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Shear zones between rock units with no relative movement

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A B S T R A C T

Shear zones are normally viewed as relatively narrow deformation zones that accommodate relative displacement between two "blocks" that have moved past each other in opposite directions. This study reports localized zones of shear between adjacent blocks that have not moved past each other. Such deformation zones, which we call wakes, form due to the movement of exotic blocks within a viscous medium (denser blocks sinking within a salt structure, (the paths) between separated boudins), melt in partially molten surroundings (melt movement during migmatisation), or solid blocks sinking through a partially molten magma body (stoping). From the fluid dynamics perspective these shear zones can be regarded as low Reynolds number deformation zones within the wake of a body moving through a viscous medium. While compact moving bodies (aspect ratio 1:1:1) generate axial symmetric (cone like) shear zones or wakes, elongated bodies (vertical plates or horizontal rod-like bodies) produce tabular shear zones or wakes. Unlike conventional shear zones across which shear indicators usually display consistent symmetries, shear indicators on either side of the shear zone or wake reported here show reverse kinematics. Thus profiles exhibit shear zones with opposed senses of movement across their center-lines or -planes.

We have used field observations and results from analytical and numerical models to suggest that examples of wakes are the transit paths that develop where denser blocks sink within salt structures, bodies of melt rise through migmatites, between boudins separated by progressive extension and (perhaps) where slabs of subducted oceanic lithosphere delaminate from the continental crust and sink into the asthenosphere. We also argue that such shear zones may be more common than they have been given credit for and may be responsible for some reverse kinematics reported in shear zones.

1. Introduction

A shear zone is usually defined as a relatively narrow deformation zone, planar or curvi-planar in geometry, composed of rocks more highly-strained/foliated than their surroundings. Even though these structures are relatively narrow, large-scale mylonitized shear zones a few kilometers wide have been observed (Hobbs et al., 1986). Conventionally, shear zones are accepted as forming between blocks that have moved in opposite directions relative to each other. In his classical review of the geometry of shear zones, Ramsay (1980) defined boundary conditions for the geometrically simplest shear zones; that the shear zone is parallel sided, and the displacement profiles along any cross sections of the zone are identical (i.e. the finite strain profiles and the orientations and characteristic geometric features of small scale structural features across profiles are also identical). However, he argued that even though these conditions are unrealistic since shear zones have finite length and their displacement profiles must change near their termination, "very many shear zones often approximate closely to these over quite large zone lengths". More 'orthodox' shear zones where shear is dominant were called S-bands by Cobbold (1977).

Shearing results in e.g. deflection of markers and pre-existing foliation, formation of porphyroclast systems and sigmoids, rotation of porphyroblasts and the development of shear bands (genesis of mineral fish and S-C fabrics) (Mason and Manley, 1957; Reed and Tryggvason, 1974; Berthé et al., 1979; Lister and Snoke, 1984; Passchier and Simpson, 1986; Van den Driessche and Brun, 1987; Passchier, 1998; Vernon et al., 2004; Passchier and Trouw,

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1998; Mukherjee, in press). These structures have a monoclinic symmetry in cross-sections parallel to the shear direction and are frequently used as kinematic indicators to deduce the sense of shear (Passchier, 1998). Ideally such kinematic indicators show a consistent sense of shear within a shear zone (Passchier, 1998; Passchier and Trouw, 1998). However, in many orogenic shear zones, e.g. the South Tibetan Detachment System in the Himalayas as reviewed by Yin (2006) and tectonic boundaries in the Scandinavian Caledonides (Bergman and Sjöström, 1997; Gee et al., 2008), kinematic indicators are locally contradictory, i.e. show opposing senses of shear. Such contradictions can be due to at least two phases of shearing (i.e. a retro-shear on a pro-sheared zone), two conjugate shear zones that become sub-parallel during progressive deformation and/or the existence/development of tectonic lenses (Carreras, 2001). Carreras (2001) described the complex shear zones in the Cap de Creus in Spain and identified shear zones with opposite senses of shear that come to lie in close parallelism. He suggested "a progressive non-coaxial deformation regime" being responsible for the development of the complex kinematic pattern.

In this study, we present field and model evidence for the existence of wakes, a kind of merged shear zones, which form between two blocks with no motion relative to each other and where reversed kinematic patterns develop during a single phase of deformation. Many of these zones are welds where rigid or ductile bodies of rocks have passed through the reference field and are no longer visible. Froitzheim et al. (2006) described similar structures for faults and called them "extraction faults". They explained that extraction faults form where a volume of rock is extracted between two faults with opposite sense of displacement allowing these two faults to merge together at the trailing edge of the extracted body. In the next section, we will describe different cases where kinematic reversal may occur and in the following sections give examples from both numerical and analytical models, and natural cases.

2. Reverse kinematics

There are many examples of kinematic reversals in shear zones which are inverted due to two phases of deformation. For example, in the Lewisian high-grade metamorphic complex of Gairloch, reversal of dextral shear sense to sinistral is attributed to renewed phases of deformation (Lei and Park, 1993). Cooper et al. (2010) suggested that the mylonite zone below the northern Snake Range decollement (Basin and Range) has locally experienced kinematic reversal inconsistent with movement along a single detachment fault. Two opposite senses of ductile shear have also been recorded from the South Tibetan Detachment System (STDS) in the Himalaya in terms of mineral fish, sigma structures and S-C fabrics by Argles and Edwards (2002), Mukherjee (2007, 2010a, 2010b), and Mukherjee and Koyi (2010a and b) and many others. Kinematic reversal in the Scandinavian Caledonides is the result of shift from compressional tectonics and top-to-the-east thrusting during collision of the continental plates Laurentita and Baltica, to extensional tectonics resulting in top-to-the-west backsliding of the orogenic wedge (Gee et al., 2008).

Kinematic complexity is probably most common in transpressional and transtensional shear zones, which include both a non-coaxial and a coaxial component (e.g. Fossen and Tikoff, 1998). However, there are many cases where kinematic markers show reverse polarity (i.e. mirror symmetry) along a shear-zone-like structure that develops during one phase of deformation. These structures form due to movement of an object through a host rock. A number of natural examples exist where an object that has moved through a viscous or granular medium has left a wake flanked by zones of ductile shear. Some of these are; (i) active diapirs of salt, mud or magma that have been downbuilt or risen through their host rock; (ii) tracks of organisms that have burrowed in sandy layers and dragged nearby soft bedding (Fig. 8V of Passchier, 2001); (iii) melt that cut across pre-existing foliations during migmatisation (Fig. 4); (iv) newly nucleated grains distorting the cleavage of the host grain in which they grow (Fig. 3a–b of Mukherjee and Koyi, 2009); (v) drag of pre-existing fabrics by separating boudins, and (vi) a 'cylindrical fault' by Hills (1953) where the litho-units are dragged in the same sense across the brittle fault.

