeding with the presentation of our numerical models, we provide an updated solution of lower crustal flow in a crustal wedge [*Wu et al.*, 2015]. This forms a theoretical basis for our model setup and a better framework for the presentation of our model results. For the following analytical solution, we assume that 1) both the regional topography and Moho relief are subhorizontal ($< \sim 10^\circ$), 2) no slip occurs along the boundaries, 3) the shear zone is initially at a depth of > 15 km, i.e., it is viscous, 4) constant viscosity for both the shear zone and lower crust, 5) a topographic gradient of $\partial h_t / \partial x$ and a Moho relief gradient of $\partial h_m / \partial x$. The flow of the lower crust and shear zone may be idealized as the channel flow of two immiscible incompressible viscous fluids under horizontal pressure between two fixed horizontal rigid plates (Figure 2a).

The overall horizontal pressure gradient in the lower crust and shear zone of a crustal wedge induced by topographic loading approximately equals to that induced by mantle buoyancy, and can be expressed [*Wu et al.*, 2015],

$$\nabla P_x \approx \rho_c g \frac{\partial h_t}{\partial x} \approx (\rho_m - \rho_c) g \frac{\partial h_m}{\partial x} \quad (1)$$

, where ρ_c is the density of crust and ρ_m the density of mantle. Following procedures in *Bird et al.* [2007], the velocity in the shear zone is,

$$v_s = \frac{\nabla P_x}{2\mu_s} \left[h_s^2 + \frac{\mu_s h_{lc}^2 - \mu_{lc} h_s^2}{\mu_{lc} h_s + \mu_s h_{lc}} (h_s - z) - z^2 \right], 0 \leq z \leq h_s \quad (2)$$

and in the lower crust it is,

$$v_{lc} = \frac{\nabla P_x}{2\mu_{lc}} \left[\frac{h_s^2 \mu_{lc}}{\mu_s} + \frac{\mu_s h_{lc}^2 - \mu_{lc} h_s^2}{\mu_{lc} h_s + \mu_s h_{lc}} \left(\frac{h_s \mu_{lc}}{\mu_s} - z \right) - z^2 \right], -h_{lc} \leq z \leq 0 \quad (3)$$

, where h_s and h_{lc} are the thicknesses of the shear zone and the lower crust, respectively, and μ_s and μ_{lc} are the viscosity of the shear zone and the lower crust, respectively. Without the shear zone, the lower crust forms a channel of single viscous layer, with a velocity of,

$$v_{lc}' = \frac{\nabla P_x}{4\mu_{lc}} (h_{lc} - z^2), -h_{lc}/2 \leq z \leq h_{lc}/2 \quad (4)$$

Therefore the average velocity in the lower crust with the shear zone is,

$$\overline{v_{lc}} = \frac{\nabla P_x}{12\mu_{lc}\mu_s} \frac{\mu_s^2 h_{lc}^3 + 4\mu_s \mu_{lc} h_s h_{lc}^2 + 3\mu_s \mu_{lc} h_s^2 h_{lc}}{\mu_s h_{lc} + \mu_{lc} h_s} \quad (5)$$

, and that without the shear zone,

$$\overline{v_{lc}'} = \frac{\nabla P_x h_{lc}^2}{12\mu_{lc}} \quad (6)$$

The schematic velocity profile (Figure 2a) shows that a weak shear zone significantly enhances the flow of the upper part of the lower crust, and that the strain in the weak shear zone is significant, and predominately simple shear. We assume a value of 2 km for the shear zone thickness and 16 km for the lower crust, and using equations 1, 5 and 6 we plot the average velocity of the lower crust with respect to viscosity (varying from 10^{19} Pa s to 10^{23} Pa s) using a fixed viscosity of 10^{19} Pa s for the shear zone. The results show that the lower crustal velocity is very sensitive to the viscosity of the lower crust, and that the existence of a weak shear zone greatly enhances the lower crustal flow (Figure 2b). For a reasonable tectonic velocity, defined as divergent plate boundary velocity, of ~ 2 cm yr^{-1} [e.g., *DeMets et al.*, 2010] it is clear that lower crustal flow is driven by the pressure gradient induced by topographic loading and mantle

buoyancy. It can be greater or comparable to the tectonic velocity if the lower crust is weak enough, i.e., generally weaker than 3×10^{20} Pa s (Figure 2b). If the velocity of the lower crust is greater than the tectonic velocity, then the lower crustal flow may decouple from the deformation occurring in the upper crust and mafic lower crust (and mantle), and the regional isostasy will control flow and deformation in the crustal wedge [Wu *et al.*, 2015].

2.2 Model setup

We use FLAC [e.g., Poliakov *et al.*, 1993; Lavier *et al.*, 1999; Lavier and Buck, 2002; Lavier and Manatschal, 2006; Tan *et al.*, 2012] for numerical simulations. The method is described in Wu *et al.*, 2015. In short, we use a crustal wedge (Figure 3) to approximate a representative orogenic hinterland and vary the strength of the lower crust across different models [Wu *et al.*, 2015]. We include a preexisting ductile shear zone between the upper and lower crust at the beginning of extension, assuming it was formed during the preceding orogeny. The regionally continuous shear zone is a model simplification. The detailed model setup is shown in Figure 3a, and its strength profile and physical parameters are shown in Figure 3b and Table 1, respectively.

We set a constant temperature of 10 °C and 800 °C for the surface and bottom, respectively and zero heat-flow boundary conditions on both sides of the model. In the model, we do not consider the complex rheological evolution of the preexisting shear zone. Instead we imposed a given frictional strength when it is transitioning in the brittle field. When it is ductile, it is always weak and has a specific low cut-off viscosity ($\sim 10^{19}$ Pa s). Processes such as magmatic activity and delamination are not simulated. No weak seed is included in the numerical models. We tested various parameters, such as topographic gradient, Moho relief, and bottom boundary temperature [Wu *et al.*, 2015] before focusing on the frictional strength of the shear zone and the viscosity of the lower crust, which we found are the most important controlling factors. The viscosity obeys dislocation creep law, is thermally controlled and varies with depth. To vary the viscosity of the lower crust for different models, we systematically changed the activation energy term in the dislocation creep law for plagioclase [Ranalli, 1995] so that it is in the regime of wet quartz with partial melting (very weak), wet quartz (intermediate strength), or plagioclase (strong) (Figure 3b). We used Mohr-Coulomb yield criteria for upper crust and most commonly accepted dislocation creep laws for the mafic lower crust and mantle [Ranalli and Murphy, 1987; Kirby and Kronenberg, 1987]. We varied the frictional strength of the shear zone by changing the internal frictional angle of the Mohr-Coulomb failure envelope in different models. Spreading that would be caused by gravitational rollback is simulated by divergent boundary conditions, gravity-driven relaxation of the topography, and changes in crustal thickness. It is not clear how each contributes to the extension, however, we know that the divergent extension rate typically ranges from 1 - 2 cm yr⁻¹ [e.g., DeMets *et al.*, 2010], and that a regional topographic gradient of < 0.05 is reasonable. Setting an extension rate as well as a variation in crustal thickness simulates

the contributions of both extensional divergence and crustal buoyancy in the numerical simulations. An overall horizontal boundary velocity of 1.25 cm yr^{-1} (i.e., $\sim 0.625 \text{ cm yr}^{-1}$ for each side) is used for all the models presented here. Winkler foundation is used for bases of our models, and material flux is allowed on both sides.

3. Results

In a previous study [Wu *et al.*, 2015], we found that a pre-existing sub-horizontal shear zone (detachment fault) or possibly one developed during extension is critical to the development of continental core complexes that are consistent with observations of their three key elements and the subhorizontal Moho. If the detachment structure between the upper and lower crust is absent, structures more characteristic of rifted margins such as half-grabens and tilted blocks form. Extension in a crustal wedge leads to the formation of continental margins unless the lower crust is very weak [Wu *et al.*, 2015]. We propose here that core complexes are the product of the feedback between extension along the detachment system and fast lateral lower crustal flow during post-orogenic extension in a crustal wedge. We present a suite of models and an analysis of their characteristics in an attempt to substantiate our hypothesis. We use an upper and lower divisions for the crust, but we should point out that the exhumed deep crust in our models presented are principally exhumed from mid-crustal level, which is sometimes termed middle crust [Wernicke, 1992].

