Fluid flow in compressive tectonic settings: Implications for midcrustal seismic reflectors and downward fluid migration

J. A. D. Connolly
Earth Sciences Department, Swiss Federal Institute of Technology, Zurich, Switzerland

Y. Y. Podladchikov
Physics of Geological Processes, University of Oslo, Oslo, Norway

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[1] Beneath the brittle-ductile transition of the Earth's crust, the dilational deformation necessary to create fluid pathways requires fluid pressure that is near to rock confining pressure. Although the deformation may be brittle, it is rate limited by the ductile compaction process necessary to maintain elevated fluid pressure; thus the direction of fluid expulsion is dictated by the mean stress gradient. The paradox posed by the conditions required to maintain high fluid pressure simultaneously with lower crustal rock strength can be explained by a model whereby fluids are localized within self-propagating hydraulic domains. Such domains would behave as weak inclusions imbedded within adjacent fluid-poor rocks. Because the mean stress gradient in a weak inclusion depends on its orientation with respect to far-field stress, the direction of fluid flow in such domains is sensitive to tectonic forcing. In compressional tectonic settings, this model implies that fluid flow may be directed downward to a depth of tectonically induced neutral buoyancy. In combination with dynamic propagation of the brittle-ductile transition, this phenomenon provides a mechanism by which upper crustal fluids may be swept into the lower crust. The depth of neutral buoyancy would also act as a barrier to upward fluid flow within vertically oriented structural features that are normally the most favorable means of accommodating fluid expulsion. Elementary analysis based on the seismogenic zone depth and experimental rheological constraints indicates that tectonically induced buoyancy would cause fluids to accumulate in an approximately kilometer thick horizon 2–4 km below the brittle-ductile transition, an explanation for anomalous midcrustal seismic reflectivity. *INDEX TERMS:* 3660 Mineralogy and Petrology: Metamorphic petrology; 5104 Physical Properties of Rocks: Fracture and flow; 5114 Physical Properties of Rocks: Permeability and porosity; 8045 Structural Geology: Role of fluids; 8102 Tectonophysics: Continental contractional orogenic belts; *KEYWORDS:* compaction, fluid flow, brittle-ductile transition, tectonic, stress, seismic reflector

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1. Introduction

[2] In compressive tectonic settings, relaxation of the yield stress developed in the brittle portion of the crust can result in a depth interval below the brittle-ductile transition characterized by an inverted pressure gradient [*Petrini and Podladchikov*, 2000; *Stuwe et al.*, 1993; *Stuwe and Sandiford*, 1994]. At depths at which this inverted pressure gradient is less than the hydrostatic gradient of an interstitial crustal fluid, if the fluid is subject to rock confining pressure, then it will migrate downward, ultimately stagnating at the depth at which the rock pressure gradient becomes identical to the fluids hydrostatic gradient. This condition defines a depth of tectonically induced neutral buoyancy that also acts as a barrier to upward fluid flow. In this paper we present an elementary model to constrain the scale of such phenomena and explore its relevance to hypotheses regarding a fluid-related origin for lower crustal seismic reflectors [e.g., *Klemperer*, 1987; *Hyndman*, 1988] and evidence for the involvement of meteoric fluids in the metamorphism of lower crustal rocks [e.g., *Wickham and Taylor*, 1987; *Upton et al.*, 1995].

[3] Numerous authors have observed that stress-induced deviations from lithostatic pressure can result in unusual fluid flow patterns such as downward fluxes and flow localization [e.g., *Ridley*, 1993; *Sibson*, 1996; *Ord and Oliver*, 1997; *Simakin and Petford*, 2003]. These workers have been concerned with small-scale perturbations associated with the development of structural features such as folds and shear zones or the influence of stress on perme-