Additional geologic examples where an object (body) moves through a viscous medium driven by gravity or a pressure gradient is where a block of country rock falls into a magma chamber (stoping; Daly, 1903) or the gravitational descent of denser blocks or sheets within a salt diapir (e.g. Koyi, 2000, 2001). During their descent, these denser blocks shear the viscous salt along their boundaries (Figs. 2 and 3). For example, a two-dimensional anhydrite block (sheet) with finite length shears the viscous salt along both its boundaries (Burchardt et al., 2011) resulting in formation of shear zones with opposed shear senses. These shear zones form as a result of viscous drag along the contacts between the anhydrite sheet and the surrounding host rock salt and keep propagating as the object moves. As the anhydrite sheet continues to sink within the viscous salt, these two shear zones merge and fuse behind it forming a wake. The wake is a shear zone (and/or secondary weld) that separated the two compartments of salt along which the anhydrite sheet moved. The wake thus consists of the merger of two initial shear zones formed on each side of the sinking sheet and is characterized by rotated foliations with the same sense of curvatures across the object reflecting the opposed senses of shear on each side of the fused surface (Fig. 1). Schmeling et al. (1988) studied the axisymmetric case of sinking or rising Stokes spheres. In their models, the cone-like wake of these bodies is characterized by strong plane strain with the axis of extension progressively tilting and lengthening into the direction of the body as one approaches the axis of movement.

Other examples of complex or even misleading kinematic patterns may form in relation to collapsing magma- or fluid-filled voids (Bons et al., 2008). Bons et al. (2008) studied how Newtonian viscous media on pure shear occupy void spaces inside them in deformation experiments, and drag the nearby markers. We do not compare these models with ours since (i) the wakes we report here do not develop under pure shear; and (ii) the deformed markers of Bons et al. (2008) are not uniformly dragged and curved at any single side of the void–viscous medium contact as the foliations/markers related to wakes do.

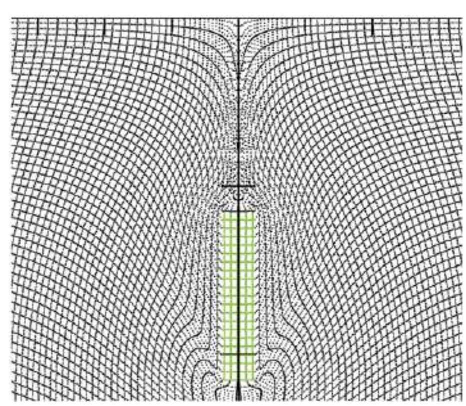

Fig. 1. Section of a two-dimensional numerical model showing a wake behind a sinking block (AR of 1:5). Note the downward drag of the horizontal markers behind the sinking block.

The understanding of crustal melting, melt extraction, migration, accumulation and flow as well as migmatisation and migmatite extrusion has increased significantly through the work by e.g. Brown and Solar (1998), Brown, 2010, Brown et al. (2011) and Sawyer et al. (2011). We argue that the recognition of wakes is important in the interpretation of field structures related to melt movements on various scales in the crust.

3. Numerical models

In order to quantify the finite strain pattern associated with the formation of wakes behind dense inclusions sinking through a linear viscous medium and study their structural effect two-dimensional Finite Differences models were prepared with the code FDCON (Weinberg and Schemling, 1988; Schmeling et al., 1999). Each model included a rectangular block with a high viscosity that sinks through the less-viscous matrix material driven by its relatively higher density (Figs. 1 and 2). Due to the two-dimensionality of the models, the block has infinite length in the third dimension. We varied the block aspect ratio (AR) from a square- with a width-to-thickness ratio of 1:1 to an elongate, rod-shaped geometry with an aspect ratio of 1:5. We also modeled the descent of blocks with ARs of 1:2 and 1:10. The block is initially placed at a depth of 1 block thickness in a rectangular box (12.5 block thicknesses wide and 50 block thicknesses deep).

The model results show that gravitational movement of a dense block induces high strain in the host material within a narrow zone in the wake of the block (Fig. 3a; cf. Burchardt et al., 2011, 2012). This wake is characterized by mirrored senses of shear on each side of the block (Fig. 1). Strain magnitudes in the wake reach a pronounced maximum that falls only slightly after a block descent of less than 1 block thickness (Fig. 2a and b). The high-strain wake is initially close to the width of the sinking object, but increases in width to about 5 times block width after only a short distance of descent to finally stabilize at about six block width. Even narrower blocks (0.5 block width) produce wakes with approximately similar width (see later). Doubling the thickness the block increases the maximum finite strain by only a factor of 1.3 (Fig. 2b and c).

4. Analytical models

How should the width of the wake be defined? We arbitrarily chose the distance from the axis of movement of the sunken body at which the finite strain measured as the ratio of the maximum to the minimum strain ellipsoid axis R exceeds a magnitude of 2. Taking this definition, the wakes of the sinking, non-deforming Stokes sphere of Schmeling et al. (1988) can be analyzed and result in the radius of the highly strained ($R > 2$) wake of 3.7 or 2.7 sphere's radii for no slip or free slip spheres, respectively. Thus low viscous bodies of constant shape (free slip boundaries) moving through low viscous media produce narrower wakes than solid bodies (no slip).

The width of the wake is also influenced by the shape of the sinking body. Fig. 3 shows the vertical velocity field in the ambient fluid of an axi-symmetric model as denser oblate and prolate ellipsoids sink through it along the vertical axis of symmetry. In both cases the long axis of the ellipsoid has been labeled 'a'. We define the half width of the shear zone near the moving body by the distance between the outer boundary of the body and the point at which the fluid velocity has dropped to 20% of the velocity of the body. In the Stokes flow solution of flow past a sphere the streamline passing through this point can be tracked and the finite strain can be integrated (Schmeling et al., 1988) resulting in approximately $R = 2$ in the wake of the sphere. The horizontal oblate (penny shaped) body produces a shear zone of half width of 1.9*a, while the vertical prolate (needle shaped) body generates a 0.9*a wide half width shear zone. Remember, "a" is the horizontal long axis in the first penny shaped, and the vertical long half axis in the second needle shaped case. Interestingly, this means that the 10 times narrower vertically falling prolate body (needle) leaves behind it a shear zone only about 50% thinner than the horizontal oblate body. We conclude that in both cases the half width of the shear zone or wake will be of the order of the longest dimension of the fallen body and not necessarily its cross section dimension.