3.1 Detachment systems

Distinct detachment systems arise in our numerical models. Here, we take the liberty to propose three types of model detachment system, defined as a shear zone across which significant displacement, strain accumulation and partitioning occur (Figure 4). Type-I is a convex upward detachment fault that behaves as a brittle shear zone at shallow depth and transitions to a ductile shear zone at greater depth. In Figure 4a the brittle domain of the detachment or normal faults is indicated by a continuous red line and the ductile domain by a fading transitional red dashed line. All the detachments initiate in the preexisting shear zone and underlie a hanging wall populated with normal faults rooted in the detachment. Type-I detachment requires a weak to intermediate-strength lower crust so that lower crustal flow is intense enough to exhume the detachment to the surface (Figure 4a). Unless the lower crust is very weak (e.g, Fig. 5A in Wu *et al.* [2015]), a shear zone is required but not necessarily weak when it is brittle. For example, the shear zone in Figure 4a is relatively strong with a friction coefficient of 0.47 in the brittle domain ($< 13 \text{ km}$ depth) and transitions to a viscosity of $\sim 10^{19} \text{ Pa s}$ in the ductile domain ($> 13 \text{ km}$ depth).

Type-II is defined as a listric detachment fault constituted of a brittle listric normal fault in the crust which soles out into the preexisting shear zone at depth. The brittle part of the shear zone transitions to a basal ductile shear zone at depth (Figure 4b). The listric normal fault is of

high-angle origin and is progressively rotated to mid-to-low angle as extension progresses. The listric detachment fault exerts a strong control over the structure of the extended region. Type-II typically forms when the lower crust fails to exhume to the surface. They require intermediate-strength lower crust and the brittle strength of the preexisting shear zones may vary.

Type-III is a subsurface one that is mainly a sub-horizontal brittle shear zone which transitions to a ductile shear zone at greater depth (Figure 4c.). As shown in Figure 4c, a LANF with a dip angle of 20° is also predicted at the range front, changing to 40° at deeper depth.

From type-I to type-III, upper and lower crustal extension becomes more coupled, and the intensity of the exhumation and lateral flow of the lower crust decreases. This is consistent with our analytic calculations (Equations 2 and 3). The frictional coefficients (0.09 - 0.46) of the shear zone used in our models are reasonable and consistent with those of common minerals in shear zones [Table 3 in *Collettini, 2011*]. In all of these cases, the models show that the orientation of the principal stresses in the detachment fault is rotated away from the vertical axis (Figure 4, σ_1 orientation). The stress is strongly rotated in the regions adjacent to the brittle and ductile detachment and close to the surface where a steep topographic gradient exists (Figure 5). The rotation of the principal axis away from the vertical establishes a local non-Andersonian state of stress that facilitates slip along the generally low-angle detachment faults. Figure 5 shows that the non-Andersonian stress persists even in the zone where the detachment fault behaves in a brittle manner following Mohr-Coulomb yield criterion (inset with plastic strain $\sim$ brittle deformation). The non-Andersonian stress extends in the ductile part of the detachment. The σ_1 , however, is still largely vertical in the brittle upper crust except in the regions mentioned above (Figure 5).

3.2 Hanging wall normal faults

The normal faults forming in the hanging wall of the detachment faults have different styles defined by the overall flow in the lower crust and the corresponding type of detachment fault. Overlying type-I detachment faults, the normal faults form a chaotic mosaic, in which synthetic and antithetic normal faults overprint each other (Figure 4a). This is partially due to the extreme thinning of the upper crust and to continuous slip along the convex basal detachment. The normal faults develop in a process similar to that predicted by the rolling-hinge model [*Wernicke, 1992; Lavier et al., 1999; Choi et al., 2013*]; however, in our models a given generation of faults may be active simultaneously and alternate its activity with another group of faults. Normal faults may get locked as they rotate to dip angles unfavorable to continuous extension and as the detachment fault becomes sub-horizontal [*Choi and Buck, 2012; Choi et al., 2013*]. Following locking a new generation of normal faults forms closer to the range front (Figure 4).

The normal faults in the hanging wall of type-II detachments are connected to the master listric fault and behave mainly as antithetic faults (Figure 4b). The hanging-wall of type III detachments is populated by domino style sets of synthetic normal faults (Figure 4c). The dip angle of the faults depends on the degree of rotation, but they generally slip at mid-to-high angle (30° - 50°).

The slip rate along the detachment exerts control over the style of hanging wall normal faults. We found that the normal faults in the hanging wall of the detachments dip in the same direction when the slip rate of the detachment fault is high (e.g., Figure 5). When the slip rate of the detachment fault is low, conjugate normal faults are formed. In other words, the strain rate in the shear zone must be large enough to cause significant enough basal shear stress to favor domino faults dipping in the same direction. For example, hanging wall normal faults consecutively form and dip in the same direction when the slip rate of the detachment fault is high during the fast exhumation of the lower crust (Figures 4c and 5).

3.3. Core complex evolution

We present four models that generate different type of core complexes (Figure 6) defined as a domal structures including multiple detachment faults with their corresponding hanging wall fault systems (Figures 7-10). In the following, we distinguish the highland and lowland (Figure 3a) as the areas of high and low topography in the initial structure of the crustal wedge, respectively.

3.3.1 Bivergent core complexes (metamorphic massifs or gneiss domes)

Bivergent core complexes develop with a very weak lower crust, featuring a type-I detachment system and mosaic synthetic and antithetic hanging wall normal faults (Figures 6a and 7). It has the features of a bivergent set of detachment faults similar to the central Menderes core complex [Gessner *et al.*, 2001]. The detachment fault formed on the highland side of the crustal wedge accumulates extension continuously, while the detachment fault on the lowland side passively adjusts to the overall extension (Figure 7). Mid-to-high-angle normal faults may develop in the massif when the lower crust is exhumed and has cooled to a temperature low enough for brittle deformation to occur (Figure 7c). Once the upper crust breaks up, the lower crust exhumed and accommodates most of the extension with stable highland and lowland upper-crustal blocks slipping along the detachment faults on both sides of a massif and defining a core complex. The exhumation rate of lower crust is very high. In the model shown in Figure 7 the shear zone and lower crust are exhumed from approximately 16 km depth to the surface over 1 Myr of extension.

The presence of a very weak lower crust (wet quartz regime with partial melting) is crucial for the development of the bivergent core complex since it allows for rapid lateral flow

and exhumation of the lower crust. In that case the topographic gradient can support a very fast flow in the lower crust driven by the overall topographic gradient (Figure 7). The presence of a preexisting sub-horizontal detachment fault significantly facilitates the exhumation of lower crust through continuous extension along the detachment system as predicted by the analytic model (Figure 2). Most of the extension in the bivergent core complex is accommodated by the exhumation of the lower crust. The Moho depth varies between 20 km below the lowland and 30 km under the highland and has therefore been significantly reduced after 8 Myrs of evolution (Figure 7c).

3.3.2. Metamorphic core complex

This model generates core complexes with characteristics observed in the Basin and Range and the Aegean Sea [e.g., *Coney et al.*, 1980; *Whitney et al.*, 2013] (Figure 1a, c, e and f). The core complexes feature type-I detachment faults, sequentially developed synthetic and antithetic normal faults, and an exhumed asymmetric lower crustal dome (Figure 6b). On the lowland side of the model, a type-III detachment fault has formed above a 20 km wide dome of lower crust that has failed to exhume to the surface (Figure 6b, zone 1 and Figure 8). On the highland side, another small lower crustal dome has formed that could potentially develop into another core complex if more extension were to accumulate in the model (Figure 8c). As opposed to the bivergent massif, both upper-crustal thinning and lower crustal exhumation play an equivalent role in accommodating extension (Figure 6b and 8). Furthermore, detachment faults also accommodate significant extension. The size of the core complexes ranges from 50 km to more than 100 km, and the exhumation rate is intermediate. In Figure 8, the shear zone is exhumed from 15-20 km deep to near the surface and in about 2-3 Myrs.

The model has an intermediate-strength lower crust (wet quartz regime) and a weak shear zone (with low frictional strength < 0.2). In its early stage, a type-II detachment fault develops which is critical for the later evolution by forming an active feedback between exhumation of lower crust and extension along the detachment system. Once formed, it dominates the extension, however, it might or might not be able to develop into a fully evolved metamorphic core complex.