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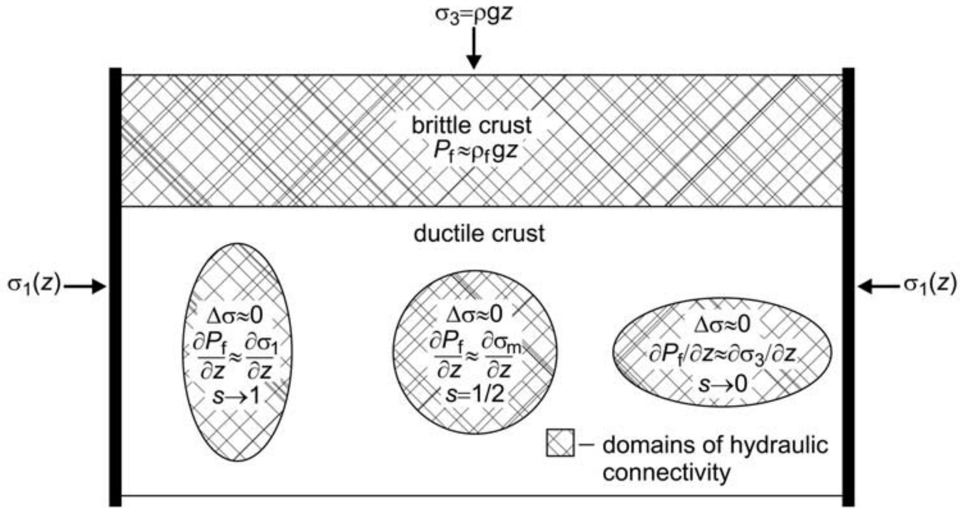

Figure 1. Conceptual model of the crust during horizontal compression in which fluid circulates freely within the brittle upper crust but is confined to isolated hydraulic domains within the ductile rocks of the lower crust. The domains might be manifest as a single structural element such as fracture zone, a magmatic diapir, or by fluids within an interconnected fracture mesh or grain-scale porosity. Nucleation and growth of self-propagating domains is a natural consequence of the inherent instability of pervasive fluid flow through a porous ductile matrix [e.g., *Scott and Stevenson*, 1984; *Richter and McKenzie*, 1984; *Connolly*, 1997; *Connolly and Podladchikov*, 1998]. Because of the presence of fluids the domains behave as weak inclusions, i.e., incapable of supporting significant differential stress, embedded within a strong matrix. Provided the stress fields about adjacent domains do not interact, then the pressure gradient within a vertically elongated domain is most strongly influenced by, and approximated here as, the vertical gradient in the far-field compressive stress, whereas if the domain is elongated in the horizontal direction, the pressure gradient within the domain is controlled by the vertical load. If the domain is spherical, the influence of both principal stresses is comparable and the far-field mean stress gradient is a suitable proxy for the pressure gradient within the inclusion.

ability [e.g., *Gavrilenko and Gueguen*, 1993; *Ge and Garven*, 1994]. In contrast, our concern is the large-scale effects of far-field tectonic stresses on the hydraulic potential responsible for fluid flow. There is, however, no general relationship between pore fluid pressure and the stress state (i.e., pressure) of the rock matrix. We therefore begin by reviewing the assumptions necessary to establish such a relationship, we then introduce the mechanism by which compressive tectonic stresses influence fluid flow, consider the scales on which it is likely to operate in nature, and conclude with a discussion of its geological implications.