While the width of conventional shear zones depends on the degree of non-linearity (or damage parameter) of the flow law of the shear zone, the widths of wake shear zones depend additionally on the characteristic dimensions of the object that has passed the

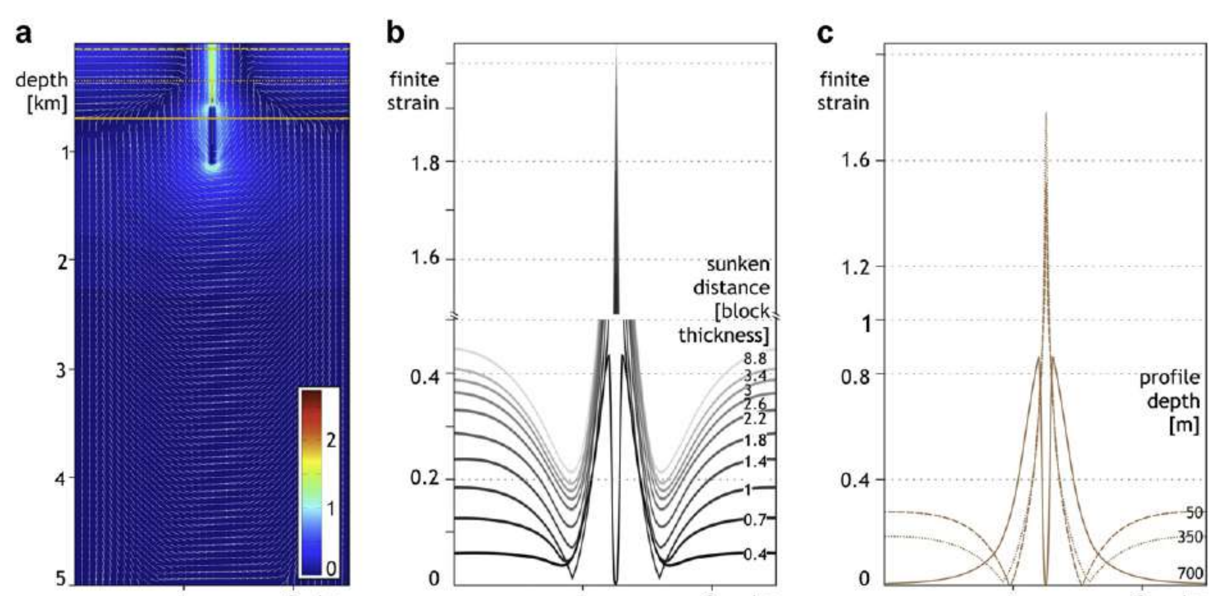

Fig. 2. Result of a numerical model with a block (AR of 1:10) sinking in a viscous medium. (a) Finite strain field (magnitude color-coded, orientation marked with white trajectories) after having sunk 1 block thickness. Orange lines display the location of strain profiles in b and c. (b) Finite strain along a horizontal profile at 350 m (initially through the middle of the block; dotted line in a) at different stages during block descent. (c) Finite strain along three horizontal profiles (locations indicated by orange lines in a) after the block has descended 1 block thickness. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

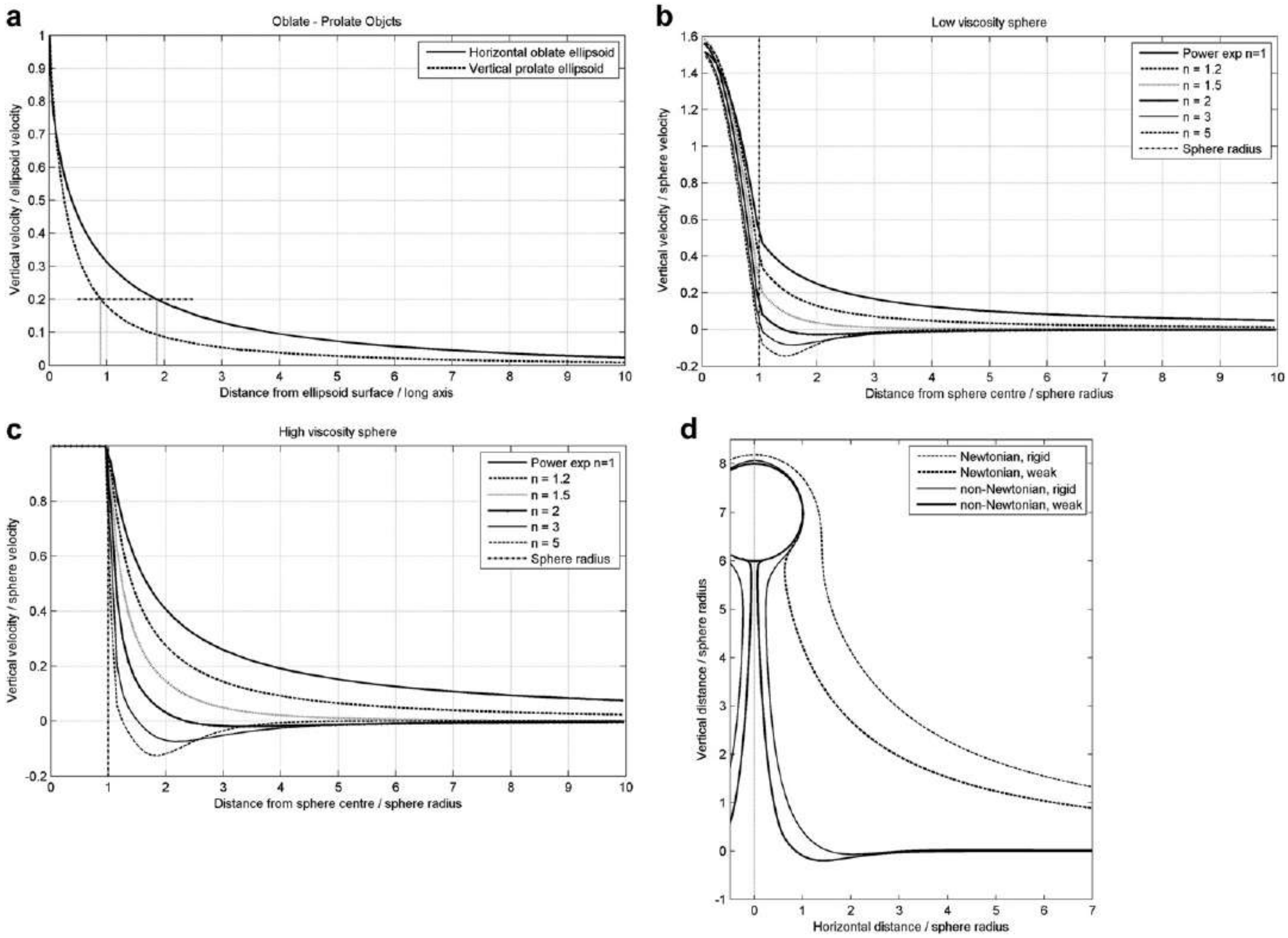

Fig. 3. Vertical velocity profiles normalized by the velocity of an object sinking or rising through a viscous medium as a function of the horizontal distance from the object. These velocity profiles can be regarded as being mapped as shear zones into the wake of the moving objects. (a) Horizontal penny shaped and vertical needle shaped rigid ellipsoids each having an axis ratio of 10 sinking in a Newtonian fluid. The distance is given starting at the surface of the object, and scaled by the long axis of the ellipsoids. The horizontal line indicates how we determine the width of the shear zone near the sinking objects. (b) A weak, low viscous sphere rising through Newtonian or non-Newtonian fluids with different power law exponent n. Here the distance is given starting at the center of the sphere and scaled by the sphere's radius. Note the negative side lobes showing the return flow near the sphere. Left of the vertical line indicating the sphere's radius the velocities represent the internal circulation within the sphere. (c) as (b) but for a rigid sphere rising through Newtonian or non-Newtonian fluids. d) Position of passive marker lines, having been penetrated by a rising sphere. The originally position of the lines were 5 sphere radii in front of the sphere, at the time shown the sphere has passed them by 7 radii. In the non-Newtonian cases the ambient medium has a power law exponent $n = 5$. The rigid (weak) spheres have Newtonian viscosities of $10^7\eta_0$ ($10^{-3}\eta_0$).