3.3.3. Boudinage structures

This model resembles the metamorphic core complex in that it is characterized by similar detachment faults and hanging wall normal faults (Figures 6b-c, 9). The boudinage structures differ from the metamorphic core complexes in their inability to exhume the lower crust to the surface. Core complexes are typically less than 50 km and have typical boudinage structures. The boudins develop a series of lower crustal domes and upper crustal swells that start in the lowland and propagate to the highland as extension continues (Figures 6c, 9). The boudinage mode is

associated with a lower crust (wet quartz regime) and a shear zone of intermediate strength (with intermediate frictional coefficient $\sim 0.2 - 0.46$).

3.3.4. Flexural core complex

This type of model generates a type-III detachment fault. It also features sequentially developed hanging-wall normal faults which are synthetic, mid-to-high-angle and in domino style that sole into the detachment fault. Underlying the type-III detachment fault is a dome of lower crust with low uplift curvature (Figures 6d and 10). If the strain rate in the shear zone is high enough to favor strongly non-Andersonian stress, a series of synthetic domino style normal faults form. Otherwise a system of conjugate synthetic and antithetic hanging wall normal faults form. Both the upper crust and lower crust accommodate extension to equivalent degree. Exhumation of the lower crust is very slow, and it requires ~ 10 s Myrs for flexural core complex to form. In the model shown in Figure 10 the shear zone is exhumed from a depth of 18 km to 6 km in approximately 6 Myrs.

Flexural core complexes have a strong lower crust (dry quartz or plagioclase creep regime) and a weak-intermediate-strength preexisting shear zone (frictional coefficient $< \sim 0.46$). However, a stronger shear zone inhibits the formation of the lower crustal dome with a smaller curvature. The extension in the upper crust through domino style faulting dominates the evolution of the deformation in the upper crust, and the exhumation of lower crust is subdued due to its high viscosity. Even with a detachment fault, a decoupling between the upper and lower crust does not occur, as predicted by the analytic model (Figure 2).

3.4. Lower crustal flow and intracrustal isostasy

Crustal and mantle flow during the formation of the different core complexes generate two kinds of feedback: 1) between faulting and stretching in the upper crust and lower crustal flow along a detachment fault, 2) between regional crustal thinning and lithospheric mantle necking. The first one controls the stretching style of the crust while the second governs how the mantle isostatically compensates for crustal thinning.

In all the numerical models presented (Figures 7-10) the lower crust flows from highland to lowland, and necking occurs in areas where faulting in the upper crust is active. When the lower crust is weak and/or the preexisting detachment is weak, the flow decouples the upper crustal deformation from the deformation in the lower lithosphere along a detachment fault, a result consistent with the theoretical predictions (Figure 2 and Wu et al, 2015). Necking of the lithospheric mantle is small as long as the lower crustal flow is not activated. When lower crustal flow occurs from highland to lowland, the mantle flows asymmetrically towards the highland to compensate for the depletion of lower crust. As a result, lower crustal flow largely compensates extension in the upper crust, while mantle upwelling beneath the highland compensates for the

regional flow of lower crust towards the lowland. A very long-wavelength isostatic compensation coexists with intracrustal isostasy. The intracrustal compensation wavelength increases as decoupling between upper crustal deformation and lower crustal flow decreases. This occurs as the lower crustal strength and detachment strength increase from bivergent, to metamorphic, to flexural core complexes. The long wavelength regional isostasy leads to the formation of a smooth subhorizontal Moho. The two coexisting wavelengths of local crustal and regional lithospheric isostasy are consistent with theoretical predictions [Zuber *et al.*, 1986]. As a consequence of the flow, the mafic lower crust is sheared away from the highland.

3.5. Thermal history

In Figure 11 we plot the thermal history of a material particle attached to the detachment fault for the bivergent core complex, metamorphic core complex, boudinage structures and the flexural core complex (black star or dot in Figures 7-10). The temperature of the detachment faults decreases from 300-400°C to around 20-50°C. This is consistent with samples from natural core complexes [e.g., Kounov *et al.*, 2015]. The exhumation rate is fastest for the bivergent case, cooling 250°C cooling over 1 myr (Figure 11i). Metamorphic and boudinage cases also cool rapidly, decreasing 250-300 °C over 1.5 to 2 Myrs (Figure 11ii-iv). Flexural uplift cases cool much slower, decreasing ~125°C over 6 Myrs (Figure 11v). The trend is generally consistent with cooling rate of natural core complexes [e.g., Fig. 6 in Whitney *et al.*, 2013], however our models may not be able to capture the magnitude of the temperature history as we are not accounting for magmatic processes (partial melting and fluids migration) or the complex thermal history associated with the orogenic processes preceding the extension. However, the models do clearly show that as the strength of lower crust and of the detachment increases from the bivergent to the flexural core complex, the time window for the formation of core complexes is greatly lengthened, and the cooling history is significantly longer (Figure 11).

4. Discussions

4.1. Preexisting shear zone at mid-crustal depth

Mylonitic shear zones are observed in almost every known continental core complex [e.g., Bartley and Wernicke, 1984; Brown and Solar, 1998], although debate persists regarding their origin. They typically display a complex deformational history, for instance, the microstructures in the mylonite shear zone of the Buckskin-Rawhide detachment fault in west-central Arizona indicate two opposing senses of slip [Singleton, 2013], and display several weakening mechanisms, such as fluid-assisted granular flow and grain size reduction [e.g., Oliot *et al.*, 2013; Smith *et al.*, 2011; Platt and Behr, 2011]. Though our model may be quite simple, it is supported by the discrepancy between the mechanical responses of the brittle upper crust and ductile lower crust which inevitably generates a zone of strong strain localization, i.e., a mid-

crustal shear zone [e.g., *Gueydan et al.*, 2004]. In addition, recent reports show intriguing examples of mid-crustal shear zones characterized by weak anastomosed phyllonitic shear bands rich in phengite and quartz which accommodate significant amount of shortening in the Alps [*Bellanger et al.*, 2014]. A core complex can be modeled without an explicit shear zone when the lower crust is very weak [e.g., *Buck et al.*, 2005; *Wu et al.*, 2015]. Such a case might be suitable for oceanic core complexes but seems not for continental core complexes.

In our models, the shear zone is assumed crustal-scale, mid-crustal, and viscously weak when extension starts, and is formed during previous orogenesis. It is a long-lived decoupling zone between upper and lower crust [e.g., *Gueydan et al.*, 2004]. Preexisting shear zones in the crust also include 1) a preexisting thrust fault [e.g., *Banks and Warburton*, 1991; *Jolivet et al.*, 2010] and 2) a recently formed mid-crustal ductile shear zone formed by ductile weakening processes [e.g., *Oliot et al.*, 2013; *Smith et al.*, 2011; *Platt and Behr*, 2011] or semibrittle processes [e.g., *Lavier and Manatschal*, 2006]. A natural shear zone could be one or any combination of these shear zones at different depths and times. Due to their weak nature, such shear zones are able to accumulate large amount of shear strain, generating attenuated structures [e.g., *Brown and Solar*, 1998] and may release energy through aseismic creep events instead of catastrophic earthquakes. They may potentially be the origin of very LANE earthquakes [e.g., *Wernicke*, 1995], although most normal fault earthquakes slip at a high dip angle [e.g., *Jackson and White*, 1989].

Our theoretical analysis and numerical models reinforce the idea that the shear zone is both geologically and geophysically significant, especially during postorogenic extension. Hence, we believe that a crustal scale shear zone of variable strength always exists at mid-crustal depth, although it may not be spatially continuous as assumed in our models. In addition, the strong shear strain suggests that a shear zone along the Moho is also highly probable (Figures 7-10). However, the origin of shear zones in the lithosphere and their tectonic significance need further investigation.

4.2. Detachment systems and low-angle slip

The detachment systems described in our models are zones of localized deformation with high strain rate that can accumulate very large amount of strain (10s kilometers of extension). Both upper crustal brittle and lower crustal ductile deformation occur in these zones of high strain accumulation, although the amount of extension accumulated in a brittle or ductile manner varies as a function of the strength contrast and decoupling between the upper and lower crust (Figure 2, *Wu et al.* 2015).

