Metamorphic core complexes: windows into the mechanics and rheology of the crust

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Abstract: Metamorphic core complexes are products of normal-fault displacements sufficient to exhumation of rocks from below the brittle–ductile transition. These faults (detachments) may initiate within the brittle crust at steep angles, but they sole into the ductile middle crust, and during displacement rotate to gentler dips due to hanging-wall extension. The exhumed footwall commonly adopts an arched or domed geometry owing to flexural isostatic readjustment, and may be overlain by strongly extended upper crustal rocks that slipped on gently dipping, low-friction shallow segments of the detachment. Metamorphic rocks exhumed beneath the detachment record progressively increasing flow stress, strain localization and strain-rate with decreasing temperature, providing a window into physical conditions and deformational processes in the mid-crust. The metamorphic and deformational history of the footwall rocks may reflect tectonic processes that predate formation of the detachment fault, in addition to those accompanying exhumation. These processes may include diapiric emplacement of gneiss domes, or exhumation in a subduction channel, and may not be directly related to formation of the core complex. Factors favouring core complex formation are high gravitational potential energy of the extending crust, weak rheology and a change in the tectonic boundary conditions such as a cessation or slowing of plate convergence.

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Metamorphic core complexes (MCCs) have attracted a great deal of interest since they were first identified around 1980, partly because their distinctive structural and tectonic style raised a number of fundamental questions (and quite a bit of scepticism), but also because of what they reveal about the mechanical properties of the crust. Because their formation results in the rapid tectonic exhumation of rock from the ductile middle crust, through the brittle–ductile transition (BDT), and on to shallow crustal levels, they provide well-constrained natural rock-mechanical experiments conducted under geological conditions of strain-rate and temperature. Studies of metamorphic core complex have substantially influenced our understanding of the mechanics of brittle faulting, the nature of the BDT, the causes of strain localization in the ductile field, strain-rates and flow stresses in ductile shear zones, and the progressive changes in deformational mechanisms and microstructure with increasing depth and temperature in the crust. This paper was motivated by the feeling that a synthesis of this new information is overdue, and by the need to place the rheological and microstructural concepts involved in a tectonic context. Much of what we have learnt bears in a fundamental way on issues such as the load-bearing capacity of the continental lithosphere, its response to deformation and the resulting changes in its thermal and mechanical structure, and the way deformation is partitioned, distributed and transferred across the thermal and rheological layering within the lithosphere.

In view of the strong focus of this paper on these mechanical and rheological questions, we have not attempted a comprehensive synthesis and review of core complexes world-wide. Whitney et al. (2013) have already presented a very thorough review of this nature, covering both oceanic and continental core complexes and their relationship to gneiss domes, and we refer the reader to their paper for this type of synthesis. Here we focus specifically on a number of mechanical and geometrical issues concerning the formation of MCCs, and on the rheological information provided by the syntectonic deformation that accompanies exhumation.

What is a metamorphic core complex?

The concept of the metamorphic core complex was first developed in the North American Cordillera, and had its seeds in the observation that medium- to high-grade metamorphic rocks in the hinterland of the Cordilleran thrust belt form dome-shaped complexes directly overlain by low-grade or unmetamorphosed sedimentary sequences. Armstrong & Hansen (1966), for example, described these relationships and interpreted them in terms of a ductile ‘infrastructure’ overlain by a competent and brittle ‘suprastructure’ from which it is separated by a regionally developed gently dipping tectonic discontinuity decorated by mylonite and cataclasite. These discontinuities were variously described as thrusts, unconformities, décollements, denudation faults and gravity slides; but by 1980 a number of workers had recognized that they cut out parts of the tectonic or stratigraphic sequence, that they had some of the characteristics of normal faults, and that they were a response to horizontal extension and vertical thinning of the crust after Cordilleran thrusting had ceased (e.g. Davis & Coney 1979; Davis et al. 1980; Crittenden 1980; Wernicke 1981). Framing a definition of a metamorphic core complex, however, was complicated by the fact the structures that came to be grouped under this heading had a wide variety of structural forms and petrological characteristics (e.g. Armstrong 1982; Coney & Harms 1984; Davis et al. 1986). As the concept came to be applied outside the Cordillera, it became even broader, and it has come to encompass extensional structures formed at mid-ocean ridges (e.g. Ranero & von Huene 2000) and metamorphic complexes exhumed in subduction and collisional settings. In its broadest sense, it has been used to refer to almost any body of rock that has experienced some form of ductile deformation, and that has been exhumed in such a way that it is in tectonic contact with lower-grade or unmetamorphosed rock. No definition will satisfy all the practitioners in the field; the following one is based on our personal experience, and is strongly influenced by discussions with G. A. Davis.
A metamorphic core complex comprises three essential elements (Fig. 1). From bottom to top these are as follows.

1. A core of metamorphic rock, commonly 10 km or more across, affected by ductile deformation and associated metamorphic recrystallization, and derived from the mid-crust or deeper.

2. The detachment is overlain by a regionally gently dipping to subhorizontal tectonic discontinuity, commonly referred to as a detachment, comprising a discrete brittle fault surface, several metres to tens of metres of cataclastic rocks in its immediate footwall, underlain in turn by a zone of ductile mylonite and ultramylonite, which may be hundreds to thousands of metres in thickness, grading downwards into the main metamorphic core (Davis et al. 1980, 2004).

3. The detachment is overlain by hanging-wall rocks that are either unmetamorphosed or of significantly lower grade than the metamorphic core, and that are commonly strongly disrupted and attenuated by normal faulting.

The detachment places upper crustal rocks against metamorphic or plutonic rocks exhumed from beneath the BDT, and hence produces vertical thinning and horizontal extension. It is now universally accepted that MCCs are the product of a distinctive style of extensional tectonics. As with all definitions, there are qualifications and exceptions. Some core complexes are cored largely by plutonic igneous rocks, including granite in some continental core complexes, such as the Colville batholith (Holder & Holder 1988), and gabbro in oceanic core complexes (Miranda & Dilek 2010). Fault rocks (mylonite and cataclasite) vary substantially in thickness and character from one complex to another, and the relationship between mylonites and the detachment is vigorously debated; this issue is discussed further in the section ‘Mylonites in MCCs’. The hanging wall in some complexes may include metamorphic and other crystalline rocks: the definitive characteristic is that they were cool and in the upper crust by the time they were involved in the formation of the core complex. In some core complexes the hanging-wall sequence is largely or entirely made up of volcanic or sedimentary rocks deposited during slip on the detachment.

As noted above, the concept has been applied very widely, and there are gradations into structures that would not generally be regarded as MCCs. The Shuswap Complex in the hinterland of the Canadian Cordillera, for example, includes several domiform bodies cored by migmatic gneiss and granite, commonly referred to as gneiss domes. These structures were central to the development of the concept of MCCs (Armstrong 1982; Brown & Read 1983), but they share many of the characteristics of gneiss domes recognized in Phanerozoic orogens and high-grade Precambrian terrains worldwide (Whitney et al. 2013). The origin of gneiss domes is also much debated, and it is likely that many of them are formed by processes other than horizontal extension. The relationship between MCCs and gneiss domes is discussed further in the section ‘MCCs and gneiss domes’.

A comparable problem is posed by MCCs in many convergent margin settings that are cored by high-pressure and ultrahigh-pressure metamorphic rocks. These were probably exhumed part way by other processes, before being involved in core-complex tectonism. This issue is discussed further in the section ‘Metamorphism in MCCs’. In our opinion, the term metamorphic core complex should not be applied to: (1) bodies of rock that have clearly been emplaced primarily by intrusion or diapirism, including plutons, salt domes and diapiric gneiss domes; (2) metamorphic complexes emplaced in the upper crust by thrust, reverse or strike-slip faults, or by normal faults that are entirely brittle and post-date any ductile deformation in the footwall; (3) metamorphic cores of orogens that lack any evidence for a bounding detachment or for extensional exhumation; (4) normal faults, gently dipping or otherwise, that have not exhumed rocks from below the BDT.

The metamorphic cores of collisional orogens present particular problems in this respect. Here are some further examples that illustrate the difficulty of defining what is and what is not a core complex.