studied section. While in Newtonian flow, conventional narrow shear zones do not form under externally applied shear stresses, narrow wakes would develop behind sinking or rising objects even in Newtonian fluids. In non-Newtonian cases, shear zones beside and in the wake behind the objects strongly narrow and even may show additional reversals of the sense of shear in hosts that progressively strain soften. To show this we carried out a series of axisymmetric models with a power law fluid with an effective viscosity of

$$\eta_{\text{eff}} = A\ \dot{e}_{\text{II}}^{\frac{1}{n}-1}$$

with

$$A = \frac{1}{2}(2\eta_0)^{\frac{1}{n}}\left(\frac{1}{3}\Delta\rho g a\right)^{1-\frac{1}{n}}$$

where $\dot{e}_{\text{II}}$ is the second invariant of the strain rate, n is the power law exponent, η_0 is a reference viscosity, $\Delta\rho$ is the density contrast, g is the gravitational acceleration, and a is the sphere's radius. The definition of the scaling factor A ensures that the effective viscosity in the vicinity of the sphere will be of order η_0. Both the rise of a Newtonian low viscous sphere with a viscosity of $10^{-3}\eta_0$ and a highly viscous, almost rigid sphere with a viscosity of $10^7\eta_0$ have been considered. The shear zones beside the weak and rigid sphere are shown in Fig. 3b and c, respectively. Clearly, already a small increase of n from 1 to 1.2 results in a significant narrowing of the shear zone for the rigid sphere (Fig. 3c, point where the curves cross the 0.2 velocity level). For n-values 2 or higher, the shear zone width becomes $<< a$. In fact, for such n-values the return flow with negative velocities is concentrated and amplified close to the sphere. For low viscous spheres (Fig. 3b) the behavior is different, as the internal circulation accommodates a significant portion of the deformation. In Fig. 3a one has to distinguish between the circulation flow within the weak sphere, left of the line denoting the sphere radius, and the shear flow in the ambient region outside of the sphere. The shear zone outside the sphere becomes significantly narrower for increasing n compared to the rigid sphere. In fact, the weak sphere allows the return flow to concentrate and actually to reach the sphere's surface for n-values ≥ 3.

To test how these side shear zones are mapped into the wake of the objects for non-Newtonian versus Newtonian ambient media we place a line of passive markers initially horizontally at $z = 0$. We trace this line as a sphere with a starting position of 5 radii below that line approaches this line, passes and rises to a level of 7 radii above the initial line position. The positions of the resulting tracer line are shown for four different cases of a weak and rigid sphere with Newtonian and non-Newtonian ($n = 5$) ambient material in Fig. 3d, respectively. Clearly, the half widths of the non-Newtonian shear zones in the wake are of the order of or less than the sphere radius, thus significantly narrower than those of the Newtonian cases. On the other hand, they are much wider than the shear zones beside the moving non-Newtonian bodies (c.f. Fig. 3b and c). In fact, for highly non-Newtonian fluids the width of the wake scales with the cross section dimension, and probably no more with the longest axis of a vertically elongated rising body as was the case for a Newtonian ambient medium (see above). Interestingly, the change of sense of shear due to the return flow (Fig. 3b and c) is only weakly inherited in the wake (Fig. 3d): the absolute backward deflection of the passive marker line is only about $-0.2 \cdot a$ at a distance of about 1.5 sphere radius from the axis of rise. It has to be noted, that in contrast to a weak sphere in an infinite Newtonian medium a non-Newtonian ambient medium deforms the spherical shape of a rising weak sphere, thus the non-Newtonian, weak case shown in Fig. 3d may have to be slightly revised if this effect would be taken into account.

5. Discussion

The conventional view is that shear zones generally form between objects that have moved relative to each other in opposed directions. We suggest a mechanism for the initiation of a shear zone pattern between two rock units that have not moved past each other. The type of shear zone reported here record relative movement between a block and its hosting medium. This relative movement can be induced by gravity or pressure gradients (e.g. descent of denser blocks and ascent of lighter bodies such as melts).

The relative movement between a block and its host material needs not involve an object moving through a stationary host, or an object moving in a direction opposed to its host. Unlike conventional shear zones where opposite senses of movement of the boundaries is essential, blocks moving in the same direction as their hosts, but at different rates, also leave wakes. Such relative movement occurs, e.g. where a relatively low-density lower crustal material carries existing eclogitic bodies during crustal rebound (Koyi et al., 1999; Milnes and Koyi, 2000; Mukherjee and Mulchrone, in press), when rock fragments are transported at the base of an ignimbrite or when magma flow velocity is higher/lower than wall rock shear velocity (Correa-Gomes et al., 2001). One of the mechanisms suggested for exhumation of ultra high-pressure rocks is by a combination of entrainment of high-density eclogite blocks within low-density lower-crustal root masses during post-orogenic rebounds and thinning of the upper crust by extension (Koyi et al., 1999; Milnes and Koyi, 2000). During such root rebound, the entrained blocks are carried upwards by a mass that moves faster than the blocks it carries. This relative movement is likely to form shear zones along the boundaries between the block and its host materials. Having passed beyond the relatively slow moving block, these shear zones are expected to merge and fuse to form a wake with reverse kinematics that could be concealed below. Such shear zones may also form during lower crustal channel flow (Beaumont et al., 2001) or in the wake of tectonic lenses that have moved out of the reference frame in transpressive shear zones. After the extrusion of the Higher Himalaya by a channel flow during ~18 Ma (reviewed by Godin et al., 2006), a wake may be represented by shear zones with opposed senses of movement along the top and bottom of the mid-crustal channel. It is also likely that delaminated lithospheric slabs (Schott and Schmeling, 1998) leave wakes within the asthenosphere and lower mantle they descend through. We do not want to speculate on the details of any tectonic implications implied by such wakes within the mantle. However, wakes are likely to be strongly anisotropic zones which may influence later deformation.

Numerical models of anhydrite blocks sinking within Newtonian salt develop wakes (Figs. 1 and 2). Model results show that the intensity of shear strain increases symmetrically inward, but in opposed directions toward their mid lines or planes (Figs. 1 and 2). Such models emphasize that even small differential movements between the blocks and their host materials result in high finite shear (Fig. 2b). Fossen (2010) classified ductile shear zones into four types depending on the rheology of the material where those form. Natural ductile shear zones could strain harden, soften, or modify their thicknesses with time. The analog modeled wakes are not to be classified in such types since the rheology in our models are predefined (Newtonian) and it was maintained constant.