2. Fluid Pressure, Mean Stress, and Hydraulic Domains

[4] Mechanical equilibrium of an interstitial pore fluid within a viscous rock matrix requires that the pressure of the interstitial fluid must be identical to the mean stress, i.e., pressure, supported by the fluid-rock aggregate. The apparent success of petrological thermobarometry premised on this equality [e.g., *Bucher and Frey*, 1994], and ubiquitous textural and structural evidence of deformation at high fluid pressures [e.g., *Etheridge et al.*, 1984] suggest that this limit must be approximately valid in the nominally ductile region of the crust. It is, however, widely appreciated that rocks that contain a fluid at, or near, the rock confining pressure have vanishing strength and deform brittlely. Thus the presence of high-pressure fluids throughout the lower crust is inconsistent with the paucity of deep crustal seismic events [e.g., *Sibson*, 1986; *Scholz*, 1988] and geomechanical models [e.g., *Brace and Kohlstedt*, 1980; *Kohlstedt et al.*, 1995; *Burov et al.*, 1998] that imply significant lower crustal strength. The paradox posed by crustal strength in the presence of high-pressure fluids can be explained if the fluids are localized within high-permeability domains. In such a domain the vertical fluid pressure gradient may approach the hydrostatic condition independent of the mean stress of the matrix [*Etheridge et al.*, 1984], but the mean fluid pressure must remain near the mean stress supported by the rock matrix with relatively small deviations determined by various hydraulic and/or rheological factors [*Connolly*, 1997; *Connolly and Podladchikov*, 1998]. Because fluid-saturated rocks have little strength [*Brace and Kohlstedt*, 1980], such domains would behave analogously to weak inclusions within a stressed solid [*Muskhelishvili*, 1963], irrespective of the nature of hydraulic connectivity within the domains. Thus the analogy applies equally to a domain, such as magmatic diapir or dike, that is entirely fluid filled, a network of fluid-filled fractures, or a domain in which fluid flow occurs through grain-scale porosity. In each case, a spherical domain would have the same mean stress gradient as the surrounding rocks, whereas the mean stress gradients in vertically and horizontally elongated domains would approach the vertical gradients in, respectively, the horizontal and vertical components of the far-field stress tensor (Figure 1). Validity of the Eshelby conjecture

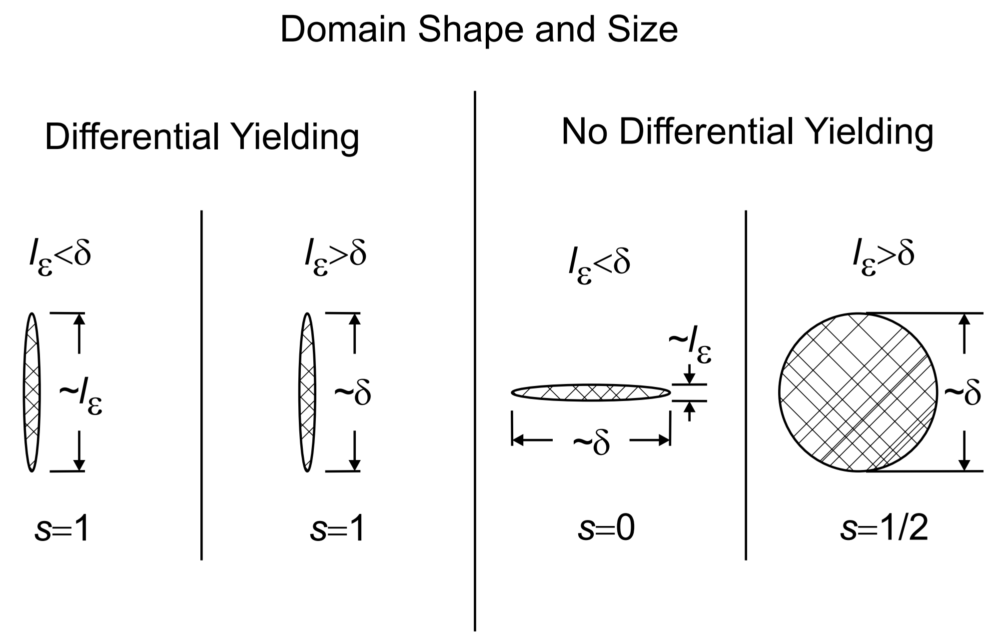

Figure 2. Dependence of the morphology and size of lower crustal hydraulic domains, in the absence of far-field tectonic stress, on deformation style. "Differential yielding" refers to behavior such that the rock matrix is drastically weakened at negative effective pressure (i.e., $P_{\text{fluid}} > P_{\text{total}}$) by processes such as hydrofracture. Within the nominally ductile lower crust, differential yielding would be increasingly favored toward the brittle-ductile transition. In view of this observation, and length scale estimates (Figure 3b), it appears probable that in the absence of tectonic stress, vertically elongated hydraulic domains predominate beneath the brittle-ductile transition. Elsewhere [*Connolly and Podladchikov*, 1998], we have shown that differential yielding increases the rate of domain propagation by up to 2 orders of magnitude. In the absence of yielding propagation, if $\delta > l_\varepsilon$, then velocities are on the order of Darcyian fluid velocity required to drain metamorphic fluid production by steady state pervasive flow ($\sim$10 km Myr^{-1} [*Connolly*, 1997]); however, if $\delta < l_\varepsilon$, then velocities (13) decay exponentially upward and may become insignificant within the ductile portion of the crust.

that the stress within a weak inclusion is constant, which is assumed here to imply a hydrostatic stress state for an inclusion in a gravitational field, is implicit in this logic. The Eshelby conjecture is of proven validity for ellipsoidal inclusions in linear elastic media [*Hardiman*, 1954]; here we also assume that any effects that might arise due to nonlinear viscous lower crustal rheology are negligible.