The High Himalayan Crystalline complex is bounded to the north by large-scale low-angle normal-sense faults and shear zones of the South Tibetan fault system (Hodges et al. 1998; Dèzes et al. 1999; Searle 1999; Cooper et al. 2012). These faults resemble core complex detachments, and clearly played an important role in exhuming the metamorphic rocks beneath them. They appear to have been active more or less simultaneously with the Main Central Thrust, which forms the lower boundary of the High Himalayan Crystallines (e.g. Grasemann et al. 1999). For this reason we would not normally consider this major element of Himalayan geology as a core complex. On the other hand, the North Himalayan gneiss domes, including the Tso Morari, Kangmar and Mabja domes, which lie north of the South Tibetan fault system, show many of the characteristics of core complexes (Lee et al. 2000; de Sigoyer et al. 2004; Aoya et al. 2005). It has been suggested that these gneiss domes are bounded by upwarped sections of the South Tibetan fault system detachments (e.g. Hauck et al. 1998), and it is clear that at some depth beneath them there is a currently active megathrust along which the Indian plate is being subducted beneath Tibet. This is an example where we may need to be flexible in our definition of a core complex.

A similar problem arises in the Alps. The Tauern window in the eastern Alps exposes medium-grade rocks and locally eclogites with early Tertiary metamorphic ages (e.g. Glodny et al. 2008). It is overlain by sedimentary and metamorphic rocks of the Austroalpine complex, which are generally regarded to have been cold, and to have formed a rigid lid, by early Tertiary time. The rocks in the Tauern window were exhumed in part by slip on a major normal fault (the Brenner Line) that outlines the western end of the window, and that has many of the characteristics of a detachment fault (Selverstone 1988). A similar situation exists in the central Alps: medium- to high-grade metamorphic rocks, including relic eclogites, in the Lepontine Dome reached peak temperatures at around 30 Ma (Vance & O’Nions 1992) and were exhumed in part along the west-directed Simplon fault to the west (Campani et al. 2010) and the east-directed Turba fault to the east, both of which exhibit many of the features of detachment faults (Nieergelt et al. 1996). Few workers consider these
large volumes of metamorphic rocks as core complexes, however, presumably because they were exhumed during continuing continental collision.

**Detachment faults**

*Nature and geometry of detachment faults*

The large-scale geometry of detachment faults above MCCs has been visualized in three significantly different ways, and these can be illustrated using the evolution of ideas about the northern Snake Range MCC in the central US Cordillera (Fig. 2). Miller *et al.* (1983), following the earlier concept of a tectonic boundary separating an ‘infrastructure’ from a ‘suprastructure’ (Armstrong & Hansen 1966), suggested that the detachment in the northern Snake Range formed as a subhorizontal tectonic boundary at the BDT, separating a zone of ductile thinning below from a zone of brittle normal faulting above (Fig. 2a). They envisaged the detachment rising towards the Earth’s surface as a result of thinning of the overlying brittle layer, but they thought that the fault accumulated little displacement in the process. In their concept, brittle normal faults in the hanging wall would have transferred displacement onto the detachment, but the detachment itself may never have broken the surface while it was active. Bartley & Wernicke (1984), by contrast, suggested that the detachment formed as a through-going normal fault, reaching the surface to the west of the Snake Range, and cutting through the crust and possibly the whole lithosphere with an average dip of about 30° (see also Wernicke 1981), with a displacement of c. 80 km (Fig. 2b). Following suggestions by Davis & Lister (1988) and Wernicke & Axen (1988) that detachments flatten out into middle crustal mylonite zones, Cooper *et al.* (2010a) used thermobarometry to show that there is no significant gradient in peak metamorphic pressure in footwall rocks in a transport-parallel direction across the Snake Range, and hence that the deeper part of the detachment was in fact subhorizontal in the mid-crust (Fig. 2c).

*Mechanics of low-angle normal faults*

Gently dipping or regionally subhorizontal extensional faults are widely recognized in extended terrains; detachments in core complexes are distinguished mainly by the fact that they have sufficient displacement to exhume rocks from below the BDT. Active normal faults with dips less than 30° (low-angle normal faults or LANFs) are widely perceived as presenting a mechanical problem, as the resolved shear stress on the fault plane has to be substantially less than that required to overcome frictional resistance with a coefficient μ typical of experimentally determined values for rock-on-rock friction (μ = 0.6–0.8). Relatively few seismic events occur on normal faults with dips <30° (Jackson 1987), although a number of examples have been documented (Rigo *et al.* 1996; Abbott *et al.* 1997; Abers *et al.* 1997; Heimsdottir & Bennett 2009).

Various explanations for slip on LANFs have been proposed, including high pore-fluid pressure (e.g. Axen 1992), lubrication by hydrous minerals, pre-existing rheological controls (e.g. Ral River shear zone, Wells 2001), or reorientation of the stress field in response to particular tectonic environments. The mechanical problem with slip on LANFs has been largely resolved by the realization in recent years that most mature brittle faults move under low shear stresses (e.g. Townend 2006), largely owing to the low coefficients of friction of smectite clays (μ = 0.1–0.2) (e.g. Holdsworth *et al.* 2011; Carpenter *et al.* 2012; Schleicher *et al.* 2012), which can be an important component of fault gouge at temperatures less than 200 °C. A coefficient of friction of 0.1 could allow a normal fault to slip at a dip of 10° or less. This also helps explain the lack of seismicity on LANFs, as smectite clays show velocity strengthening behaviour during slip, which favours creep, rather than seismic failure (Carpenter *et al.* 2012).

**Initiation of low-angle normal faults: domino and rolling-hinge models**

Low frictional strength of existing LANFs does not account for their initiation: normal faults propagating through intact isotropic rock are likely to form with an initial dip of around 60°. An important insight into the problem of how LANFs could initiate was reached by Proffett (1977), who showed in the Yerington district of Nevada that sets of normal faults formed with initial dips of around 60°, and were then progressively rotated to lower angles during continued extension. Once the dip of the faults had been reduced to <30°, they ceased to be active, and were cut by new faults with greater dips. As many as three generations of normal faults formed during a period of a few million years, and the earliest faults in some cases were rotated into the horizontal. The angle between initially horizontal volcanic units and the steep faults was

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**Fig. 2.** Evolution of ideas about the nature of detachment faults, based on different published models for the northern Snake Range. (a) After Miller *et al.* (1983); (b) after Bartley & Wernicke (1984); (c) after Cooper *et al.* (2010a). Not drawn to scale.
largely preserved, allowing the history of slip and rotation to be determined. This domino-style pattern of fault rotation reflects the difference in bulk vorticity between simple shear on an array of normal faults and horizontal coaxial extension, which is dictated by the requirement that the Earth’s surface remain roughly horizontal during deformation.

Some detachments may have reached low dips by this mechanism (e.g. the Sierra Mazatán, Wong & Gans 2008), but many regionally subhorizontal detachments have dimensions of tens of kilometres in the direction of motion, without any significant cross-cutting faults. A second fundamental insight into this problem was reached simultaneously by Buck (1988), Hamilton (1988) and Wernicke & Axen (1988), who independently proposed that isostatic rebound during unloading by slip on normal faults will rotate the footwall from an initially steep orientation into the horizontal, or even to the point that the direction of dip is reversed. Buck (1988, 1993) and Lavier et al. (1999) modelled this process (Fig. 3), and showed that the length scale of the isostatic response is mediated by the effective elastic thickness of the lithosphere, which in regions of extension may be fairly low (~10km) (e.g. Lowry et al. 2000). As modelled by Lavier et al. (1999), the active part of the fault remains steeply dipping, but as the footwall rocks reach the surface, they tilt into a subhorizontal orientation. This process has come to be known as the rolling-hinge model, as it implies that rocks in the footwall of the normal fault roll around a hinge close to the Earth’s surface into a gentle dip. This model helps resolve the mechanical issues about the initiation of the normal fault, as it can form at the mechanically favoured dip of 60°, and it explains how the exposed faults come to be subhorizontal. The effect of elastic flexure also helps explain the dome-like geometry, as isostatic rebound can rotate the detachments through the horizontal. The Black Mountains on the eastern side of Death Valley may provide an example of this process still in operation (Holm et al. 1993). Palaeomagnetic evidence from both continental (Holm et al. 1993; Livaccari & Geissman 2001) and oceanic core complexes (Garcés & Gee 2007; Morris et al. 2009) provides strong support for substantial tilting of detachment foothills.