Identifying such shear zones is critical for deciphering the tectonic evolution of an area. Shear zones showing kinematic reversal are usually attributed to two phases of shear. However, our model results and field observations indicate that shear reversal may also develop during a single phase of deformation without the need for superimposed deformations or reversal of stress field. Our findings also emphasize that kinematic indicators within many shear zones showing different and/or opposed movement senses, may in reality indicate relative movements between different components of the shear zone due to different densities and/or viscosity ratios. In effect, we argue here that wakes represent complex shear zones between rock units that have not moved relative to each other.

The shear zones reported here may also form when less viscous bodies leave wakes as they migrate down differential pressure gradients. Melt escaping from migmatites can shear its surroundings (Druguet et al., 1997; Bons et al., 2004). Bons et al. (2004, 2008) described primary intrusive rocks from the Cap de Creus Peninsula of NE Spain as strings of beads and argued that such structures form in hot rocks intruded by dykes where magma solidifies slowly enough to allow enough ductile flow of the wall rock to accommodate the formation of the beads. We adapt Bons et al. (2004)'s explanation and use field evidence from the same area to argue that during their movement within the host rock, the individual beads formed wakes with reverse kinematics behind them (Fig. 4a and b). When melt pods accumulate enough mass and pressure to start migrating, they may suck the surrounding material radially into almost axisymmetric wakes in plan view (Fig. 4c and d). Any profile across such axisymmetric wakes will show kinematic reversal across their axes (Fig. 4d). Lack of concentric layering around the uniaxial wake in Fig. 4 is due to the fact that layering was subvertical when the melt was transported through them. These wakes can equally well develop during extension as more competent boudins drag a pre-existing fabric in the less competent host rock (Fig. 5a).

Explaining the mechanisms for the formation of tectonic lozenges bounded by shear zones in the Cap de Creus, Ponce et al. (2013) emphasized that the angular relationship between the pre-existing foliation and the bulk kinematic axes, and shear zone interaction rather than the bulk kinematics govern the deformation mechanism. Our observations from the same area (Fig. 4a and b) show that the angular relationship between the pre-existing fabric and the movement direction of the melt (or the object which moves through the host rock) dictates the nature of the wake symmetry. A moving object making a high-angle with a foliated host rock,

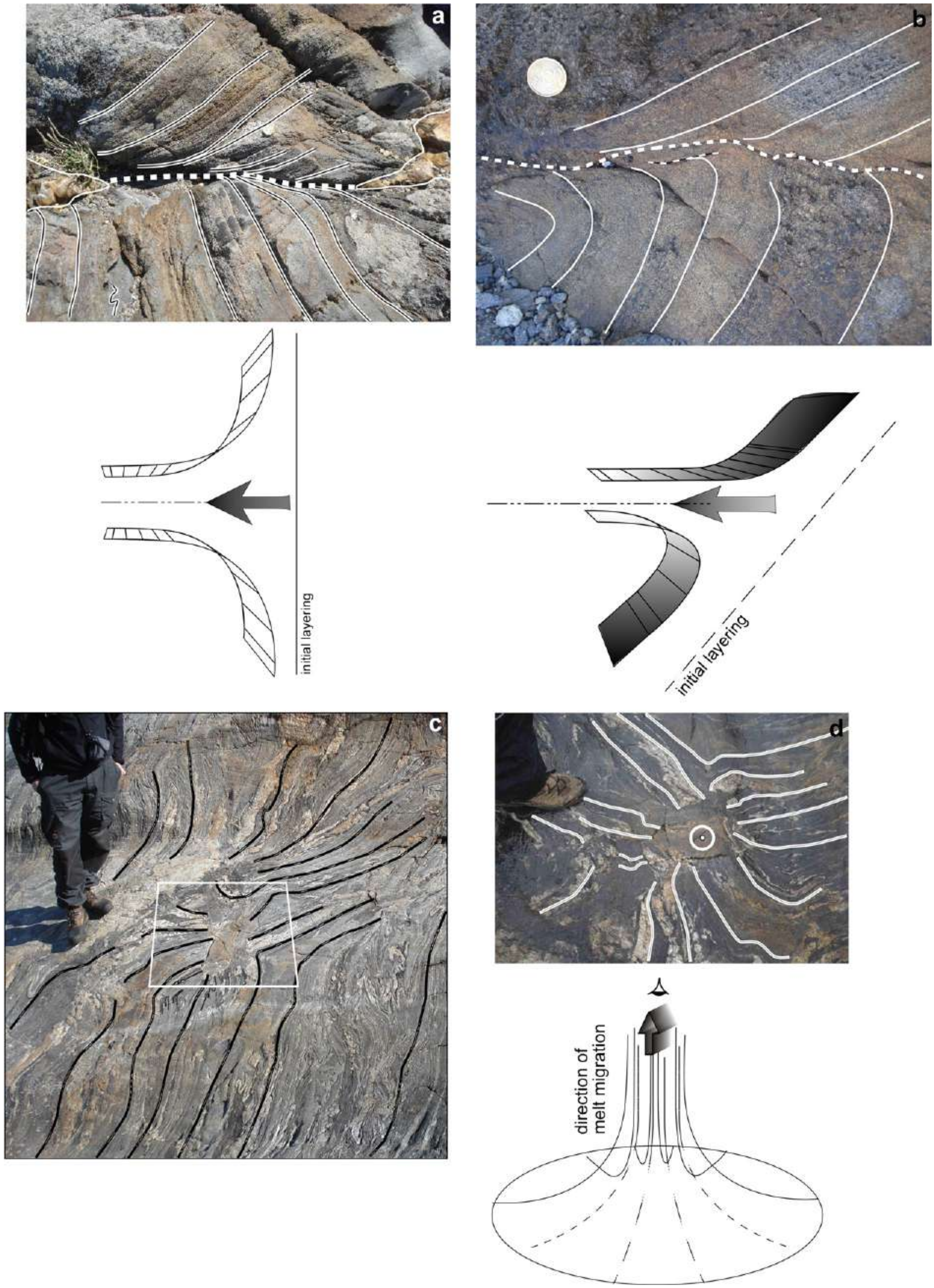

Fig. 4. Photograph of (a) a symmetric and (b) an asymmetric wake between beads (initially discontinuous veins (Druguet et al., 1997; Bons et al., 2004)), Cap de Creus, Spain. The continuous white lines show layering and dashed line outlines the wake. Note that the pre-existing fabric in the pelitic schist is bent in the same direction on either side of the symmetric and asymmetric wakes. Note also that layering which make a larger angle with the wake show a larger degree of rotation in b. The arrows in the schematic line drawings indicate the possible direction of melt movement. Photographs (c) overview and (d) close-up, of a radial pattern in migmatitic banding that was dragged into the wake of a small magmatic diapir that rose (either in or out) of this view on its way to higher structural levels. A mafic block was caught in the wake. The Island of Utö, Stockholm Archipilago, Sweden.