[5] The preceding argument neglects the relatively small influence of the deviation of the fluid pressure and mean stress gradients on the stress state of the rock matrix. This deviation is significant in that it induces deformation that not only influences the shape of the domain, but can also cause the entire domain to propagate in response to a gradient in mean stress, i.e., as a "porosity wave" [e.g., *Scott and Stevenson*, 1984; *Richter and McKenzie*, 1984]. It is well established that porosity waves would nucleate, and grow as the preferred mechanism of fluid expulsion, from uniform flow regimes in ductile rocks as a natural consequence of heterogeneities [*Scott and Stevenson*, 1986; *Wiggins and Spiegelman*, 1995; *Connolly and Podladchikov*, 1998] or perturbations caused by metamorphic reactions [*Connolly*, 1997]. Although the term porosity wave evokes the image of grain-scale porosity, the only restriction on the porosity wave mechanism is that the hydraulically conductive features occur on a spatial scale that is small in comparison to the scale of the domain. With this proviso, the concept applies equally, and perhaps with more relevance, to a domain defined by a network of interconnected network of fluid-filled fractures. A deformation-propagated mode of fluid flow is not essential to our arguments, but the effect of rheology on domain shape is important. *Connolly and Podladchikov* [1998] investigate this effect for compaction-driven flow regimes and show that the vertical scale for self-nucleating hydraulic domains is controlled by the shorter of two length scales: the viscous compaction length δ [*McKenzie*, 1984] and the scale for variation in the rheology due to the geothermal gradient l_ε. The horizontal scale is determined by whether or not the matrix strength is drastically reduced at negative effective pressure by processes such as hydrofracture, a phenomenon we designate as differential yielding. When differential yielding is suppressed, then the horizontal scale is the viscous compaction length, and the domains are spherical or oblate ellipsoids (Figure 2). The horizontal length scale under differential yielding is not easily constrained, but it must be less than the viscous compaction length, and consequently causes fluid flow to be channeled into vertically elongated structures (Figure 2) with spacing comparable to the viscous compaction length.

[6] To quantify the operative length scales in the nominally ductile portion of the crust, we assume a constitutive relationship of the general form

$$\dot{\varepsilon} = A \exp\left(\frac{-Q}{\mathrm{R}T}\right)\Delta\sigma^n, \quad (1)$$