Our models also show that a brittle detachment can accumulate strain at very low angle. In an elastoplastic layer yielding according to the Mohr-Coulomb criterion, the slip on a normal

fault at dip angle lower than 30° is inhibited even for low coefficients of static friction [Anderson, 1951; Wills and Buck, 1998]. Rather than slip on the preexisting LANF, a new high-angle normal fault forms in the layer to accumulate strain [e.g., Wills and Buck, 1998; Lavier *et al.*, 2000]. However in the cases presented here two factors are simultaneously at play: 1) the initially weak preexisting shear zone acts as a heterogeneity that can slip at low angle as long as the state of stress in the hanging wall has not reached yield (Figure 12), and 2) the shear flow enhanced by the regional topographic gradient (Figure 2) at the base of the detachment rotates the principal stresses away from the vertical axis [Melosh, 1990], inducing a non-Andersonian state of stress and facilitating normal slip at low angle.

An analysis of the Mohr-Coulomb circle (Figure 12) [Sibson, 1977; Axen, 1992; Axen, 1999; Collettini *et al.*, 2011] illustrates how a weak preexisting shear zone allows for the initiation and slip along the detachment faults originated from the shear zone even at unusually low angles while resisting the formation of a new high-angle normal fault in the hanging-wall. The presence of a preexisting weak brittle and ductile shear zone in the crustal wedge acts as a low-angle heterogeneity at the base of the upper crust. The detachment and upper crustal yield stress envelope are represented with a Mohr-Coulomb circle corresponding to a hypothetical state of stress in the crustal wedge (Figure 12). The detachment has already yielded (black dot on the circle) while the upper crust has not yet yielded. Therefore, the formation of new faults in the upper crust is delayed by the weak preexisting shear zone as well as the rotation of the stresses to a non-Andersonian state allowing further slip at low angle.

These effects are modulated by the rotation and exhumation of the lower crust when deformation is decoupled (bivergent and metamorphic core complexes). For low coupling and weak detachment faults, slip along the detachment is initially facilitated by the state of stress and later by the rotation of the detachment to a higher dip angle during lower crustal exhumation. When coupling increases, slip over the detachment fault is more difficult since the coefficient of static friction is larger (boudinage structures). Since the frictional strength of the brittle part of the detachment fault is higher and resists continuous slip, it eventually locks up when the dip of the detachment becomes too low. As a result, multiple zones of active faulting and exhumation form, sequentially generating the pinch and swell structures characteristic of crustal scale boudinage. Finally, when the strength of the upper crust is similar to that of the lower crust, the brittle detachment accommodates shear at low angle before high-angle normal faults form. When high-angle normal faults are rotated to a dip angle unfavorable to slip, the detachment slips again. Detachment faults and hanging wall normal faults work intermittently to accommodate extension. Sometimes slip is accommodated along the detachment at low angle, at other time it is accommodated by a set of hanging-wall normal faults slipping at high angle. Seismological evidences of seismicity on low-angle normal faults are documented in the northern Apennines of Italy [Chiaraluce *et al.*, 2007].

The origin of detachment faults in our models is a sub-horizontal crustal-scale mid-crustal shear zone that is warped and exhumed to variable degrees in the brittle and/or ductile regimes (Figures 4, 7-10). The characteristic LANF at the range front (75 – 90 km in Figure 5b) forms because of the large topographic gradient forcing a high degree of rotation of the maximum principal stress (σ_1) from the vertical direction (non-Andersonian). Primary detachment faults as well as faults which rotate from high to low angles are generated in our numerical models. The models therefore suggest that both high-angle origin and primary detachment faults are probable. The numerical models also suggest that the visible parts of the detachment faults with extremely low-angle dip overprint shear zones that are originally ductile. The results are consistent with geological observations in many core complexes. Finally, it is probable that many of the exhumed shear zones were aseismic and creeping with very small earthquakes during its evolution.

4.3. Lower crustal flow and intracrustal isostasy

Lower crustal flow effectively transports hot lower crust from the stable highland towards the extended lowland. The flow is driven by the topographic loading and mantle buoyancy, thereby reducing both topography and Moho relief in the highland [Wu *et al.*, 2015]. The decreasing thickness of lower crust from highland to lowland also intensifies the lower crustal flow due to restrictions imposed by mass conservation. As a result, intracrustal isostasy predominates when the lower crust is weak enough. For a specific crustal wedge, the driving forces are determined once the geometry and boundary conditions are set [Wu *et al.*, 2015]. The biggest factor affecting crustal flow is then the viscosity of the ductile lower crust [Wu *et al.*, 2015]. A weak lower crust facilitates flow, while a strong one resists it (Figure 2). Strong decoupling and depth-dependent extension is observed with a weak-to-intermediate-strength lower crust (Figures 7-9). Strong crustal flow enhances intracrustal isostasy, and largely inhibits mantle flow for compensation. The mantle passively adapts to the overall deformation pattern of the crust, and leads to upwelling below the highland domain of the crustal wedge. The relative strength of crustal flow when compared with extension driven by divergent boundary conditions affects both the style of extension and compensation, as is evident in Figures 7-10 (see Wu *et al.*, 2015 for details). Clearly, the numerical models cannot sustain such a decoupled flow for long since the flow itself reduces both topography and Moho relief. If a large plateau seats next to the highland, it is likely that the decoupled flow can last as long as the plateau subsists. Such a mechanism for decoupled flow may be tested in the future.

4.4. Geological and geophysical observations

The models are not meant to be an accurate representation of the formation of core complexes during the collapse of mountain belts. Particularly, the initiation of the deformation in the topographically low areas of the models is likely due to the overly simplistic assumptions (two dimensional wedge) meant to represent the hinterland of a continental collision. Moreover

the fact that the core of modeled MCCs is topographically lower than the hanging wall normal faults is not consistent with the observations and shows that some dynamic processes generating anomalously high topography are omitted.

However, each core complex or detachment fault in our models shares characteristics similar to natural examples, particularly in the US Cordillera and the Aegean Sea (See Figures 1, 7-10 and references therein). The bivergent core complex model resembles geological reconstruction of southwest Rhodope in the Aegean [Gautier *et al.*, 1999] and the Mendere Massif in Turkey [Hetzel *et al.*, 1995] where extensional shearing partly localizes in a décollement and the exhumation rate is high. The strong and fast intrusion of middle and lower crust suggest they are weak, which is consistent with our numerical model. Our model, however, further suggests that the two detachment systems take different roles: one is more “active”, and the other passively adapts to the overall extension. The metamorphic core complex model is consistent to the Whipple Mountains [Davis, 1980], Harcuvar Mountains [Rehrig and Reynolds, 1980], and Snake Range core complex [e.g., Miller *et al.*, 1983; Bartley and Wernicke, 1984; Wernicke, 1992] in the Basin and Range. The modeled boudinage structures resemble the Death Valley turtlebacks [Miller and Pavlis, 2005] in the Basin and Range. Finally, the modeled flexural core complex resembles those reconstructed for the Elba-Adriatic coast [Collettini *et al.*, 2006]. The Crete core complex in the Aegean Sea [Gautier *et al.*, 1999] shares characteristics of bivergent and metamorphic core complexes. Temperature histories and exhumation rates extracted from the models are consistent with most natural core complexes [e.g., Table 2 in Ketchum, 1996; Gautier *et al.*, 1993; Kounov *et al.*, 2015]. The numerical models could be used to interpret a broad range of core complexes known to date.

The style and migration of extension, the bulk thinning of the crust, and the associated intra-crustal compensation predicted by our models could be used to interpret the regional evolution of the Basin and Range [Dickinson, 2002]. The earlier stage generates many core complexes when the lower crust is hot and fluids/magma activity is strong with predominantly intermediate to weakly silicic volcanism [Gans *et al.*, 1989]. The later stage, where mid-to-high-angle normal faults predominate, occurs when the lower crust has cooled and the magmatic activity and fluid migration have been reduced. The strain partitioning in the crust [e.g., Wernicke, 1992] seems to be consistent with our model predictions. Our models are also consistent with the observations of sub-horizontal Moho in the Basin and Range and Western Europe where intensive extensions occurred [Klemperer *et al.*, 1986; McCarthy *et al.*, 1991; Sobouti *et al.*, 2006].