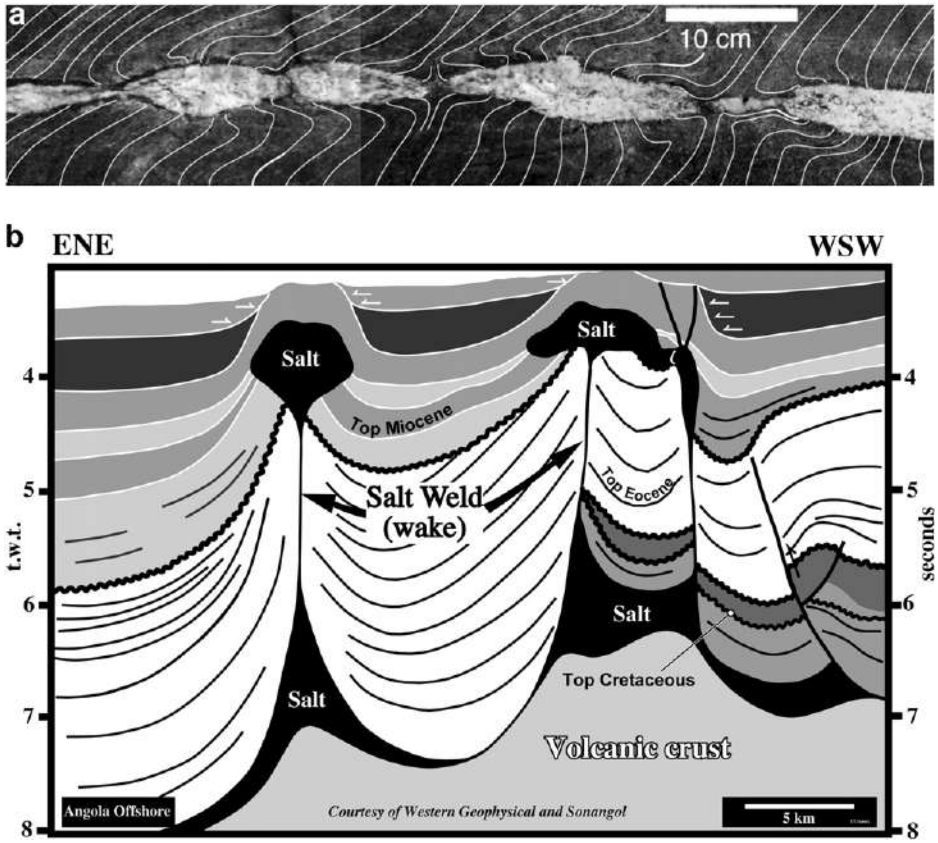

Fig. 5. (a) White lines trace foliation distorted into incipient welds (wake) between boudins of quartz in quartz porphyry, Utö, Sweden (Fig. 8J in Talbot, 2008). (b) Line drawing of a seismic section from offshore Angola showing secondary vertical welds behind salt diapirs whose stems are pinched off. Dr. Carlos Cramez kindly provided this figure, which is accessible at the link (http://homepage.ufp.pt/biblioteca/WebBasPrinTectonics/BasPrincTectonics/Page1.htm#Universidade.).

produces a symmetric wake (Fig. 4a), whereas a shallow angle between them forms an asymmetric wake (Fig. 4b). The asymmetry develops due to the higher rotation of the fabric on that side where the object makes an obtuse angle with the pre-existing fabric. In contrast, on the side where they are at an acute angle, the fabric rotates less to form an asymmetric wake (Fig. 4b). It is worth underlining here that objects crossing passive markers of a homogeneous host at a shallow angle are expected to result in an apparent asymmetric wake although the finite strain distribution on either *r* sides is expected to be symmetric. In contrast, objects moving at a shallow angle across active markers (foliation, bedding), which are associated with an anisotropic viscosity, are expected to result in complicated anisotropic and asymmetric viscosity distributions leading to asymmetric strain and wake. It is also correct to assume that melt migrating parallel to the pre-existing fabric in a Newtonian host is less likely to produce any wakes with mirrored polarity. The effect of orientation of pre-existing foliations on the localization and evolution of the shear zones in the Cap de Creus (Spain) was also emphasized by Carreras (2001). He argued that the orientation of the pre-existing foliation may dictate whether one or two sets of shear zones develop because the pre-existing foliation acts as mechanical anisotropy during shearing “leading to the development of instabilities, which commonly leads to the development of a self-similar pattern of shear zones in a wide range of sizes”.

A possible macro-scale example indicating a migmatitic wake exists along a steep, possible terrane boundary in the Paleoproterozoic rocks of central Sweden. To the south of the 25 km wide Gävle-Rättvik Shear Zone (GRZ), the large-scale structural and lithological trends in the Bergslagen province to the south rotate clockwise into the GRZ indicating dextral shear. In the Ockelbo Domain (North to the GRZ), exists a less distinct anticlockwise (sinistral) shear within the migmatites (Högdahl et al., 2009). Inconsistencies between kinematic indicators, stretching lineation and metamorphic variation north of the GRZ were interpreted to reflect sub-vertical extrusion of the migmatites. This attribution referred to strain partitioning and a perpendicular relationship amongst sub-horizontal fold axes, the steep mylonitic lineations, and the maximum stretching axis, *X*. This relation is based on a model for magma flow in migmatites of the Karakoram Shear Zone (Weinberg and Mark, 2008).

Salt diapirs rising through or downbuilt by their overburden is another example where a wake may develop. The opposite movement of salt relative to its overburden units result in drag along the flank of the diapir (the contact between the salt and its surrounding). Salt diapirs may drag its overburden units as it rises (Davison et al., 1996; Alsop et al., 2000). When the stem of the diapir is pinched (Koyi, 1998), these dragged overburden layers fuse behind the diapir to form a secondary weld (Jackson and Talbot, 1991; Giles and Lawton, 1999), which are generally steep. These secondary welds are in essence wakes showing mirror symmetry of the dragged overburden layers which converge upwards behind the diapir (Fig. 5b). If the salt diapir whose stem is pinched off is an axisymmetric structure, then the wake it leaves behind will also be axisymmetric similar to the case illustrated in Fig. 4d. However, a salt wall will leave a two-dimensional wake behind it once its stem is pinched off.

Marques and Cobbold (1995) modeled the formation of non-cylindrical folds nucleated around competent ellipsoidal inclusions in bulk simple shear regimes. They concluded that sheath folds form where competent inclusions retard flow and generate local three-dimensional differential flow during bulk simple shear. Natural sheath folds formed in a similar fashion were reported by Cobbold and Quinquis (1980). Our model results show that, even in the absence of layer-parallel bulk simple shear, blocks or bodies of melt moving across the layering can shear the layering/foliation into sheath folds. We are therefore proposing another mechanism for the formation of sheath folds in addition to the conventional mechanisms (Cobbold and Quinquis, 1980; Ghosh and Sengupta, 1984; Brun and Merle, 1988; Alsop et al., 2007, 2009).