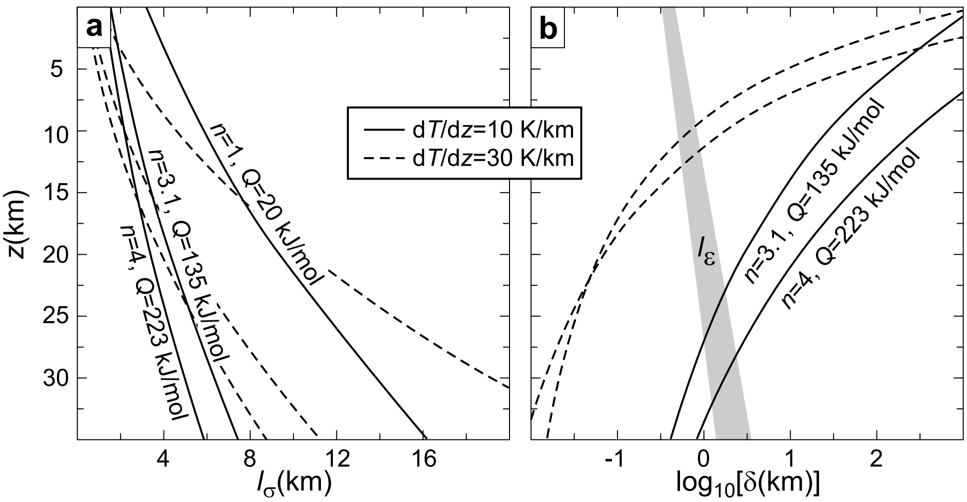

Figure 3. Characteristic length scales as a function of crustal depth for relaxation of (a) differential stress, l_σ (equation (2)) and (b) hydraulic domains maintained by ductile fluid expulsion, $l_\varepsilon = l_\sigma/n$ and δ (equation (6)). Solid and dashed curves correspond to geothermal gradients of 10 and 30 K km^{-1}, respectively. Curves are labeled by the relevant activation energy for viscous creep. For $Q = 223$ and $Q = 135$ kJ mol^{-1} the curves correspond, respectively, to the experimentally determined power law rheologies for quartz aggregates of *Gleason and Tullis* [1995] ($A = 1.1 \times 10^{-26}$ Pa^{-n}, $n = 4$) and *Paterson and Luan* [1990] ($A = 1.6 \times 10^{-24}$ Pa^{-n}, $n = 3.1$). The length scale for hydraulic domains is defined by the shorter of l_ε (shaded field in Figure 3b) or δ, the former being the length scale for depth-dependent variation in strain rate and the latter being the length scale for compaction in the absence of such variations [*Connolly and Podladchikov*, 1998]. In Figure 3b, δ is computed with $\phi = 0.01$and $q = 10^{-9}$ m s^{-1}, a plausible value for regional metamorphism [*Connolly*, 1997; *Manning and Ingebritsen*, 1999]. For $n \approx 3$, δ increases by a factor of 2 for an order of magnitude increase in ϕ or q. Although δ is uncertain, the estimates for δ in conjunction with the much narrower range for l_ε suggest hydraulic domains will develop on a length scale of ~1 km. Because of its strong depth dependence, δ is unlikely to dictate domain size over any significant depth interval within the crust unless it is orders of magnitude shorter than estimated, in which case fluid accumulation would occur on spatial scales that cannot be resolved by geophysical methods. The $Q = 20$ kJ mol^{-1} curves in Figure 3a correspond to pressure solution creep in sandstones as deduced from theoretical models [*Rutter*, 1976], natural compaction profiles [*Connolly and Podladchikov*, 2000], and experiment [*Niemeijer et al.*, 2002]. These values are likely to provide an upper limit on l_σ for the linear-viscous rheologies that might operate episodically during crustal metamorphism [e.g., *Rutter and Brodie*, 1995; *Kruse and Stunitz*, 1999].

where $\dot{\varepsilon}$ is the strain rate in response to a differential stress $\Delta\sigma$, T is the temperature, and A, Q, and n are material constants. To derive a length scale that accounts for the effect of temperature in equation (1), we characterize the depth dependence of the differential stress $\Delta\sigma$ that develops in response to horizontal compression of the crust by

$$\Delta\sigma = \left(\frac{\dot{\varepsilon}}{A'}\right)^{\frac{1}{n}} \exp\left(-\frac{z}{l_\sigma}\right), \quad (2)$$

where $\dot{\varepsilon}$ is the strain rate, the depth variable z increases downward, and l_σ is the characteristic length scale for variation of the differential stress, from which it follows the length scale l_ε for variation in the ductile rheology is l_σ/n. Rearranging equation (1) to obtain an explicit function for the differential stress and differentiating with respect to depth yields

$$l_\sigma = nT^2 \Bigg/ \left(\frac{Q}{\mathrm{R}}\frac{\mathrm{d}T}{\mathrm{d}z}\right), \quad (3)$$

and equality of equations (1) and (2) requires

$$A' = A\exp\left(-\frac{Q}{\mathrm{R}T_{\mathrm{ref}}} - \frac{nz_{\mathrm{ref}}}{l_\sigma}\right), \quad (4)$$