If the ductile roots of detachments are also subhorizontal, as discussed above, there may be two rolling hinges, one at depth where the subhorizontal ductile shear zone curves into the steep orientation of the active brittle fault zone, and the other at the surface, as illustrated for the Snake Range in Figure 2c. This double-hinged detachment geometry may be supported by thermo-chronological data in some examples, such as the northern Snake Range MCC (Lee 1995) and the Buckskin–Rawhide MCC (Scott et al. 1998), but detailed structural analysis of footwall rocks has failed to find conclusive evidence for this repeated flexure (Axen & Bartley 1997). Several workers have suggested that the inclined part of the fault plane may not dip much more than 30° (e.g. Lister & Davis 1989; John & Foster 1993), which would greatly reduce the amount of strain associated with the rolling hinges and still satisfy the mechanics of the brittle faulting, if we assume a coefficient of friction of c. 0.4. This opens the possibility that the brittle section of the detachment may initiate with a mechanically favourable dip of 50° or more (Fig. 4a), and the hanging-wall side of the fault then rapidly rotates to a dip of c. 30° as a result of horizontal stretching of the overlying rock sequence (Fig. 4b). Continued motion on the fault rotates the footwall around the upper rolling hinge into a subhorizontal orientation, together with stranded fragments of hanging-wall rocks, as envisaged by Buck et al. (1988) (Fig. 4c).

Numerous studies have documented deformed hanging-wall sequences above subhorizontal detachments, in which normal fault structures appear to be truncated downwards onto the detachment (e.g. Davis et al. 1980; Davis 1988; Scott & Lister 1992; Konstantinou et al. 2012). Bedding in hanging-wall rocks commonly dips steeply, consistent with tilting by either the domino or the rolling-hinge mechanism. In places, however, there are subhorizontal sedimentary sequences lying on a horizontal detachment, and these may include conjugate sets of normal faults with steep dips. The sediments involved may be pre-extensional cover sequences, as in the Raft River MCC (Konstantinou et al. 2012) or they may include syn-extensional volcanic and sedimentary rocks, as in the Whipple Mountains (Davis et al. 1980). These observations suggest that in some cases the shallow parts of detachment faults were active with dips of perhaps 10°, requiring very low coefficients of friction, as discussed above. Reconstruction of these extended hanging-wall sequences suggests that they formed by slip on arrays of listric normal faults that soled down onto a gently dipping detachment surface (Fig. 5). This process has been modelled by Hayman et al. (2003) in terms of an extensional critical wedge. These extended sequences of hanging-wall rocks come from the location where the detachment reaches the surface (known as the breakaway, Fig. 1), and this section of the detachment may develop a ramp-flat geometry analogous to thrust faults developed in sedimentary sequences.

The net result of these processes is that hanging-wall sequences, deformed above a gently dipping section of the detachment at shallow depth, may come to lie on a footwall section of the fault that has been rotated around the upper rolling hinge from a 30° dip into a gently dipping orientation, which in turn lies on mid-crustal mylonites with a subhorizontal foliation that formed with this orientation below the lower rolling hinge (Fig. 4). These paradoxes have been responsible for several decades of confusion.

**Core complexes with multiple or conjugate detachments**

Multiple detachments have been recorded from some MCCs. In the Valhalla dome of the Shuswap Complex, the lower amphibolite-facies Valkyr shear zone has been arched over the dome in response to motion on the greenschist-facies Slocan Lake normal fault on the east side (Parrish et al. 1988) (Fig. 6a). Both shear zones have the same top-to-east sense, and are interpreted as extensional. In the western Betic Cordillera, the Ronda peridotite massif appears to have been exhumed beneath a set of stacked ductile to brittle detachments, the lowest of which defines the upper surface of the peridotite, and the highest separates a greatly attenuated sequence of crustal metamorphic rocks from unmetamorphosed rocks above (Platt et al. 2003; Fig. 6b).
Many MCCs are bounded by detachments with opposing shear senses, which may or may not be coeval. The Santa Catalina core complex in Arizona is bounded by shear zones that displace away from the complex on either side and appear to be coeval (Naruk & Bykerk-Kaufman 1990). The Albion-Raft River-Grouse Creek Complex is bounded on the west by a relatively high-T west-directed shear zone, of probable late Eocene to early Oligocene age (Salzer & Hodges 1988; Wells et al. 2000; Gottardi & Teyssier 2013), and on the east side by the younger east-directed Raft River detachment, of mid- to late Miocene age (Konstantinou et al. 2012). The Shuswap Complex in the Canadian Cordillera is bounded by the west-dipping Okanagan detachment on the west side, and the somewhat younger east-dipping Columbia River fault on the east (Parrish et al. 1988; Fig. 6c), and more complex patterns are exhibited by the Menderes massif in Turkey (Gessner et al. 2001), and the Priest River complex in the NW USA (Doughty & Price 1999).

The implications of rolling-hinge models for multiple and conjugate detachments are clearly complicated (e.g. Gessner et al. 2001). Motion in the same sense on successive detachments could in principle cause large tilts; motion on detachments with opposite shear senses could result in very small net tilts. In some of the examples cited above, however, such as the Albion-Raft River-Grouse Creek Complex and the Valhalla dome, early ‘detachments’ are in fact normal-sense ductile shear zones that did not exhume rocks to the BDT; these may have been gently dipping over their whole trajectory, and hence may not have caused significant tilting (Simony & Carr 1997).

![Fig. 4. Suggested evolution of a detachment fault. (a) Initiation of detachment along a steep (50° dip) trajectory through the brittle crust. The detachment cuts through the brittle–ductile transition (BDT), and soles out at the transition from localized to distributed ductile deformation (LDT). (b) Rapid reduction in dip (to 30°) of the detachment, by horizontal stretching of the upper plate. Upper plate structure shown schematically. (c) Rollover of the detachment into a subhorizontal orientation following the rolling-hinge model, with stranded hanging-wall slivers as envisaged by Buck (1988). Ductile middle crust rolls up along the detachment at the lower rolling hinge, creating a dome-shaped metamorphic core. True scale (V = H).](image1)

![Fig. 5. (a) Section through the hanging wall of the detachment in the SW Whipple Mountains (from an unpublished section by G. A. Davis). (b) Restoration, showing that deformation resulted from slip on a listric normal fault (probably the breakaway fault) that soled down onto a horizontal section of the detachment, which may have been short (<1 km). The entire section was subsequently transported onto a section of the detachment derived from deeper in the crust, where it would have had a dip of 30° or more, but rolled around the upper rolling hinge into its present horizontal orientation. This is underlain in turn by mylonites that formed below the lower rolling hinge with a gently dipping fabric; these rolled around both hinges, and now dip gently SW.](image2)
Core complexes with no breakaway

In some MCCs, a breakaway has not been positively identified (e.g. the Ruby Mtns), and it is not entirely clear whether the detachment did in fact break the Earth’s surface. This question is particularly pertinent in some of the MCC's in the Mediterranean region, which have very large areal extents. In the Betic Cordillera of southern Spain, for example, the subducted southern margin of Iberia (known as the Nevada–Filabride Complex) has been exhumed in a series of linked domal core complexes (Martínez-Martínez et al. 2002; Fig. 7). The detachment can be traced for 180 km in the direction of motion (west to WSW), and no breakaway has been found. The footwall and hanging-wall rocks have independent tectonic and metamorphic histories, and no offset markers can be identified. If the detachment is regarded as a normal fault that cut down from the surface, these observations would require a displacement on the fault of >180 km. Zircon fission-track ages from the footwall differ by only 3 Ma over this distance (Johnson & Harbury 1997), which would imply a remarkably high rate of motion. The alternative explanation is that the detachment formed along the upper boundary of the subduction channel, within which most of the exhumation of the Nevada–Filabride Complex took place, and that the final stages of exhumation were achieved by widely distributed normal faulting of the hanging-wall sequence (Behr & Platt 2013). This explanation does not require the detachment to reach the surface, and allows more or less simultaneous exhumation and cooling of the footwall over its whole trajectory. The implication is that the detachment is not a normal fault in the usual sense, although it formed as a result of horizontal extension and vertical shortening. Normal faults in the hanging wall, which varied in their direction of motion, transferred displacement onto the detachment, so that the direction and amount of displacement is likely to vary from place to place over the detachment surface.