The detachment faults are by far the most controversial feature in the interpretation of core complexes. Geological or seismological evidence suggests detachment faults dip at angles between 10° - 30° [Collettini, 2011]. The models predict they can slip at dip lower than 10° and

that primary detachment faults have their probable origin in Cretaceous orogenic belts for the Basin and Range province. The Black Mountains and the Whipple Mountains detachments where hanging wall grabens are made up of and/or filled with tertiary sediments and volcanic rocks [Hayman *et al.*, 2003; Miller and Pavlis, 2005; Davis, 1980] suggest that these visible detachments have a more recent origin than those exhumed from the lower crust. However, type-I and type-II detachment systems, where large amounts of principal stress rotation occur near the surface, would allow slip at low angle at the contacts between the sediments and exhumed lower crust. This mechanism has the effect of rejuvenating the detachment surface and is consistent with the observations.

When identifying the portion of the modeled LANFs that accumulate strain at low angle, we find that areas with a non-Andersonian state of stress and where high-angle and listric normal faults root in the preexisting ductile shear zone characterize them (Figure 5 - 10). Zones of low-angle slip also correspond to range front faults that share many characteristics with the basal detachment of observed at landslides (Figure 5). One important feature of the model is that the regional state of stress stays Andersonian while it departs from that state locally where either fast shear flow in the middle crust or a high topographic gradient is present. Moreover our models predict that brittle deformation in the detachment faults overprints ductile deformation in the preexisting shear zone. While they do not explain the potential existence of initially brittle LANFs, the results are consistent with many of the observations at metamorphic core complexes [e.g., Collettini, 2011].

5. Concluding remarks

Our models predict that core complexes are natural products of the decoupling between the extension in the upper crust and the lateral flow and exhumation of lower crust along a detachment. We presented three types of detachment systems and four types of core complexes based on numerical models. A crustal wedge with a very weak lower crust (likely hot and intruded by magmatic products) leads to the development of a bivergent core complex (or metamorphic massif). An intermediate-strength lower crust (wet quartz regime) leads to the development of classic Cordilleran metamorphic core complexes as interpreted in the Basin and Range and the Aegean Sea. A strong lower crust (colder or stronger) leads to the formation of a flexural core complex for which little lower crustal exhumation occurs. Our models suggest that detachment faults develop in hot young orogens where strong fluids/magma activity and a preexisting weak zone or décollement exist.

Acknowledgements

We thank Roger Buck and Whitney Behr for valuable comments on our earlier manuscript. We also thank Nicholas Hayman, Daniel Stockli, and Richard Katcham for helpful

discussions. We are grateful to reviewers Gary Axen and Cristiano Collettini for their constructive comments and suggestions that help to improve the paper. The work is partially funded by King Abdullah University of Science and Technology (KAUST) Global Collaborative Research. The raw data is available, and can be requested by e-mail (glwu@utexas.edu).

Author Manuscript

References

- Anderson, E. M. (1951), *The Dynamics of Faulting*, Oliver and Boyd, Edinburgh.
- Axen, G. J. (1992), Pore pressure, stress increase, and fault weakening in low-angle normal faulting, *Journal of Geophysical Research*, 97(B6), 8979-8991, doi:10.1029/92JB00517.
- Axen, G. J. (1999), Low-angle normal fault earthquakes and triggering, *Geophysical Research Letters*, 26(24), 3693-3696, doi:10.1029/1999GL005405.
- Banks, C. J., and J. Warburton (1991), Mid-crustal detachment in the Betic system of southeast Spain, *Tectonophysics*, 191(3), 275-289, doi:10.1016/0040-1951(91)90062-W.
- Bartley, L. M., and B. P. Wernicke (1984), The Snake Range decollement interpreted as a major extensional shear zone, *Tectonics*, 3(6), 647-657, doi:10.1029/TC003i006p00647.
- Behr, W. M., and J. P. Platt (2011), A naturally constrained stress profile through the middle crust in an extensional terrane, *Earth Planet. Sci. Lett.*, 303(3-4), 181-192, doi:10.1016/j.epsl.2010.11.044.
- Bellanger, M., N. Bellahsen, L. Jolivet, T. Baudin, R. Augier, and A. Boutoux (2014), Basement shear zones development and shortening kinematics in the Ecrins Massif, Western Alps, *Tectonics*, 33(2), 84-111, doi:10.1002/2013TC003294.
- Bird, F. (1991), Lateral extrusion of lower crust from under high topography in the isostatic limit, *J. Geophys. Res.*, 96(B6), 10275-10210,10286, doi:10.1029/91JB00370.
- Bird, K. B., W. E. Stewart, and E. N. Lightfoot (2007), *Transport phenomena*, 2nd ed. John-Wiley & Sons., New York.
- Block, L., and L. H. Royden (1990), Core complex geometries and regional scale flow in the lower crust, *Tectonics*, 9(4), 557-567, doi:10.1029/TC009i004p00557.
- Brown, M., and G. S. Solar (1998), Shear-zone systems and melts: feedback relations and self-organization in orogenic belts, *J. Struct. Geol.*, 20(2), 211-227, doi:10.1016/S0191-8141(97)00068-0.
- Buck, W. R. (1988), Flexural rotation of normal faults, *Tectonics*, 7(5), 959-973, doi:10.1029/TC007i005p00959.
- Buck, W. R., L. L. Lavie, and A. N. Poliakov (2005). Modes of faulting at mid-ocean ridges. *Nature*, 434(7034), 719-723, doi:10.1038/nature03358

- Bürgmann, R. and G. Dresen (2008). Rheology of the lower crust and upper mantle: Evidence from rock mechanics, geodesy, and field observations. *Ann. Rev. Earth Planet. Sci.*, 36(1), 531-567, doi: 10.1146/annurev.earth.36.031207.124326
- Chiaraluce, L., C. Chiarabba, C. Collettini, D. Piccinini, and M. Cocco (2007), Architecture and mechanics of an active low-angle normal fault: Alto Tiberina Fault, northern Apennines, Italy, *J. geophys. Res.*, 112, B10310, doi:10.1029/2007JB005015.
- Choi, E. and W. R. Buck (2012), Constraints on the strength of faults from the geometry of rider blocks in continental and oceanic core complexes, *J. Geophys. Res.*, 117(B4), doi:10.1029/2011JB008741.
- Choi, E., W. R. Buck, L. L. Lavier, and K. D. Petersen (2013), Using core complex geometry to constrain fault strength, *Geophys. Res. Lett.*, 40(15), 3863-3867, doi:10.1002/grl.50732.
- Collettini, C. (2011), The mechanical paradox of low-angle normal faults: Current understanding and open questions, *Tectonophysics*, 510(3), 253-268, doi:10.1016/j.tecto.2011.07.015.
- Collettini, C., N. De Paola, R. Holdsworth, and M. Barchi (2006), The development and behaviour of low-angle normal faults during Cenozoic asymmetric extension in the Northern Apennines, Italy, *J. Struct. Geol.*, 28(2), 333-352, doi:10.1016/j.jsg.2005.10.003.
- Collettini, C., A. Niemeijer, C. Viti, S. A. Smith, and C. Marone (2011), Fault structure, frictional properties and mixed-mode fault slip behavior, *Earth and Planetary Science Letters*, 311(3), 316-327, doi:10.1016/j.epsl.2011.09.020.
- Collettini, C., A. Niemeijer, C. Viti, and C. Marone (2009), Fault zone fabric and fault weakness, *Nature*, 462(7275), 907-910, doi:10.1038/nature08585.
- Coney, P. J., D. L., Jones, and J. W. Monger (1980), Cordilleran suspect terranes, *Nature*, 288, 329-333, doi:10.1038/288329a0.
- Davis, G. H. (1980), Structural characteristics of metamorphic core complexes, southern Arizona, *Geol. Soc. Am. Mem.*, 153, 35-78, doi:10.1130/MEM153-p35.
- Davis, G. H. (1983), Shear-zone model for the origin of metamorphic core complexes, *Geology*, 11(6), 342, doi:10.1130/0091-7613(1983)11<342:SMFTOO>2.0.CO;2.
- Davis, G. H., and P. J. Coney (1979), Geologic development of the Cordilleran metamorphic core complexes, *Geology*, 7(3), 120-124, doi:10.1130/0091-7613(1979)7<120:GDOTCM>2.0.CO;2.