where T_{ref} is the temperature at reference depth z_{ref}. Substitution of appropriate parameters into equation (3) gives l_σ (Figure 3a) in the range 3–8 km ($l_\varepsilon \sim 1$–2 km) for power law creep in quartzite ($z = 15$–35 km), the lithology and mechanism commonly assumed to define crustal strength [*Kohlstedt et al.*, 1995]. Linearization of the steady state solution to the compaction equations (as formulated by *Connolly and Podladchikov* [1998]) generalized to power law rheology, defines the viscous compaction length as

$$\delta \approx \sqrt[n+1]{\frac{\eta k}{\mu\phi}\left(\frac{\phi}{\Delta\rho g}\right)^{1-n}}, \quad (5)$$

where $\eta \equiv \Delta\sigma^n/\dot{\varepsilon}$, k and ϕ are the matrix permeability and porosity, μ is the fluid viscosity, and $\Delta\rho$ is the difference between the density of the fluid and solid. Because permeability is a dynamic property of metamorphic systems [*Connolly*, 1997; *Connolly et al.*, 1997], it is convenient to reformulate equation (5) in terms of the magnitude of the metamorphic fluid flux $q \approx k/\mu\Delta\rho g$ necessary to maintain lithostatic fluid pressure in a hypothetical crust with uniform porosity as

$$\delta \approx \sqrt[n+1]{\frac{\eta q \phi^{n-2}}{(\Delta\rho g)^n}}. \tag{6}$$

The expulsion of metamorphic fluids through a uniform porosity places a lower limit on the compaction length relevant to the formation of hydraulic domains. Arguments based on heat conduction timescales [*Connolly*, 1997] and geochemical evidence [*Manning and Ingebritsen*, 1999] constrain regional metamorphic fluxes to be on the order of 10^{-9} m s^{-1}. Taking this value for q, with $\phi = 10^{-1}$, and pertinent rheological parameters, comparison of the estimates for l_ε and δ (Figure 3b), suggests that the vertical length scale for lower crustal hydraulic domains is on the order of a kilometer and that in the vicinity of the brittle-ductile transition domain propagation is limited by rheology rather than hydraulic properties such as permeability. This conclusion implies that equant domains within which the mean stress mimics the mean tectonic stress are improbable, but unfortunately does not eliminate either of extreme cases of sill-like domains and vertically elongated domains (Figure 2) that have, respectively no and the maximum possible sensitivity to compressional stress.

[7] It has been argued that during periods of active metamorphism ductile crustal deformation may be accomplished by linear-viscous mechanisms [e.g., *Rutter and Brodie*, 1995; *Kruse and Stunitz*, 1999]. While these mechanisms are not easily quantified, the low activation energies estimated for pressure solution creep in quartzites (Figure 3a) define an upper limit on l_σ of ~8–16 km that would allow a substantially wider range of behavior than inferred here for power law rheology.

[8] Although there is evidence for high fluid pressure and localized ductile deformation in the upper crust, in general upper crustal rock permeabilities are sufficient to permit fluids to circulate independently of the stress state of the host rocks [e.g., *Etheridge et al.*, 1984; *Zoback and Townend*, 2001]. For simplicity, we therefore assume hydrostatic fluid pressures maintain in the upper crust. As we are not concerned with fluid flow in this regime, this assumption is of importance only in the sense that it bears on upper crustal rock strength.

3. Stress Gradients During Tectonic Compression

[9] It is generally accepted that in compressional tectonic regimes the differential stress supported by the brittle upper crust increases with depth until depths are reached at which ductile mechanisms begin to operate, thereafter the differential stress decays as ductile deformation becomes increasingly efficient with depth (Figure 2a [after *Brace and Kohlstedt*, 1980]). Making the conventional assumptions that the minimum principal stress component is vertical and identical to the overburden weight; that the intermediate principle stress is identical to the mean stress; and that the variation in total density, ρ, with depth is insignificant; then the mechanical pressure within a hydraulic domain can be expressed [*Stuwe and Sandiford*, 1994; *Petrini and Podladchikov*, 2000]