If detachments (and their footwalls) can be exhumed wholly or in part by thinning of the hanging-wall sequence, this has important implications for rates of slip calculated using thermochronological data. Several papers have presented data from samples collected along traverses parallel to the slip direction, and have calculated rates of slip assuming that the exhumation was entirely a result of slip on the detachment. Some of the calculated rates are high, (e.g. c. 4 mm a⁻¹ on the Buckskin–Rawhide detachment in Arizona; Brady 2002); and in some cases they substantially exceed the likely overall rate of tectonic extension in the region as a whole (e.g. 30 mm a⁻¹ on the Bullard detachment, Arizona; Carter et al. 2004). As pointed out by Wells et al. (2000), hanging-wall extension will modify the pattern of cooling ages, and can lead to erroneously high estimates of slip rate.

Mylonites in MCCs

An essential characteristic of MCCs is that they exhume rocks from below the BDT, and they generally show a zone a few tens to hundreds of metres thick of relatively high-strain non-coaxial ductile deformation in the footwall of the detachment. The rocks in these zones of high strain typically show evidence of crystal plastic deformation and dynamic recrystallization in quartz, and contain porphyroclasts of feldspar and mica fish. They are generally referred to as mylonites or mylonitic gneisses, and they commonly (but not always) show kinematic indicators of the shear sense consistent with the motion on the brittle detachment. The precise relationship between the mylonitic rocks and the brittle detachment has proved somewhat controversial, however. In many core complexes, radiometric ages dating crystallization or cooling events in the mylonites suggest that the ductile deformation was older than the displacement on the brittle detachment, as determined on stratigraphic grounds. Discordant relationships between the brittle detachment and the underlying mylonites are not uncommon.

These questions can be well illustrated from the Whipple Mountains MCC in SE California. Here, the brittle detachment is locally discordant to the underlying mylonitic foliation (Davis et al. 1980; Fig. 8). On the west side of the Whipple Mountains, the mylonitic foliation dips moderately west, and disappears beneath non-mylonitic footwall rocks along a surface described by Davis & Lister (1988) as the mylonitic front (Fig. 8a). The detachment dips more gently, and can be traced westward to a breakaway where it reached the Miocene ground surface; the mylonitic front on the other hand can be traced seismically in the subsurface (Wang et al. 1989), where it descends to a depth of c. 10 km. These relationships...
led Davis & Lister (1988) to suggest that the mylonites are not directly related to the detachment, but were “captured” by the detachment as it descended below the BDT. The mylonites have the same kinematics as the brittle detachment, but they predate it by a few million years. Both the brittle detachment and the mylonites can reasonably be related to the overall process of core complex formation (Davis & Lister 1988; Behr & Platt 2011).

In some MCCs, mylonites may predate the development of the core complex altogether. In the Nevado–Filabride Complex of southern Spain, early mylonites appear to have been formed along the upper surface of a subduction channel that accommodated the initial stages of exhumation (Behr & Platt 2013). These mylonites are crosscut by the detachment, which is underlain by a concordant layer of late ultramylonite with different kinematics.

In most MCCs, the metamorphic core has one or more sets of foliations, folds and lineations, some of which may entirely predate the tectonic event that produced the MCC. The dominant set of structures may be related to the processes of crustal thickening, subduction and exhumation that immediately predated, and created the required conditions for, core complex formation. In this situation it may be difficult to make a clear distinction between structures and fabrics directly attributable to core complex formation and those...
that are not. In the Albion–Raft River–Grouse Creek Complex, for example, high-strain fabrics have been variously attributed to an Eocene west-directed detachment (Saltzer & Hodges 1988; Gottardi & Teyssier 2013), or to diapiric emplacement of an Oligocene pluton (Konstantinou et al. 2012). In the Ruby–East Humboldt MCC, fabrics in the high-grade gneissic core have been attributed to early stages of motion on the detachment (McGrew et al. 2000), but these fabrics formed under amphibolite-facies conditions, with different kinematics (MacCready et al. 1997), and at depths well below the BDT (McGrew et al. 2000). Mylonites spatially related to the detachment formed at lower temperatures, and are separated in time from the earlier structures by a suite of 29Ma biotite monzonite dykes and sheets (MacCready et al. 1997).

In both the northern Snake Range and the Ruby–East Humboldt MCCs, the interpretation of Ar–Ar chronotours (lines on the map connecting rocks with the same Ar–Ar cooling age) has led to some confusion about the timing of motion on the detachment and the metamorphic and deformational history of the core rocks. In the northern Snake Range, chronotours trend north–south, and young from 50 to 20Ma from west to east, as described by Lee & Sutter (1991) and Lee (1995). This led those researchers to suggest that the chronotours reflect westward tilting beneath the east-dipping detachment, which was therefore inferred to have been active over this period of time. A similar pattern of chronotours is documented in the Ruby–East Humboldt MCC, trending NNE–SSW, and younging from 56 to 21Ma from ESE to WNW. These have been interpreted to indicate eastward tilting beneath the west-dipping detachment over this period. In both cases the thermochronological data can be alternatively interpreted as a result of slow cooling after Late Cretaceous to Eocene metamorphism, followed by relatively rapid displacement and tilting during the Miocene. This is more consistent with the evidence from sedimentary basins in the region for rapid displacement on normal faults in the Miocene (Henry et al. 2011), and the evidence that early Tertiary fabrics in the metamorphic cores formed at depth in the mid-crust, and are geometrically unrelated to the detachment (e.g. MacCready et al. 1997; Cooper et al. 2010b).

In the section ‘Core complexes as windows into the rheology of the deep crust’ we make the case that many MCCs have a relatively narrow zone of high-stress mylonite that is genetically related to the detachment, underlain by a broader zone of high-strain rock that formed in the mid-crust below the detachment, and in some cases before the detachment was initiated.

Domiform geometry

A striking feature of metamorphic core complexes is their domiform geometry, which is commonly defined both by the geometry of the detachment and the mylonitic foliation beneath. Two types of dome can be distinguished: those that are elongate parallel to the direction of motion on the detachment, and those that are elongate normal to this direction (Jolivet et al. 2004); some are almost circular in present-day outcrop. The domes appear to reflect warping about axes both normal and parallel to the direction of motion on the detachment. A number of explanations for this geometry have been proposed, all of which are likely to be true in some cases.

The simplest explanation for warping of the detachment about an axis normal to the direction of motion arises from the rolling-hinge model. Isostatic uplift and flexure of the footwall and detachment create a broad antiform normal to the direction of motion (Fig. 3), and possibly a synform close to the breakaway (Lavier et al. 1999). Alternative explanations include rollovers related to later normal faults, or boudinage of a stiff crustal layer during extension. The first explanation may account for many of the elongate core complexes in the US Cordillera, where extending Basin and Range extension has modified core complexes such as the Snake Range and Ruby–East Humboldt MCCs, warping the detachments down towards younger range-bounding normal faults (e.g. Henry et al. 2011). The second explanation has been invoked to explain the geometry of some of the detachments in the Aegean (Jolivet et al. 2004). Multiple extension-normal folds have been reported from the Hohhot MCC in Inner Mongolia (China) by Davis & Darby (2010).

Folds and corrugations parallel to the slip direction are also common in detachments. Some of these may reflect the primary geometry of the active fault: Spencer (1999) has suggested that the detachment footwall may have such corrugations imprinted in it by a process akin to continuous casting or extrusion, and this seems a likely explanation for the widely documented corrugations in oceanic core complexes (Cann et al. 1997). Davis (1988) has shown that corrugations in the mylonitic gneisses of the Whipple Mtns are cross-cut by the overlying nearly planar brittle detachment, suggesting that they formed within the ductile shear zone below the BDT. Alternative explanations include horizontal shortening normal to the motion direction, possibly concurrent with motion (Holm et al. 1994; Braathen et al. 2004), which would imply a constrictional deformation field, or shortening owing to a later, unrelated phase of contractional tectonics. The second explanation may apply in the Betic Cordillera, where some of the east–west-trending domes in the detachment above the Nevada–Filabride Complex (Fig. 7) are a result of a latest Miocene phase of north–south contraction, unrelated to the formation of the detachment in the early Eocene (Konstantinou 2012). In the Ruby–East Humboldt MCC, trending NNE–SSW, and younging from 56 to 21Ma from ESE to WNW, these have been interpreted to indicate eastward tilting beneath the west-dipping detachment over this period. In both cases the thermochronological data can be alternatively interpreted as a result of slow cooling after Late Cretaceous to Eocene metamorphism, followed by relatively rapid displacement and tilting during the Miocene. This is more consistent with the evidence from sedimentary basins in the region for rapid displacement on normal faults in the Miocene (Henry et al. 2011), and the evidence that early Tertiary fabrics in the metamorphic cores formed at depth in the mid-crust, and are geometrically unrelated to the detachment (e.g. MacCready et al. 1997; Cooper et al. 2010b).