Deformation in shear zones may involve volume change (Cobbold, 1977; Ramsay, 1980). The shear zones we describe here may involve volume change that depend on whether the object moving through the host is in-situ and domestic (i.e. melt from the host rock) or exotic and "just" passes through the host rock. When melt accumulates in-situ and eventually leaves its host, it contributes to a bulk volume loss leading to radial convergence toward the wake left behind the melt (Fig. 4c and d). However, melt transported within a Newtonian host is less likely to cause much volume change; the host experiences volume increase when the melt enters it (host inflation) and the host experiences volume loss (i.e. retains its initial volume) when the melt leaves it (host deflation).

6. Conclusions

Geological wakes form where (i) the flow between two materials is locally obstructed, slowed or impeded due to the movement of an object that moves at either a different rate, or in a different direction relative to its surroundings.

Geological wakes develop downstream of bodies that move faster (or "upstream" of bodies that move slower) than their ductile surroundings with different rheologies or densities.

Unlike conventional shear zones, geological wakes show reverse kinematics on either sides of their center-lines or -planes.

Geological wakes are symmetric where blocks or melts have moved through a uniform host material or when they move nearly orthogonal across a pre-existing fabric in the host rock. Geological wakes are asymmetric where blocks or melts have moved at a shallow angle across a pre-existing fabric.

The prerequisite of geologic wakes is that the body/material that caused the changes in relative flow is upstream or downstream of the wake but may no longer be within the available reference frame.

References

Alsop, G.I., Holdsworth, R.E., 2009. Discussion of sheath fold development with viscosity contrast: analogue experiments in bulk simple shear by Marques, F.O., Guerreiro, S.M., and Fernandes, A.R. Journal of Structural Geology 31, 215–217.

Alsop, G.I., Brown, J.P., Davison, I., Gibling, M.R., 2000. The geometry of drag zones adjacent to salt diapirs. Journal of the Geological Society of London 157, 1019–1030.

Alsop, G.I., Holdsworth, R.E., McCaffrey, K.J.W., 2007. Scale invariant sheath folds in salt, sediments and shear zones. Journal of Structural Geology 29, 1585–1604.

Argles, T.W., Edwards, M.A., 2002. First evidence for high-grade, Himalayan-age synconvergent extension recognized within the western syntax's – Nanga Parbat, Pakistan. Journal of structural Geology 24, 1327–1344.

Beaumont, C., Jamieson, R.A., Nguyen, M.H., Lee, B., 2001. Himalayan tectonics explained by extrusion of low-viscosity crustal channel flow coupled to focused surface denudation. Nature 414, 738–742.

Bergman, S., Sjöström, H., 1997. Accretion and lateral extension in an orogenic wedge: evidence from a segment of the Seve-Köli terrane boundary, central Scandinavian Caledonides. Journal of Structural Geology 19, 1073–1091.

Berthé, D., Choukroune, P., Jegouzo, P., 1979. Orthogneiss, mylonite and non coaxial deformation of granites: the example of the South Armorican Shear Zone. Journal of Structural Geology 1, 31–42.

Bons, P.D., Druguet, E., Hamann, I., Carreras, J., Passchier, C.W., 2004. Apparent boudinage in dykes. Journal of Structural Geology 26, 625–636.

Bons, P.D., Druguet, E., Castano, L.-M., Elburg, M.A., 2008. Finding what is now not there anymore: recognizing missing fluid and magma volumes. Geology 36, 851–854.

Brown, M., Solar, G., 1998. Granite ascent and emplacement during contractional deformation in convergent orogens. Journal of Structural Geology 20, 1365–1393.

Brown, M., Korhonen, F.J., Siddoway, C.S., 2011. Organizing melt flow through the crust. Elements 7, 261–266.

Brown, M., 2010. The spatial and temporal patterning of the deep crust and implications for the process of melt extraction. Philosophical Transactions of the Royal Society A 368, 11–51.

Brun, J.P., Merle, O., 1988. Experiments on folding in spreading-gliding nappes. Tectonophysics 145, 129–139.

Burchardt, S., Koyi, H.A., Schmeling, H., 2011. Strain pattern within and around denser blocks sinking within Newtonian salt. Journal of Structural Geology 33, 145–153.

Burchardt, S., Koyi, H.A., Schmeling, H., 2012. The influence of viscosity contrast on the strain pattern and magnitude within and around dense blocks sinking through Newtonian rock salt. Journal of Structural Geology 35, 102–116.

Carreras, J., 2001. Zooming on Northern Cap de Crues shear zones. Journal of Structural Geology 23, 1457–1486.

Cobbold, P.R., Quinquis, H., 1980. Development of sheath folds in shear regimes. Journal of Structural Geology 2, l19–126.

Cobbold, P.R., 1977. Description and origin of banded deformation structures. I. Regional strain, local perturbations, and deformation bands. Canadian Journal of Earth Sciences 14, 1721–1731.

Cooper, F.J., Platt, J.P., Platzman, E.S., Grove, M.J., Seward, G., 2010. Opposing shear senses in a subdetachment mylonite zone: implications for core complex mechanics. Tectonics 29, TC4019. http://dx.doi.org/10.1029/2009TC002632.

Correa-Gomes, L.C., Souza Filho, C.R., Martins, C.J.F.N., Oliveira, E.P., 2001. Development of symmetrical and asymmetrical fabrics in sheet like igneous bodies: the role of magma flow and wall-rock displacements in theoretical and natural cases. Journal of Structural Geology 23, 1415–1428.

Daly, R.A., 1903. The mechanics of igneous intrusion. American Journal of Science 15, 269–298.

Davison, I., Bosence, D., Alsop, I., Al-Aawah, M.H., 1996. Deformation and sedimentation around active miocene salt diapirs on the Tihama Plain, NW Yemen. In: Alsop, I., Blundell, D.J., Davison, I. (Eds.), Salt Tectonics. Geological Society, London, Special Publications, 100, pp. 23–40.

Druguet, E., Passchier, C.W., Carreras, J., Victor, P., den Brok, S., 1997. Analysis of a complex high-strain zone at Cap de Creus, Spain. Tectonophysics 280, 31–45.

Fossen, H., Tikoff, B., 1998. Extended models of transpression and transtension, and application to tectonic settings. In: Holdsworth, R.E., Strachan, R.A., Dewey, J.F. (Eds.), Continental Transpressional and Transtensional Tectonics. Geological Society, London, Special Publications, 135, pp. 15–33.

Fossen, H., 2010. Structural Geology. Cambridge University Press, pp. 306–307.

Froitzheim, N., Pleuger, J., Nagel, T.J., 2006. Extraction faults. Journal of Structural Geology 28, 1388–1395.

Gee, D.G., Fossen, H., Henriksen, N., Higgins, A.K., 2008. From the Early Paleozoic Platforms of Baltica and Laurentia to the Caledonide Orogen of Scandinavia and Greenland. Episodes 31, 44–51.

Ghosh, S.K., Sengupta, S., 1984. Successive development of plane noncylindrical folds in progressive deformation. Journal of Structural Geology 6, 703–709.

Giles, K.A., Lawton, T.F., 1999. Attributes and evolution of an exhumed salt weld, La Popa basin, northeastern Mexico. Geology 27, 323–326.