$$P = \rho g z + s\Delta\sigma, \tag{7}$$

where the differential stress is $\Delta\sigma = \sigma_1 - \sigma_3$, and the factor s is introduced to differentiate the cases where pressure (Figures 1 and 2) is related to the mean far-field stress (s = 1/2), compressional far-field stress (s = 1), or the vertical load (s = 0). Equation (7) implies that pressure is in general supralithostatic. This pressure increases with depth in the brittle crust where rock strength is limited pressure-dependent yielding, whereas in the lower crust increasing efficacy of thermally activated ductile yielding causes the pressure to decay toward lithostatic values with increasing depth. *Petrini and Podladchikov* [2000] demonstrate that it is to be expected that decay of the differential stress in the upper portion of the ductile region occurs so rapidly that the rock pressure gradient will be inverted beneath the brittle-ductile transition. To assess the importance of this effect on fluid flow, we characterize the differential stress by equation (2), assuming that l_σ in the vicinity of the brittle-ductile transition can be taken as a constant characteristic of the lower crust. Although the computed variation in l_σ is seemingly large (Figure 3a), the effect of such variations is dampened by the logarithmic dependence of differential stress on l_σ in equation (2); thus the accuracy of our analysis is not reduced significantly by this approximation. With this simplification, equation (2) can be written in terms of the differential stress σ_Y at the depth of the brittle-ductile transition z_Y

$$\Delta\sigma = \sigma_Y \exp\left(\frac{z_Y - z}{l_\sigma}\right). \tag{8}$$

If the ductile deformation occurs by homogeneous pure shear (Figure 1), then from equations (7) and (8) the potential for vertical fluid flow ($\hat{P} = P - \rho_f g z$) is

$$\hat{P} = \Delta\rho g z + s\sigma_Y \exp\left(\frac{z_Y - z}{l_\sigma}\right), \tag{9}$$

where $\Delta\rho = \rho - \rho_f$. Because the first term in equation (9) grows linearly with depth, while the second term decays exponentially, there must be a depth z_0 at which the gradient in the hydraulic potential is zero, i.e.,

$$\frac{\partial \hat{P}}{\partial z} = 0 = \Delta\rho g + \frac{s\sigma_Y}{l_\sigma}\exp\left(\frac{z_Y - z_0}{l_\sigma}\right). \tag{10}$$

At this depth the buoyancy forces acting on the fluid are balanced by the stress gradient in the rock matrix and the driving force for vertical flow vanishes. Above this point, the hydraulic potential gradient is negative and fluid flow must have a downward component, whereas below the point fluid flow will have an upward component (Figure 4b). The existence of this point is only relevant if it lies beneath the brittle-ductile transition at z_Y. Rearranging equation (10), the depth interval $\Delta z = z_0 - z_Y$ over which