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Metamorphism in MCCs

MCCs show a wide variety of metamorphic histories, and these have given rise to substantial debate about the relationship between metamorphism and core complex formation. Many Cordilleran MCCs show Barrovian-style metamorphism that is plausibly related to the Late Cretaceous peak of crustal thickening in the Cordilleran orogen, and is associated with partial melting, migmatization and granitic magmatism (e.g. Parrish et al. 1988; Miller & Gans 1989; Lee 1995; Sullivan & Snake 2007). The likely heat source for metamorphism is radioactive heat generation in the thickened crust, possibly accentuated by thinning of the underlying lithosphere in the back-arc region behind the subduction zone along the western margin of North America (Sonder et al. 1987; Ranalli et al. 1989; Currie & Hyndman 2006; Wells & Hoisch 2008). In this case there is no direct connection between Late Cretaceous metamorphism and the formation of the MCCs, except that the processes that caused the metamorphism (crustal thickening and removal of lithospheric mantle) may have increased the gravitational potential energy that drove the later extension (e.g. Jones et al. 1996), and thermal weakening and partial melting associated with the metamorphism facilitated deformation. The story in the US Cordillera is complicated, however, by petrological and structural evidence for multiple stages of decompression and burial during the Late Cretaceous (e.g. Applegate & Hodges 1995; Camilleri & Chamberlain 1997; Wells et al. 2005; Harris et al. 2007), which have been interpreted as reflecting alternating stages of extension and shortening in a critical orogenic wedge or thickened plateau subject to varying tectonic boundary conditions (Wells et al. 2012). Cooling, and possibly exhumation, during the Eocene is clearly indicated by thermochronological data in many of the US core complexes (e.g. Saltzer & Hodges 1988; McGrew & Snee 1994; Lee 1995; Henry et al. 2011), and is well documented in the Canadian MCCs (e.g. Vanderhaeghe et al. 2003; Gordon et al. 2008;
Kruckenberg et al. (2008), where it has been attributed to diapiric emplacement of the migmatitic gneiss complexes, followed by motion on extensional detachments (Parrish et al. 1988). There is only limited evidence for upper crustal extension south of the Snake River Plain during the early Tertiary, however (Rahl et al. 2002; Druschke et al. 2009; Henry et al. 2011; Miller et al. 2012), and Eocene exhumation appears to predate motion on the detachment faults in the central and southern Cordillera, most of which have been dated to between 25 and 16Ma (Dokka et al. 1986; Davis 1988; Miller et al. 1999; Carter et al. 2004, 2006; Colgan et al. 2010; Henry et al. 2011; Konstantinou et al. 2012).

The metamorphic evolution of many of the MCCs identified in the Alpine–Himalayan system more closely reflects plate-boundary processes. In the western Mediterranean and in the Aegean Sea, core complexes exhumed blueschist- and locally eclogite-facies rocks formed by Tertiary subduction of continental margin or microcontinental basement and sedimentary sequences. In most cases exhumation to depths of c. 20 km was rapid, occurred soon after subduction, and was most probably driven by buoyancy forces within the subduction channel or accretionary wedge where convergence was continuing (Avigad et al. 1997; Ring et al. 2009; Behr & Platt 2013). The MCCs formed subsequently, in a back-arc setting as the subduction zone retreated, and exhumed the partly retrogressed high-pressure rocks through the BDT. Unravelling the precise sequence and timing of events, and relating the metamorphic history to the structure, has proved difficult and controversial.

In the Himalayan orogen, the Tso Morari dome is distinguished from the other North Himalayan gneiss domes by the presence of eclogite-facies rocks representing the early stages of collision (de Sigoyer et al. 2000). These rocks appear to have returned to the surface in the subduction channel, and were then exhumed to the surface in a structure that resembles an MCC (de Sigoyer et al. 2004).

In the Woodlark Basin, north of Papua New Guinea in the western Pacific, the D’Entrecasteaux Islands have long been described as active metamorphic core complexes (e.g. Hill et al. 1992). Sea-floor spreading in the basin at around 40 mm a⁻¹ (Abers et al. 1997) separates the Woodlark microplate from the Australian plate, and the rift is propagating westwards into the continental collision zone that accommodated Cenozoic convergence between Australia and New Guinea. Intracontinental extension in this region has exhumed blueschist- and eclogite-facies rocks from up to 90 km depth during the last 8 Ma (Baldwin et al. 2008). These high-pressure rocks are now exposed on the D’Entrecasteaux Islands (Fig. 9), which form a series of domes bounded on the north side by north- to NE-dipping recent or active normal faults forming part of the rift system (Little et al. 2007). The exposed surfaces of similar active normal faults in the basin have been imaged using side-scan sonar (Speckbacher et al. 2011), and seismic imaging shows that they dip 15–35° (Abers et al. 1997) in the subsurface. The islands form a series of nearly circular to ovoid domes, exposing high-pressure metamorphic rocks in their cores, surrounded by a mylonitic carapace, and overlain by remnants of the Papuan ophiolite complex, which forms the upper plate of the original subduction system (Hill et al. 1992; Davies & Warren 1998; Little et al. 2007). The mylonitic carapaces to the domes show north to ENE shear senses, broadly consistent with the sense of motion on the active normal faults that bound the domes on the north side. The rate of exhumation determined by thermobarometry and geochronology on the eclogites is 10–12 mm a⁻¹ (Baldwin et al. 2008), consistent with the rate of plate divergence, if accommodated by normal faults dipping 25° (Webb et al. 2008). The gneisses show evidence for high-temperature metamorphism and decompressional melting during their exhumation, followed by cooling and mylonitic deformation. Little et al. (2007) have suggested that these gneiss domes were emplaced as diapirs. It is clear that the rocks were partially molten and behaving as low-viscosity fluids at depths of around 30 km, and exhumation from 90 to 30 km depth may have been diapirc, but the later stages of exhumation were facilitated by large-displacement normal sense shear zones forming the mylonitic carapace, and ultimately by brittle normal faults.

In the Western Gneiss Region of the Scandinavian Caledonides, the Laurentian margin was subducted during the Scandinavian orogeny and then exhumed en masse beneath major detachments overlain by Devonian sedimentary basins (e.g. Andersen 1998; Hacker & Gans 2005). The exhumation of ultrahigh-pressure eclogites, locally carrying coesite and diamond, within the Western Gneiss Region, was probably accomplished in part by diapiric ascent to the base of the orogenic crust, followed by detachment faulting (Root et al. 2004; Walsh & Hacker 2004; Johnston et al. 2007).

Pressure–temperature paths from the metamorphic cores of MCCs are plotted in Figure 10 (metamorphic events that clearly predate and are completely unrelated to the evolution of the MCC are
(1) As discussed above, extension on detachment faults with a rolling-hinge geometry can produce domiform structures cored by rocks that have been exhumed from at least 15–20 km depth (Fig. 4), and this provides an adequate explanation for many MCCs.

(2) Granitic crust at a temperature of 700–800 °C can deform under low stresses (c. 1 MPa) at a rate (e.g. c. $10^{-14} \text{s}^{-1}$) sufficient for it to form a diapir, driven by buoyancy forces, that could rise through the crust on a time scale of c. 10 Ma. The presence of significant granitic melt would greatly increase this rate. On this time scale the upward motion of a structure tens of kilometres across will be close to isothermal until it reaches depths <20 km, and is likely to be accompanied by decompressional melting. Hence it is plausible that domes cored by high-grade gneiss, migmatitic gneiss, or foliated granite could be emplaced by this mechanism, and we would call the resulting structure a gneiss dome. Diapirs are in general likely to have steep or even overturned margins, and predominantly steeply dipping foliation and steeply plunging lineations and folds in the core of the structure, although the top of the diapir can experience significant vertical shortening and develop a flat-lying foliation (e.g. Dixon 1975; Jackson & Talbot 1989).