- DeMets, C., R. G. Gordon, and D. F. Argus (2010), Geologically current plate motions, *Geophys. J. Int.*, 181(1), 1-80, doi:10.1111/j.1365-246X.1990.tb06579.x.
- Dewey, J. F., and J. M. Bird (1970), Mountain belts and the new global tectonics, *J. Geophys. Res.*, 75(14), 2625-2647, doi:10.1029/JB075i014p02625.
- Dingus, W. R. (2002), The Basin and Range province as a composite extensional domain, *International Geology Review*, 44(1), 1-38, doi:10.2747/0020-6814.44.1.1.
- Faulkner, D., T. Mitchell, D. Healy, and M. Heap (2006), Slip on 'weak' faults by the rotation of regional stress in the fracture damage zone, *Nature*, 444(7121), 922-925, doi:10.1038/nature05353.
- Forryth, D. W. (1992). Finite extension and low-angle normal faulting. *Geology*, 20(1), 27-30, doi: 10.1130/0091-7613(1992)020<0027:FEALAN>2.3.CO;2
- Gars, P. B., and W. A. Bohrsen (1998), Suppression of volcanism during rapid extension in the basin and range province, united states, *Science*, 279(5347), 66-68, doi:10.1126/science.279.5347.66.
- Gars, P. B., G. A. Mahood, and E. Schermer (1989), Synextensional magmatism in the Basin and Range Province; a case study from the eastern Great Basin, *Geological Society of America Special Papers*, 233, 1-53, doi:10.1130/SPE233-p1.
- Gautier, P., J.-P. Brun, and L. Jolivet (1993), Structure and kinematics of upper Cenozoic extensional detachment on Naxos and Paros (Cyclades Islands, Greece), *Tectonics*, 12(5), 1180-1194, doi:10.1029/93TC01131.
- Gautier, P., J. P. Brun, R. Moriceau, D. Sokoutis, J. Martinod, and L. Jolivet (1999), Timing, kinematics and cause of Aegean extension: a scenario based on a comparison with simple analogue experiments, *Tectonophysics*, 315(1), 31-72, doi:10.1016/S0040-1951(99)00281-4.
- Gessner, K., U. Ring, C. Johnson, R. Hetzel, C. W. Passchier, and T. Güngör (2001), An active bivergent rolling-hinge detachment system: Central Menderes metamorphic core complex in western Turkey, *Geology*, 29(7), 611-614, doi:10.1130/0091-7613(2001)029<0611:AABRHD>2.0.CO;2.
- Gueydan, F., Y. M. Leroy, and L. Jolivet (2004), Mechanics of low-angle extensional shear zones at the brittle-ductile transition, *J. Geophys. Res.*, 109(B12), B12407, doi:10.1029/2003JB002806.

- Hayman, N.W., J.R. Knott, D.S. Cowan, E. Nemser, and A.M. Sarna-Wojcicki (2003). Quaternary low-angle slip on detachment faults in Death Valley, California. *Geology*, 31(4), 343-346, doi: 10.1130/0091-7613(2003)031<0343:QLASOD>2.0.CO;2
- Hetzel, R., C. W. Passchier, U. Ring, and Ö. O. Dora (1995), Bivergent extension in orogenic belts: the Menderes massif (southwestern Turkey), *Geology*, 23(5), 455-458, doi:10.1130/0091-7613(1995)023<0455:BEIOBT>2.3.CO;2.
- Jackson, J., and N. White (1989), Normal faulting in the upper continental crust: observations from regions of active extension, *J. Struct. Geol.*, 11(1-2), 15-36, doi:10.1016/0191-8141(89)90033-3.
- Jolivet, L., and J. P. Brun (2010), Cenozoic geodynamic evolution of the Aegean, *Int. J. Earth Sci.*, 99(1), 109-138, doi:10.1007/s00531-008-0366-4.
- Jolivet, L., C. Faccenna, B. Huet, L. Labrousse, L. Le Pourhiet, O. Lacombe, E. Lecomte, E. Burov, Y. Denèle, and J.-P. Brun (2013), Aegean tectonics: Strain localisation, slab tearing and trench retreat, *Tectonophysics*, 597, 1-33, doi:10.1016/j.tecto.2012.06.011.
- Jolivet, L., E. Lecomte, B. Huet, Y. Denèle, O. Lacombe, L. Labrousse, L. Le Pourhiet, and C. Mehl (2010). The north cycladic detachment system. *Earth Planet. Sci. Lett.*, 289(1), 87-104, doi:10.1016/j.epsl.2009.10.032
- Ketchum, R. A. (1996), Thermal models of core-complex evolution in Arizona and New Guinea: Implications for ancient cooling paths and present-day heat flow, *Tectonics*, 15(5), 933-951, doi:10.1029/96TC00033.
- Kirby, S., and A. Kronenberg (1987), Rheology of the lithosphere: selected topics, *Rev. Geophys.*, 25(6), 1219-1244, doi:10.1029/RG025i006p01219.
- Klemperer, S. L., T. Hauge, E. Hauser, J. Oliver, and C. Potter (1986), The Moho in the northern Basin and Range province, Nevada, along the COCORP 40 N seismic-reflection transect, *Geol. Soc. Am. Bull.*, 97(5), 603-618, doi:10.1130/0016-7606(1986)97<603:TMITNB>2.0.CO;2.
- Kounov, A., E. Wüthrich, D. Seward, J.-P. Burg, and D. Stockli (2015), Low-temperature constraints on the Cenozoic thermal evolution of the Southern Rhodope Core Complex (Northern Greece), *Int. J. Earth Sci.*, 104(5), 1337-1352, doi:10.1007/s00531-015-1158-2.

- Lavier, L. L., and W. R. Buck (2002), Half graben versus large-offset low-angle normal fault: Importance of keeping cool during normal faulting, *J. Geophys. Res.*, 107(B6), 2122, doi:10.1029/2001JB000513.
- Lavier, L. L., W. R. Buck, and A. N. B. Poliakov (2000), Factors controlling normal fault offset in an ideal brittle layer, *J. Geophys. Res.*, 105(23), 431–423, doi:10.1029/2000JB900108.
- Lavier, L. L., and G. Manatschal (2006), A mechanism to thin the continental lithosphere at magma-poor margins, *Nature*, 440(7082), 324–328, doi:10.1038/nature04608.
- Lavier, L. L., W. Roger Buck, and A. N. B. Poliakov (1999), Self-consistent rolling-hinge model for the evolution of large-offset low-angle normal faults, *Geology*, 27(12), 1127–1130, doi:10.1130/0091-7613(1999)027<1127:SCRHMF>2.3.CO;2.
- Lecomte, E., L. Jolivet, O. Lacombe, Y. Denèle, L. Labrousse, and L. Le Pourhiet (2010). Geometry and kinematics of Mykonos detachment, Cyclades, Greece: Evidence for slip at shallow dip. *Tectonics*, 29(5), doi: 10.1029/2009TC002564
- Lister, G. S., and G. A. Davis (1989), The origin of metamorphic core complexes and detachment faults formed during Tertiary continental extension in the northern Colorado River region, USA, *J. Struct. Geol.*, 11(1-2), 65–94, doi:10.1016/0191-8141(89)90036-9.
- McCarthy, J., S. P. Larkin, G. S. Fuis, R. W. Simpson, and K. A. Howard (1991), Anatomy of a metamorphic core complex: Seismic refraction/wide-angle reflection profiling in southeastern California and western Arizona, *J. Geophys. Res.*, 96(B7), 12259–12291, doi:10.1029/91JB01004.
- McKenzie, D., F. Nimmo, J. A. Jackson, P. Gans, and E. Miller (2000), Characteristics and consequences of flow in the lower crust, *J. Geophys. Res.*, 105(B5), 11,029–011,046, doi:10.1029/1999JB900446.
- Melosh, H. (1990), Mechanical basis for low-angle normal faulting in the Basin and Range province, *Nature*, 343, 331–335, doi:10.1038/343331a0.
- Miller, E. L., P. B. Gans, and J. Garing (1983), The Snake Range décollement: An exhumed Mid-Tertiary ductile-brittle transition, *Tectonics*, 2(3), 239–263, doi:10.1029/TC002i003p00239.
- Miller, M., and T. Pavlis (2005), The Black Mountains turtlebacks: Rosetta stones of Death Valley tectonics, *Earth Sci. Rev.*, 73(1-4), 115–138, doi:10.1016/j.earscirev.2005.04.007.