Godin, L., Grujic, D., Law, R.D., Searle, M.P., 2006. Channel flow, extrusion and extrusion in continental collision zones: an introduction. In: Law, R.D. (Ed.), Channel Flow, Extrusion and Extrusion in Continental Collision Zones. In: Searle, M.P. (Ed.),, Geological Society of London, Special Publication, 268, pp 1–23.

Hills, E.S., 1953. Outlines of Structural Geology, 182 pp.

Hobbs, B.E., Ord, A., Teyssier, C., 1986. Earthquakes in the ductile regime. Pure and Applied Geophysics 124, 310–336.

Högdahl, K., Sjöström, H., Bergman, S., 2009. Ductile shear zones related to crustal shortening and domain boundary evolution in the central Fennoscandian Shield. Tectonics 28, 1–18.

Jackson, M.P.A., Talbot, C.J., 1991. Glossary of Salt Tectonics, Bureau of Economic Geology, Geological Circular 91–4, 44 pp.

Koyi, H.A., Milnes, A.G., Schmeling, H., Talbot, C.J., Zeyen, H., 1999. Numerical models of ductile rebound of crustal roots beneath mountain belts. Geophysical Journal International 139, 556–562.

Koyi, H.A., 1998. The shaping of salt diapirs. Journal of Structural Geology 4, 321–338.

Koyi, H.A., 2000. Analogue modelling of entrainment of non-evaporitic rocks by salt diapirs. In: Geertman, R.M. (Ed.), Proceedings of the 8th World Salt Symposium, vol. 1, pp. 149–153.

Koyi, H.A., 2001. Modelling the influence of sinking anhydrite blocks on salt diapirs targeted for hazardous waste disposal. Geology 29 (5), 387–390.

Lei, S.H., Park, R.G., 1993. Reversals of movement sense in Lewisian brittle-ductile shear zones at Gairloch, NW-Scotland, in the context of Laxfordian kinematic history. Scottish Journal of Geology 29, 9–19.

Lister, G.S., Snoke, A.W., 1984. S-C Mylonites. Journal of Structural Geology 6, 617–638.

Marques, F.O., Cobbold, P., 1995. Development of highly non-cylindrical folds around rigid ellipsoidal inclusions in bulk simple shear regimes: natural

examples and experimental modeling. Journal of Structural Geology 17, 589–602.
Mason, S.G., Manley, R.St.J., 1957. Particle motion in sheared suspension: orientations and interactions of rigid rods. Proceedings of the Royal Society of London Series A 238, 117–131.
Milnes, A.G., Koyi, H.A., 2000. Ductile rebound of an orogenic root: case study and numerical model of gravity tectonics in the western Gneiss Complex, Caledonides, S. Norway. Terra Nova 12, 1–7.
Mukherjee, S., Koyi, H.A., 2009. Flanking microstructures. Geological Magazine 146, 517–526.
Mukherjee, S., Koyi, H.A., 2010a. Higher Himalayan Shear Zone, Zanskar Indian Himalaya: microstructural studies and extrusion mechanism by a combination of simple shear and channel flow. International Journal of Earth Sciences 99, 1083–1110. http://dx.doi.org/10.1007/s00531-009-0447-z.
Mukherjee, S., Koyi, H.A., 2010b. Higher Himalayan Shear Zone, Sutlej section: structural geology and extrusion mechanism by various combinations of simple shear, pure shear and channel flow in shifting modes. International Journal of Earth Sciences 99, 1267–1303.
Mukherjee, S., Mulchrone K. Estimating the viscosity and Prandtl number of the Tso Morari crystalline gneiss dome, Indian western Himalaya. International Journal of Earth Sciences, in press.
Mukherjee, S., 2007. Geodynamics, Deformation and Mathematical Analysis of Metamorphic Belts of the NW Himalaya. Unpublished Ph.D. thesis, Indian Institute of Technology, Roorkee, 267 pp.
Mukherjee, S., 2010a. Structures at meso and micro-scales in the Sutlej section of the higher Himalayan shear zone in Himalaya. e-Terra 7, 1–27.
Mukherjee, S., 2010b. Microstructures of the Zanskar shear zone. Earth Science India 3, 9–27.
Mukherjee, S., 2012. Simple shear is not so simple! Kinematics and shear senses in Newtonian viscous simple shear zones. Geological Magazine 149, 819–826.
Passchier, C.W., Simpson, C., 1986. Porphyroclast systems as kinematic indicators. Journal of Structural Geology 8, 831–843.
Passchier, C.W., Trouw, R.A.J., 1998. Microtectonics. Springer-Verlag, Berlin, 289 pp.
Passchier, C.W., 1998. Monoclinic model shear zones. Journal of Structural Geology 20, 1121–1137.
Passchier, C.W., 2001. Flanking structures. Journal of Structural Geology 23, 951–962.
Ponce, X., Drugeut, E., Carreras, J., 2013. Development of shear zone-related lozenges in foliated rocks. Journal of Structural Geology 50, 176–186.
Ramsay, J.G., 1980. Shear zone geometry: a review. Journal of Structural Geology 2, 83–99.
Reed, L.J., Tryggvason, E., 1974. Preferred orientations of rigid particles in a viscous matrix deformed by pure shear and simple shear. Tectonophysics 24, 85–98.
Sawyer, E.W., Cesare, B., Brown, M., 2011. When the continental crust melts: elements. Vil 2, 229–334.
Schmeling, H., Cruden, A.R., Marquart, G., 1988. Progressive deformation in and around a fluid sphere moving through a viscous medium: implications for diapiric ascent. Tectonophysics 149, 17–34.
Schmeling, H., Monz, R., Rubie, D.C., 1999. The influence of olivine metastability on the dynamics of subduction. Earth and Planetary Science Letters 165, 55–66.
Schott, B., Schmeling, H., 1998. Delamination and detachment of a lithospheric root. Tectonophysics 296, 225–247.
Talbot, C.J., 2008. Palaeoproterozoic crustal building in NE Utö, southern Svecofennies, Sweden. GFF 130, 49–70.
Van den Driessche, J., Brun, J.-P., 1987. Rolling structures at large shear strain. Journal of Structural Geology 9, 691–704.
Vernon, R.H., Johnson, S.E., Melis, E.A., 2004. Emplacement-related microstructures in the margin of a deformed pluton: the San Jose Tonalite, Baja California, Mexico. Journal of Structural Geology 26, 1867–1884.
Weinberg, R.F., Mark, M., 2008. Magma migration, folding and disaggregation of migmatites in the Karakoram shear zone, Ladakh, NW India. Geological Society of America Bulletin 120, 994–1009.
Weinberg, R.F., Schemling, H., 1988. Polydiapirs – multiwavelength gravity structures. Journal of Structural Geology 14, 425–436.
Yin, A., 2006. Cenozoic tectonic evolution of the Himalayan orogen as constrained by along-strike variation of structural geometry, extrusion history, and foreland sedimentation. Earth-Science Review 76, 1–131.