(3) Even in the absence of buoyancy forces, if the lower crust is at or above the solidus, rifting can result in the upward flow of the fluid lower crustal material, driven by the difference between the lithostatic pressure in the fluid and the reduced horizontal normal stress associated with extension in the overlying rock. This could generate a structure that might be structurally indistinguishable from a diapirc gneiss dome, but is a consequence of regional extension. This process has been referred to as isostasy-driven flow (Tirel et al. 2008; Rey et al. 2011), but the resulting structures are probably best referred to as piercement structures, a non-genetic term long used to describe the upward motion of relatively low-viscosity materials through the crust.

The various gneiss domes that make up the Shuswap Complex in the Canadian Cordillera (Fig. 6a and c) are cored by high-grade granitic gneiss, migmatite and granite that reached temperatures of 750–800 °C at pressures of around 1 GPa (e.g. Norlander et al. 2002). Crystallization of melt took place between 50 and 60 Ma (e.g. Gordon et al. 2008; Kruckenberg et al. 2008). These bodies show metamorphic evidence for isothermal decompression and partial melting (Norlander et al. 2002), followed by rapid cooling in the period 56–48 Ma (Vanderhaeghe et al. 2003). It is therefore plausible that they were emplaced as diapirs or as piercement structures.

Several features of the internal structure of the Shuswap domes are not fully consistent with this interpretation, however. Foliation is for the most part gently dipping within the domes, and the foliation dip is spatially related to the normal faults that bound many of the domes on all sides (e.g. Crowley et al. 2001; Fig. 6a and c). These normal faults, which include the Okanagan detachment, the Columbia River fault and Valkyr–Slocan shear zone, show many of the characteristics of detachment faults, including gentle dips, an upward transition from ductile to brittle microstructures, and estimated displacements of >30 km (Parish et al. 1988). In addition, large-displacement gently dipping shear zones interpreted as east-directed Mesozoic thrust structures occur within or around several of these domes. The Monashee décollement, for example, places metamorphic rocks of the Selkirk allochthon above the migmatitic gneisses of the Frenchman Cap and Thor–Odin domes (Fig. 6c). This structure is thought to have as much as 80 km displacement, and may link contractional structures in the metamorphic hinterland with those of the thin-skinned Rocky Mountain thrust belt to the east (Crowley et al. 2001). The Gwillim Creek shear zones may have played a similar role in the Valhalla dome (Parish et al. 1988). These structures, as well as early extensional shear zones, are folded around the open antiforms defined by the main detachments.

The close association between the large-scale antiforms defining the domes and the bounding normal detachments suggests that the present-day geometry is largely a result of detachment faulting, and that they can reasonably be described as MCCs (Armstrong 1982; Brown & Read 1983; Brown & Murray Journeay 1987; Parish et al. 1988; Crowley et al. 2001). As pointed out by Teysseier & Whitney (2003), however, earlier structures related to diapirism may have been flattened and partly transposed by subsequent ductile vertical thinning.
Metamorphic core complexes

Another very interesting example is the island of Naxos in the Aegean Sea (Fig. 11). The outer part of the structure resembles a classic Cordilleran-style MCC, with a detachment fault separating unmetamorphosed upper crustal rocks from a core with a varied and complex metamorphic history (Lister et al. 1984). As in much of the Aegean, there are traces of early blueschist metamorphism in the core (Avigad 1998), but in the deeper parts of the complex this was overprinted by Barrovian metamorphism, which caused extensive partial melting. The inner core is largely made up of migmatic gneiss, with tight upright foliations on various scales, and dominantly steeply dipping foliation, and steeply plunging lineations and fold hinges (Kruckenberg et al. 2011; Fig. 11). The inner core is in fact divided into three sub-domes, separated by screens of gneiss with a strong, steeply dipping foliation (Kruckenberg et al. 2011). These features are entirely consistent with an origin as a diapir or piercement structure for the inner core. The boundary between the inner core and the surrounding medium-grade metamorphic rocks, which have a more gently dipping foliation, shows evidence for strong NNE-directed shear, consistent with the motion on the detachment that bounds the whole complex (Kruckenberg et al. 2011). Rey et al. (2011) suggested, on the basis of a modelling study, that the multiple sub-dome structure on Naxos could result from the horizontal inflow of mobile material at depth into the rising dome, driven primarily by pressure differences produced by rifting (see above). Naxos can therefore be regarded as a prime example of a structure with a composite origin, involving a phase of diapirism or piercement in the early Miocene (Keay et al. 2001), followed by detachment faulting in the late Miocene (Seward et al. 2009).

As noted above, Little et al. (2011) made a case for a diapiric origin for the eclogite-cored domes in the D’Entrecastaux Islands. A major part of the exhumation trajectory for these rocks took place within the mantle, as they rose from c. 90 to c. 30 km depth, and the predominantly quartzofeldspathic gneisses making up the domes were at temperatures close to or above the solidus during that stage of exhumation. A buoyancy-driven mechanism is therefore very reasonable. The structural case for diapirism is not very strong, however: the domes show a clear geometrical relationship to the normal faults that bound them (Fig. 9), and the present domiform geometry developed at a late stage, after a period of vertical shortening that produced a flat-lying foliation at 30–40 km depth (Little et al. 2011).

Magnatism and MCCs

Many Cordilleran MCCs contain substantial bodies of granite and other plutonic rocks in their cores, and this led to an early perception that extension was facilitated, permitted, or driven by magma emplacement (Armstrong 1982; Gans 1987; Gans et al. 1989). It was subsequently discovered that most of the large plutons are Mesozoic (Miller & Gans 1989; Henry et al. 2011), whereas the main phase of core complex formation is Tertiary. There was in fact considerable volcanic activity during the mid-Tertiary in the US Cordillera, accompanied by the emplacement of suites of dykes with compositions ranging from basaltic to rhyolitic (e.g. Gans et al. 1989), but Gans & Bohrson (1998) made the case that during periods of rapid extension magma tended to crystallize at depth, thereby inhibiting volcanic activity. The documented volumes of Tertiary intrusive rocks are relatively minor compared with the scale of the extensional tectonics during this period.

In the Canadian Cordillera the high-grade gneiss cores in the Shuswap Complex contain abundant migmatic and anatectic granite of Eocene age, and there appears to be a close association between partial melting and core complex formation. There too, however, the major plutons are of Cretaceous age, and single plutons are not closely associated with core complexes.

MCCs in the Mediterranean region largely lack any direct association with plutons, although small granite intrusions pre-, syn- and post-kinematic with the detachments are widespread. Much the same is true of most of the MCCs developed in the Alpine–Himalayan system, including those that are cored by high-pressure and ultrahigh-pressure metamorphic rocks.

In summary, it appears that although continental MCCs commonly contain substantial plutons, as might be expected from their location within orogenic belts, there is no direct cause-and-effect relationship between magmatism and core complex formation. Partial melting undoubtedly facilitates tectonic activity, and extension and decompression may trigger partial melting, but many core complexes lack evidence for substantial syntectonic magmatism.

The situation in oceanic core complexes may be different. Mid-ocean ridges are the locus of voluminous magmatism, and oceanic MCCs are likely to form at ridge–transform intersections, for example (Ranero & Reston 1999), close to the locus of magmatism. Tucholke et al. (2008) suggested that if melt emplacement in the form of dykes and plutons accounts for 50% or more of the total rate of extension, then detachment faults that initiate at the ridge axis can accumulate large displacements; if melt emplacement accounts for <50% of extension, then the detachment becomes broken up by younger faults, and does not accumulate much displacement. This implies a very close relationship between magmatism and the formation of detachment faults; a relationship that clearly does not apply in a continental environment.

Driving forces for core complex formation

The defining feature of a core complex is the brittle–ductile detachment fault, which is responsible for a vertical component of displacement of the order of 10 km and a horizontal component of several tens of kilometres. This displacement clearly implies regional horizontal extension and vertical thinning of at least the upper 20–25 km of crust. This may be coeval with lower crustal flow (as suggested for the North American Cordillera), or continuing subduction and underthrusting of continental crust, as in many of the core complexes of the Alpine–Himalayan system. Most MCCs are found in orogenic systems, and crustal extension appears to be a late or post-contractional phenomenon.