- Nagel, T. J., and W. R. Buck (2006), Channel flow and the development of parallel-dipping normal faults, *J. Geophys. Res.*, *111*(B8), B08407, doi:10.1029/2005JB004000.
- Oliot, E., P. Goncalves, K. Schulmann, D. Marquer, and O. Lexa (2013), Mid-crustal shear zone formation in granitic rocks: Constraints from quantitative textural and crystallographic preferred orientations analyses, *Tectonophysics*, *612-613*, 63-80, doi:10.1016/j.tecto.2013.11.032.
- Platt, J. P., and W. M. Behr (2011), Grainsize evolution in ductile shear zones: Implications for strain localization and the strength of the lithosphere, *J. Struct. Geol.*, *33*(4), 537-550, doi:10.1016/j.jsg.2011.01.018.
- Platt, J. P., W. M. Behr, and F. J. Cooper (2014), Metamorphic core complexes: windows into the mechanics and rheology of the crust, *J. Geol. Soc.*, *172*(1), 9-27, doi:10.1144/jgs2014-036.
- Platt, J. P., W. M. Behr, K. Johanesen, and J. R. Williams (2013), The Betic-Rif arc and its orogenic hinterland: A review, *Annu. Rev. Earth Planet. Sci.*, *41*, 313-357, doi:10.1146/annurev-earth-050212-123951.
- Polakov, A., P. Cundall, Y. Podladchikov, and V. Lyakhovsky (1993), An explicit inertial method for the simulation of viscoelastic flow: an evaluation of elastic effects on diapiric flow in two-and three-layers models, *NATO ASI Ser. C Math. Phys. Sci.*, *391*, 175-175, doi:10.1007/978-94-015-8206-3_12.
- Ranalli, G., 1995. *Rheology of the Earth*, 2nd ed. Chapman and Hall, London.
- Ranalli, G., and D. C. Murphy (1987), Rheological stratification of the lithosphere, *Tectonophysics*, *132*(4), 281-295, doi:10.1016/0040-1951(87)90348-9.
- Rehing, W. A., and S. J. Reynolds (1980), Geologic and geochronologic reconnaissance of a northwest-trending zone of metamorphic core complexes in southern and western Arizona, *Cordilleran metamorphic core complexes: Geol. Soc. Am. Mem.*, *153*, 131-157, doi:10.1130/MEM153-p131.
- Rey, P., C. Teyssier, and D. Whitney (2009), Extension rates, crustal melting, and core complex dynamics, *Geology*, *37*(5), 391, doi:10.1130/G25460A.1.
- Sibson, R. (1977), Fault rocks and fault mechanisms, *Journal of the Geological Society*, *133*(3), 191, doi:10.1144/gsjgs.133.3.0191.

- Singleton, J. S. (2013), Development of extension-parallel corrugations in the Buckskin-Rawhide metamorphic core complex, west-central Arizona, *Geol. Soc. Am. Bull.*, 125(3-4), 453-472, doi:10.1130/B30672.1.
- Smith, S., R. Holdsworth, C. Collettini, and M. Pearce (2011), The microstructural character and mechanical significance of fault rocks associated with a continental low-angle normal fault: the Zuccale Fault, Elba Island, Italy, *Geol. Soc., London, Spec. Pub.*, 359(1), 97-113, doi:10.1144/SP359.6.
- Sodoudi, F., R. Kind, D. Hatzfeld, K. Priestley, W. Hanka, K. Wylegalla, G. Stavrakakis, A. Vafidis, H. P. Harjes, and M. Bohnhoff (2006), Lithospheric structure of the Aegean obtained from P and S receiver functions, *J. Geophys. Res.*, 111(B12), doi:10.1029/2005JB003932.
- Spencer, J. E. (1984), Role of tectonic denudation in warping and uplift of low-angle normal faults, *Geology*, 12(2), 95-98, doi:10.1130/0091-7613(1984)12<95:ROTDIW>2.0.CO;2.
- Spencer, J. E., and C. G. Chase (1989), Role of crustal flexure in initiation of low-angle normal faults and implications for structural evolution of the Basin and Range province, *J. Geophys. Res.*, 94(B2), 1765-1775, doi:10.1029/JB094iB02p01765.
- Tan, E., L. L. Lavier, H. J. Van Avendonk, and A. Heuret (2012), The role of frictional strength on plate coupling at the subduction interface, *Geochem., Geophys., Geosys.*, 13(10), doi:10.1029/2012GC004214.
- Wdowinski, S., and G. J. Axen (1992), Isostatic rebound due to tectonic denudation: A viscous flow model of a layered lithosphere, *Tectonics*, 11(2), 303-315, doi:10.1029/91TC02341.
- Wernicke, B. (1985), Uniform-sense normal simple shear of the continental lithosphere, *Can. J. Earth Sci.*, 22(1), 108-125, doi:10.1139/e85-009.
- Wernicke, B. (1992), Cenozoic extensional tectonics of the US Cordillera, in *The Cordilleran orogen: Conterminous US*, edited by B. C. Burchfiel, P. W. Lipman and M. L. Zoback, pp. 553-581, *Geol. Soc. Am.*, Boulder, Colorado.
- Wernicke, B. (1995), Low-angle normal faults and seismicity: A review, *J. Geophys. Res.*, 100(B10), 20159-20174, doi:10.1029/95JB01911.
- Wernicke, B., and G. J. Axen (1988), On the role of isostasy in the evolution of normal fault systems, *Geology*, 16(9), 848-851, doi:10.1130/0091-7613(1988)016<0848:OTROI>2.3.CO;2.

- Westaway, R. (1999), The mechanical feasibility of low-angle normal faulting, *Tectonophysics*, 308(4), 407-443, doi:10.1016/S0040-1951(99)00148-1.
- Whitney, D. L., C. Teyssier, P. Rey, and W. R. Buck (2013), Continental and oceanic core complexes, *Geol. Soc. Am. Bull.*, 125(3-4), 273-298, doi: 10.1130/B30754.1
- Whitney, D. L., C. Teyssier, and O. Vanderhaeghe (2004), Gneiss domes and crustal flow, *Geol. Soc. Am. Spec. Pap.*, 380, 15-33, doi:10.1130/0-8137-2380-9.15.
- Wills, S., and W. R. Buck (1997), Stress-field rotation and rooted detachment faults: A Coulomb failure analysis, *J. Geophys. Res.*, 102(B9), 20503-20514, doi:10.1029/97JB01512.
- Wu, G., L. L. Lavier, and E. Choi (2015), Modes of continental extension in a crustal wedge, *Earth Planet. Sci. Lett.*, 421, 89-97, doi:10.1016/j.epsl.2015.04.005.
- Xiao, K. B., F. Dahlen, and J. Suppe (1991), Mechanics of extensional wedges, *Journal of Geophysical Research*, 96(B6), 10,301-310,318, doi:10.1029/91JB00222.
- Yin, A. (1989), Origin of regional, rooted low-angle normal faults: A mechanical model and its tectonic implications, *Tectonics*, 8(3), 469-482, doi:10.1029/TC008i003p00469.
- Zuber, M., E. Parmentier, and R. Fletcher (1986), Extension of continental lithosphere: A model for two scales of Basin and Range deformation, *J. Geophys. Res.*, 91(B5), 4826-4838, doi:10.1029/JB091iB05p04826.

Figures and Figure Captions

Figure 1. Geological and geophysical reconstructions of core complexes in the US Cordillera and the Aegean showing the characteristic detachment faults, overlying upper-crustal normal faults, and underlying exhumed lower crust: (a) Harcuvar Mountains [Buck, 1988; after *Rehrig and Reynolds*, 1980]; (b) Whipple Mountains in California, with inferred dome on the left side removed [Buck, 1988; after *Davis*, 1980]; (c) the Snake Range [after *Miller et al.*, 1983]; (d) SW Rhodope massif [after *Gautier et al.*, 1999]; (e) the Central Cyclades [after *Gautier et al.*, 1999]; (f) Crete [after *Gautier et al.*, 1999]. The orange band in each panel denotes a shear zone. In all of these cross sections, the Moho is sub-horizontal.