Fig. 11. East–west section across Naxos, showing the relationships between the detachment, the metamorphic envelope and the migmatitic gneiss dome (modified from Vanderhaeghe (2004) and Kruckenberg et al. (2011)). It should be noted that the section is normal to the motion direction on the detachment (to the north); this is to illustrate the upright folding in the migmatitic core.
The following are some of the ideas that have been proposed to explain late- to post-orogenic extension associated with the formation of core complexes.

1. Change in plate motions. This seems the most plausible explanation for the D’Entrecasteaux Islands core complexes, where UHP metamorphism was caused by subduction of the Australian margin beneath Papua New Guinea, but exhumation was temporally related to the westward propagation of the Woodlark Rift (Webb et al. 2008). The MCCs in the US Cordillera have been attributed to the change from subduction to transform motion along the western North American margin (e.g. Glazner & Bartley 1984), although the timing does not support this as a general explanation (Sonder & Jones 1999). MCCs in the Aegean and the western Mediterranean have been attributed to the onset of rollback in adjacent subduction zones (Avidov et al. 1997; Brun & Sokoutis 2010).

2. Releasing bends in strike-slip systems. The Gurla Mandhata dome in the High Himalaya lies in a releasing bend near the eastern termination of the Karakoram Fault (Murphy & Copeland 2005). The Simpion fault in the Alps, which exhumes the western part of the Lepontine Dome, may be the expression of a releasing bend in the dextral Tonale fault (Jiménez-Munín et al. 2005). Some of the Eocene extensional faulting in the Canadian and northern US Cordillera has been attributed to right stepping of the Late Cretaceous–Early Tertiary dextral fault system along the interior margin of the Canadian Rocky Mountain thrust belt (Parrish et al. 1988). The Black Mts in Death Valley appear to be a core complex in process of formation, and are kinematically related to right-stepping dextral faults within the Eastern California Shear Zone (Serpa et al. 1988). In most of these cases, however, extension cannot be regarded as simply an accidental consequence of an irregularity in a strike-slip fault. The overall motion is transtensional, and the component of vertical thinning and horizontal extension requires a dynamic explanation.

3. Late or post-orogenic collapse. Elevated topography associated with thickened crust in orogenic belts produces an increased vertical load, and this can be accentuated by processes such as delamination or convective removal of lithospheric mantle (England & Houseman 1989). The increase in gravitational potential energy (GPE) may be sufficient to drive vertical thinning and horizontal extension even while plate convergence continues, or extension may be triggered by slowing or cessation of convergence (Sonder & Jones 1999). The core complexes of the North American Cordillera are localized along the axis of the Sevier–Rocky Mountain orogen (Coney & Harms 1984), and the MCCs in the US Cordillera may have been triggered by founding of the Laramide “flat slab” ( Humphreys 1995). Three major belts of Cretaceous core complexes in China and adjacent areas of north-eastern Asia developed within thickened crust created by earlier collisional orogeny (Darby et al. 2004; Wang et al. 2011).

4. Core complex formation in the Aegean and western Mediterranean is located in areas of continental subduction and accretion, followed by the removal of subcontinental lithosphere by some combination of delamination, slab detachment and slab retreat (Platt & Vissers 1989; Carminati et al. 1998; Faccenna et al. 2001; Jolivet et al. 2008).

5. Extension initiated or driven by diapirism. Analogue and numerical simulations of diapirism (e.g. Dixon 1975; Jackson & Talbot 1989), together with observations on natural structures such as salt domes (e.g. Muehlerberger & Clabaugh 1968), support the idea that there can be substantial vertical shortening and localized horizontal extension over the top of the rising diapir. In the absence of regional extension, this local extension has to be accommodated locally, in the form of contractual structures flanking the dome. Diapirism cannot explain the patterns of regional extension seen, for example, in the US Basin and Range province.

6. Many MCCs are located in orogenic hinterlands (which is commonly the location of earlier crustal thickening), and some lie behind a subduction-related magmatic arc. This “back-arc” tectonic setting therefore raises the possibility that they are related to the extension widely observed behind oceanic subduction zones (Lallemant et al. 2008). The mechanics of back-arc extension are still very poorly understood, however, even though the process has been successfully reproduced in analogue experiments on subduction (Becker et al. 1999). It is clearly related kinematically to retreat of the trench relative to the upper plate of the subduction zone, but this does not provide a dynamic explanation. Possibilities include the motion of the upper plate in a deep mantle reference frame (Uyeda 1982), tractions at the base of the upper plate induced by corner-flow circulation in the mantle wedge (Schellart et al. 2010), the high GPE of back-arc crust (Platt 2007), or some combination of these.

MCCs are clearly extensional in origin, but this does not account for their distinctive tectonic style. The formation of low-angle normal faults with sufficient displacement to exhume rocks from below the BDT requires a more specific explanation, and the common occurrence of HP, UHP, and high-T metamorphic rocks in core complexes requires exhumational processes beyond what results from displacement along the detachments. Explanations in terms of magmatism, high thermal gradients, or melting in the lower crust do not encompass the full range of conditions under which MCCs can form.

Continental MCCs are almost exclusively found within orogens, and they develop in crust that has been thickened by continental collision, intra-continental shortening, or subduction and accretion of low-density crustal material. Orogenic crust has low effective elastic thickness and low effective viscosity, as a result of some combination of inherently weak materials (e.g. metasediments rich in phyllosilicates), abundant faults and shear zones formed during tectonic contraction, continuing metamorphic reactions, elevated thermal gradient and partial melting. Orogenic crust also has high GPE because of its thickness, and it tends to undergo loss of mantle lithosphere, which increases both GPE and thermal gradients. These conditions favour the generation of normal-sense faults and shear zones with low dips, controlled by pre-existing gently dipping faults and fabrics. Low effective viscosity allows bulk vertical thinning of the crust, and favours exhumational processes such as return flow in subduction channels, diapirism and pillow structures. Low effective elastic thickness favours rolling-hinge behaviour of the brittle detachments, and hence large displacements.

Core complexes as windows into the rheology of the deep crust

As discussed above, MCCs expose rocks that have been exhumed from deep in the crust, and in some cases from as much as 90 km below the surface. During exhumation, rocks experience conditions of decreasing pressure and water activity, and in the later stages, a rapid decrease in temperature. At depth, decompression of hot rock may result in partial melting, but in general decreases in temperature and water fugacity in the ductile field cause rocks to become stronger. The stress required for deformation increases, and the degree of strain localization increases. This may seem counter-intuitive, as strain localization requires weakening, but it requires weakening relative to the strength of undeformed rock under the same conditions. Deformed rocks in ductile shear zones record this evolution in conditions, and increasing localization means that structures and fabrics formed during earlier stages can be preserved. The resulting distinctive sequence of structures and microstructures informs us of the mechanics of deformation as the rocks are exhumed up to and through the BDT (Gueydan et al. 2005; Behr & Platt 2011, 2013). This allows us to define a sequence
of deformational styles, rheologies and structures, as a function of depth. These are summarized below, from the surface down (note that in the present-day situation, these microstructures are spatially collapsed onto one another, and commonly overprint one another at various scales).

(1) Brittle deformation along the detachment fault. Fault gouge developed at shallow levels \((T < 200°C)\) may be rich in smectite and mixed-layer clay minerals (e.g. Carpenter et al. 2012; Schleicher et al. 2012), which smear out along shear planes \((R\) or \(Y\) planes), reducing friction coefficients to 0.1–0.2 (Collettini et al. 2009; Holdsworth et al. 2011). This allows slip on normal faults with dips as low as 7–14°. At deeper levels, but still above the BDT, clay minerals are no longer stable (Haines & van der Pluijm 2012), fault rocks are enriched in hydrothermal minerals including feldspars, quartz, micas and chlorite, and commonly have a breccia or microbreccia texture (Fig. 12a). Coefficients of friction are likely to be around 0.4 in such rocks, and hence fault dips are likely to be >30°. Fault weakening may occur as a result of the breakdown of feldspars to micas, resulting in the formation of through-going shear bands rich in phyllosilicates (Imber et al. 2001; Bos & Spiers 2002). Large dynamic stress drops may occur as a result of frictionally induced melting, producing pseudotachylite (Di Toro et al. 2006; Fig. 12a).