Figure 2. (a) Idealized shear zone and lower crust as two immiscible fluids between two rigid plates of upper crust and mantle. The scaled velocity profile (blue) in the weak shear zone (10^{19} Pa s) and lower crust (10^{20} Pa s) is shown with magnitude being greater toward the right side. The profile for the lower crust with the same thickness but without a weak shear zone is shown in green. (b) Mean velocity of lower crust as a function of viscosity of the lower crust for the case with a weak shear zone (blue), for that of without a shear zone, and ratio of mean velocity of lower crust for the case with a shear zone with respect to that of without a weak shear zone (10^{19} Pa s). The red dashed line indicates a tectonic velocity of $6 \times 10^{-10} \text{ m s}^{-1}$ ($\sim 2 \text{ cm yr}^{-1}$). The velocity is calculated with 2 km for h_{ss} , and 16 km for h_{lc} .

Figure 3. (a) Model setup. The shear zone thickness is initially 2 km. The temperature for the top surface is 10°C , and bottom 800°C . There is no heat flow boundary condition on either side of the model. The heavy arrows indicate horizontal divergent boundary velocity. (b) Schematic strength profiles across the left side of the modeled crustal wedge based on the parameters in Table 1, with three characteristic strengths of lower crust: (1) wet quartz with partial melting, (2) wet quartz, and (3) plagioclase. (c) Plot of temperature in the mid-crustal shear zone, at 25 km depth in the lower crust, and Moho after 0.1 Myr of model run with intermediate-strength lower crust. (d) Plot of viscosity in the mid-crustal shear zone, at 25 km depth in the lower crust, and in the mantle along the Moho discontinuity for the same model as in c after 0.1 Myr of model run.

Figure 4. Three distinct detachment systems and their associated hanging wall normal faults based on numerical simulations. Principal stress is non-Andersonian in the shear zone and in the regions where topographic gradient is large, but is still largely vertical in the upper crust. (a) A type-I detachment fault consists of a detachment fault residing in the brittle domain or close to the surface that bears ductile inheritance, and a ductile shear zone as a continuation of the brittle shear zone. (b) A type-II detachment fault is a listric normal fault soling into a detachment fault residing in the brittle domain and extending as a ductile shear zone a deeper depth. (c) A type-III detachment fault is a subsurface fault zone, which has a subhorizontal dip in the brittle domain

and transitions slowly to a ductile shear zone at depth. In all three panels, σ_1 denotes the maximum principal stress for an element in the shear zone.

Figure 5. Snap shots of velocity vector and principal stress (σ_1) in a numerical model with relatively strong lower crust and a weak shear zone. (a) Velocity vectors show that the upper crust and lower crust is decoupled. (b) The σ_1 significantly deviates from vertical direction (non-Andersonian state of stress) in regions adjacent to the shear zone and in regions with high topographic gradient (range front), but is still largely Andersonian in most of the upper crust. The plastic strain inset shows the brittle low-angle detachment fault with its associated hanging wall normal faults.

Figure 6. Typical core complexes and detachment systems generated by the numerical models. (a) A bivergent core complex with an exhumed metamorphic massif (type-I). (b) A metamorphic core complex or classic Cordilleran core complex (numbered 2) including a type-I detachment system and a classic asymmetric lower crustal dome. Another core complex (numbered 1) has a type-II detachment system and a lower crustal dome not yet exhumed to the surface. (c) Boudinage structures consisting of several consecutive core complexes with typical type-II detachment systems. (d) The type-III detachment system overlying a flexural core complex with lower crustal dome exhibiting a very small curvature.

Figure 7. Bivergent core complex. The model has a weak lower crust and a weak mid-crustal shear zone (frictional angle, 5°). It develops bivergent detachment faults (Type-I) and a large massif (or batholith). The black stars trace the same particle in the shear zone, whose thermal history is plotted in Figure 11.

Figure 8. Metamorphic core complex. The model has an intermediate-strength lower crust (wet quartz regime) and a weak mid-crustal shear zone (frictional angle, 5°). It generates the classic Cordilleran core complex (Red rectangle in c) as well as type-I and type-II detachment faults. The black stars trace the same particle in the shear zone, whose thermal history is plotted in Figure 11.

Figure 9. Boudinage structures. The model has an intermediate-strength lower crust (wet quartz regime) and a strong mid-crustal shear zone (frictional angle, 25°). It develops a series of small

core complexes as well as type-I and type-II detachment faults. The black stars or dots trace the same particles in the shear zone, whose thermal history is plotted in Figure 11.

Figure 10. Flexural core complex. The model has a strong lower crust (dry quartz regime), and a weak mid-crustal shear zone (frictional angle, 5°). It generates a subsurface detachment fault (Type-III) and a series of domino style hanging wall normal faults. The black stars trace the same particle in the shear zone, whose thermal history is plotted in Figure 11.

Figure 11. Thermal history of sampled particles in the shear zones of modeled core complexes: (i) Metamorphic massif in the model shown in Figure 8 (star); (ii) Metamorphic core complex in the model shown Figure 9 (star); (iii) Metamorphic core complex in the model shown in Figure 9 (star); (iv) Boudinage structure in the model shown in Figure 9 (dot); (v) Flexural core complex in the model shown in Figure 10 (star).

Figure 12. Mohr-Coulomb analysis with a preexisting weak subhorizontal shear zone. The yield stress of the shear zone has crossed the Mohr-Coulomb (MC) circle and failed, while the failure envelope in the hanging wall has not yet crossed. The heterogeneity allows the low-angle detachment fault residing below the hanging wall to slip as long as the hanging wall has not yielded.

Table 1. Summary of key model parameters.

Phase	ρ (kg m ⁻³)	θ (°)	θ' (°)	C (MPa)	A (MPa ^{n} s ⁻¹)	E_a (J mol ⁻¹)	n	C_p (J Kg ⁻¹ K ⁻¹)	h_r (W kg ⁻¹)
Upper crust	2700	30	15	40	0.125	2.76 x 10 ⁵	3.05	1000	10 ⁻⁹
Weak lower crust	2800	30	15	40	0.125	1.8 x 10 ⁵	3.05	1000	10 ⁻⁹
Intermediate-strength lower crust	2800	30	15	40	0.125	2.4 x 10 ⁵	3.05	1000	10 ⁻⁹
Strong lower crust	2800	30	15	40	0.125	2.76 x 10 ⁵	3.05	1000	10 ⁻⁹
Mafic lower crust	3000	30	15	40	0.125	3.5 x 10 ⁵	3.05	1000	10 ⁻⁹
Mid-crustal shear zone	2800	5-25	5-25	4	0.125	1.76 x 10 ⁵	3.05	1000	10 ⁻⁹
Mantle	3300	30	15	40	7.0 x 10 ⁴	5.2 x 10 ⁵	3.0	1000	-

Abbreviation in column headings: ρ (density), θ (initial frictional angle), θ' (frictional angle after weakening), C (cohesion), A (coefficient in the creep law), E_a (activation energy in creep law), n (exponential component in the creep law), C_p (heat capacity), and h_r (radioactive heating).

See main text for references.

2016tc004245-f01-z-.eps

2016tc004245-f02-z-.eps

pt

Al

2016tc004245-f03-z-.eps

2016tc004245-f04-z-.eps

2016tc004245-f05-z-.eps

ipt

AL

2016tc004245-f07-z-.eps

ipt

Au

2016tc004245-f08-z-eps

ipt

Au

2016tc004245-f09-z-eps

ipt

Au

2016tc004245-f10-z-eps

2016tc004245-f11-z-.eps

2016tc004245-f12-z-.eps

Minerva Access is the Institutional Repository of The University of Melbourne

Author/s:

Wu, G;Lavier, LL

Title:

The effects of lower crustal strength and preexisting midcrustal shear zones on the formation of continental core complexes and low-angle normal faults

Date:

2016-09-01

Citation:

Wu, G. & Lavier, L. L. (2016). The effects of lower crustal strength and preexisting midcrustal shear zones on the formation of continental core complexes and low-angle normal faults. *TECTONICS*, 35 (9), pp.2195-2214. <https://doi.org/10.1002/2016TC004245>.

Persistent Link:

<http://hdl.handle.net/11343/291677>