(2) At and just below the BDT, there is an association of ductile structures (very narrow ductile shear zones, folds in foliated or layered rocks) and discrete discontinuities (Fig. 12b). Ultramylonites and phyllonites form as a result of extreme grain-size reduction by dynamic recrystallization, combined with breakdown of feldspars and mafic minerals to micas and chlorite (Fig. 12c and d). Weakening and localization processes include development of interconnected phyllosilicate-rich layers (Holyoke & Tullis 2006), and switches to dislocation-accommodated grain-boundary sliding in quartz mylonites (Behrmann 1985; Behr & Platt 2013; Fig. 13a). Shear stress at this depth is in the range 60–100 MPa (Behr & Platt 2011, 2013). Stress of this magnitude on the shear zone requires it to have a dip of >30°.

(3) For several kilometres below the BDT, ductile deformation is strongly localized in one or more shear zones with a cumulative width of a few tens of metres to 200 m (Behr & Platt 2013), with moderate to high-stress microstructures (20–60 MPa) (Fig. 14a and b). Localization processes include recrystallization-assisted dislocation creep (Platt & Behr 2011) in quartz mylonites (Fig. 14b), and mixing of phases (quartz, feldspar and micas) to form fine-grained ultramylonites (Fig. 13b). These processes generate a relatively thin layer of mylonite and ultramylonite that is genetically and geometrically related to the detachment (e.g. Behr & Platt 2011, 2013), although it may be cut out during exhumation along the brittle section of the fault (Fig. 8b).

(4) At a depth corresponding to a temperature of about 500°C, deformation in quartz-rich rocks ceases to be localized, and there is
a subhorizontal transition into a mid-crustal layer of distributed deformation. We refer to this transition as the localized–distributed transition, or LDT (Cooper et al. 2010b). The transition itself is likely to be marked by a zone of relatively high strain, as it separates the region above it, in which there may be undeformed blocks tens of kilometres in horizontal dimension bounded by narrow ductile shear zones, from the region of distributed deformation below. In the northern Snake Range, Cooper et al. (2010b) showed that this region is occupied by both west- and east-directed shear zones, which interfere and overprint each other as they sole into the LDT. The detachment fault itself is also likely to sole into the LDT, transferring displacement onto it. Our observations suggest that this happens when the shear stress drops to around 10 MPa. This corresponds approximately to the transition from Regime 2 to Regime 3 microstructures in quartz (Hirth et al. 2001; Stipp et al. 2002; Fig. 14b and c), and hence may mark conditions at which surface-energy driven grain-boundary migration proceeds at the same rate as strain-energy driven grain-boundary migration (Platt & Behr 2011). This may also mark the transition from recrystallization-assisted dislocation creep to recovery-assisted dislocation creep, which would explain the lack of strain localization. A fossil LDT is preserved in the Whipple Mountains, in the form of the ‘mylonitic front’ identified by Davis (1988), which forms a sharp, mappable boundary between mylonitized and unmylonitized granitic gneiss (Fig. 8a). It may also be preserved in the Snake Range, where it marks the western limit of lineated mylonites (Miller et al. 1983; Lee et al. 1987), but it is much more diffuse, because it is developed in a mechanically heterogeneous sequence of metasediments.

Based on the distribution of Ar–Ar cooling ages in the Snake Range (Lee & Sutter 1991; Lee 1995), Cooper et al. (2010b) suggested that successive positions of the LDT migrated down through the rock pile, as a result of continued exhumation and cooling during the Palaeogene.

(5) Increasing temperature with depth below the LDT causes a decrease in flow stress, because of the temperature dependence of dislocation creep. As temperatures approach the solidus, stress may drop to the order of 1 MPa, and there may be a significant rate of flow driven by pressure differences related to density or topographic gradients (Wernicke & Getty 1997; Rey et al. 2011). The resulting deformation may have the same kinematics as the overlying detachment, but this is not necessarily the case, and is characterized by high-temperature microstructures (Fig. 14d). Partial melting results in a dramatic drop in strength, and is likely to trigger diapirism, piercement structures, or the segregation and intrusion of magma (Rushmer 1996; Rosenberg & Handy 2005; Vanderhaeghe 2009). In subduction-zone environments, where temperatures at depth may be low, deformation below the LDT may be dominated by pressure-solution creep (Stöckhert 2002; Behr & Platt 2013). Depending on the composition of the crust, this zone of distributed deformation may be bounded below by more mafic or feldspar-dominated lower crust, or by the Moho.

Conclusions

Metamorphic core complexes reflect a distinctive mode of extensional tectonics, resulting from large displacements on normal faults that end up with subhorizontal, commonly arched or domed, orientations. These faults are probably initiated with relatively steep dips (>50°) within the brittle upper crust, and sole downwards into subhorizontal shear zones at a level (the LDT) where ductile deformation becomes distributed rather than localized. Extension in the hanging-wall wedge results in a rapid decrease in dip with time of the main trajectory of the fault through the upper crust, to about 30°. This then allows large displacement of the footwall towards the surface, passing through two ‘rolling hinges’, one at or below the BDT, and the other near the surface. The rolling-hinge mechanism results from isostatic readjustment to unloading by the fault, mediated by elastic flexure of the upper crust. Motion on gently dipping sections of the fault (as low as 10°) at shallow depths is facilitated by low coefficients of friction on clay-rich fault gouge. Gently dipping hanging-wall sections of the fault may therefore come to lie on rotated, gently dipping footwall sections.

Displacement of 20 km or more on the detachment exhumes ductilely deformed middle crustal rocks in the footwall, creating the assemblage we refer to as an MCC. During displacement, footwall rocks move progressively from a middle crustal zone of distributed ductile deformation at relatively low shear stress (<10 MPa) into a zone of localized ductile shear, involving switches to grain-size-sensitive deformation mechanisms under progressively decreasing temperature and increasing shear stress. Near the BDT, rocks fluctuate between brittle and ductile deformation within narrow zones of very high strain-rate (c. 10^-11 s^-1) and high shear stress (c. 100 MPa). They then pass into a zone of cataclastic deformation, also at high shear stress, accompanied by extensive hydrothermal alteration, and by intermittent seismic slip producing narrow veins of fractional melt. Continued exhumation brings the footwall into the stability region of clay minerals, resulting in a marked reduction in shear stress. This sequence results in the distinctive zonation of fault rocks along many detachment faults.

The double rolling-hinge mechanism results in a strongly arched or domed geometry in the footwall, characteristic of MCCs. This geometry is in some cases accentuated by fluid upwelling of partially molten rock from the lower crust, driven in part by buoyancy forces, and in part by the head of pressure produced by the crustal
load on either side of the extending zone. This produces migmatitic gneiss domes in the central parts of some MCCs. Flow of partially molten rock does not, however, appear to be an essential factor in the formation of MCCs, as many of these structures occur in low- to medium-grade metamorphic terrains, and in some cases in blueschist- or eclogite-facies rocks.

Most, and possibly all, continental MCCs occur in regions that have experienced crustal thickening by convergent tectonics, associated with zones of subduction, continental collision, or Cordilleran-type back-arc contraction. The switch to extension and normal faulting may be facilitated by thermal relaxation in the thickened orogen, or by heating associated with removal of lithospheric mantle. In many cases, a key factor in triggered extension in the middle- to upper-crust is exhumation of material from within a subduction channel or a deep collisional crustal root by buoyant upwelling. As a result, many core complexes contain rocks that have been exhumed from depths of 30–90 km; much of that exhumation was achieved by ductile processes, and predated the formation of the detachment fault that defines the resulting MCC. This multistage exhumational process can substantially confuse discussion about the significance and mechanics of MCCs.

Buoyant upflow of deeply buried rocks in the pre-MCC stage of exhumation may be facilitated by decompressional melting, or by pressure-solution creep in wet metasediments deformed at low to moderate temperatures in a subduction-zone environment. The transition into a core-complex mode of deformation in the middle–upper crust may result from increases of surface elevation associated with the return flow of deeply buried crustal rock. Alternatively, slowing or cessation of plate convergence may trigger both the return flow and the transition to extensional tectonics.

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