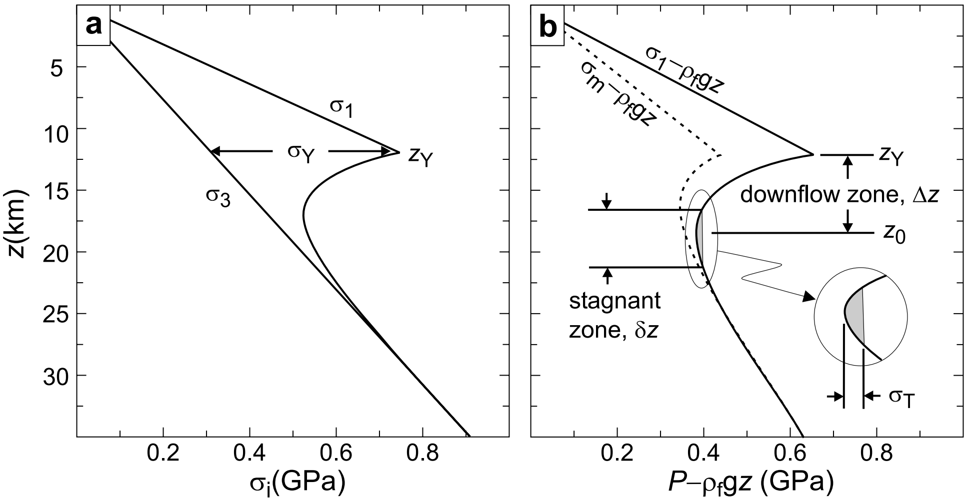

Figure 4. (a) Principal stress and (b) hydraulic potential depth profiles for a two-dimensional pure shear model of the crust. The vertical principal stress is the lithostatic load ($\rho = 2700$ kg m^{-3}). In the brittle crust, which extends to depth z_Y, the differential stress is computed from Byerlee's law for an internal angle of friction of $\pi/6$ and assuming the presence of hydrostatically pressured pore fluid with density $\rho_f = 800$ kg m^{-3}. The hydraulic potential for fluid flow at rock pressure is computed from equation (9) with $\Delta\rho = 1900$ kg m^{-3} and $l_\sigma = 3000$ m (Figure 3), values appropriate for aqueous fluids in a quartz-dominated rock matrix. The solid curve represents the potential within a vertically elongated domain ($s = 1$), whereas the dashed curve represents the potential within a spherical domain.

fluid flow would be driven downward in response to the mean stress gradient is

$$\Delta z = l_\sigma \ln\left(\frac{s\sigma_Y}{l_\sigma \Delta\rho g}\right), \qquad (11)$$

where negative values of Δz (i.e., $s\sigma_Y < l_\sigma\Delta\rho g$) indicate conditions for which there is no region of downward fluid flow. Positive values of Δz have the implication that fluids within the ductile region of the crust would tend to collect at depth z_0.

[10] Since the rheological parameter l_σ is well constrained (Figure 3a) and the variation in $\Delta\rho$ ($\sim$1900 kg m^{-3}) would be minor for aqueous fluids along the cool geotherms characteristic of compressive tectonic settings, the key unknown in equation (11) is the yield strength of the crust at the brittle-ductile transition (Figure 4a). The depth of the seismogenic zone (e.g., 10–20 km [*Sibson*, 1986; *Scholz*, 1988]) is commonly taken as evidence of a frictional sliding mechanism for brittle deformation in the crust such as described by Byerlee's law [*Brace and Kohlstedt*, 1980], a model supported by in situ stress studies [e.g., *Zoback*, 1992; *Zoback and Townend*, 2001]. Assuming that near hydrostatic fluid pressures maintain in the brittle crust, that the effective pressure is the difference between the rock and fluid pressure, and that the angle of internal friction is $\pi/6$, then the Mohr-Coulomb criterion gives the yield strength as a function of depth as

$$\sigma_Y = 2^m \Delta\rho g z, \qquad (12)$$

where the exponent m is introduced to distinguish the Mohr-Coulomb ($m = 1$) criterion from "Goetze's criterion" ($m = 0$) as discussed below. Employing quartzite power law creep rheology as a proxy for the ductile portion of the crust (l_σ = 2–5 km), equations (11) and (12) imply yield stresses at the base of the modern seismogenic zone would be adequate to cause an inverted gradient for fluid flow over a depth interval extending at 4–10 km beneath the brittle-ductile transition (Figure 5). This result is not greatly modified if the ductile crust deforms by pressure solution creep for which $l_\sigma \sim 8$ km is appropriate (Figures 3a and 5). In the case of a granitic magma the density difference between the magma and surrounding rocks would be more than an order of magnitude smaller than in the case of an aqueous fluid. Substituting such values into equation (11) increases the depth of the neutral buoyancy by el_σ ($\sim$10 km) over those estimated for aqueous fluid.

[11] *Kohlstedt et al.* [1995] suggest that the brittle-ductile transition is likely to be gradual, with ductile deformation becoming dominant at conditions such that the differential stress is comparable to the effective least principle stress, i.e., "Goetze's criterion" (equation (12)). This somewhat ad hoc criterion gives yield stresses half as large as from the Mohr-Coulomb criterion, but does not alter the conclusion that in compressional settings the brittle-ductile transition can be expected to act as a barrier to upward propagation of equant or vertically elongated hydraulic domains.

4. Stress-Induced Fluid Stagnation and Hydrofracture

[12] The foregoing analysis has the counterintuitive implication that the brittle-ductile transition is most likely to act as an obstacle to fluids within vertically oriented structural features that normally would be expected to be the most favorable means of accommodating fluid expulsion. In contrast, fluids concentrated in sill-like structures are not affected by compressive stress and therefore fluid

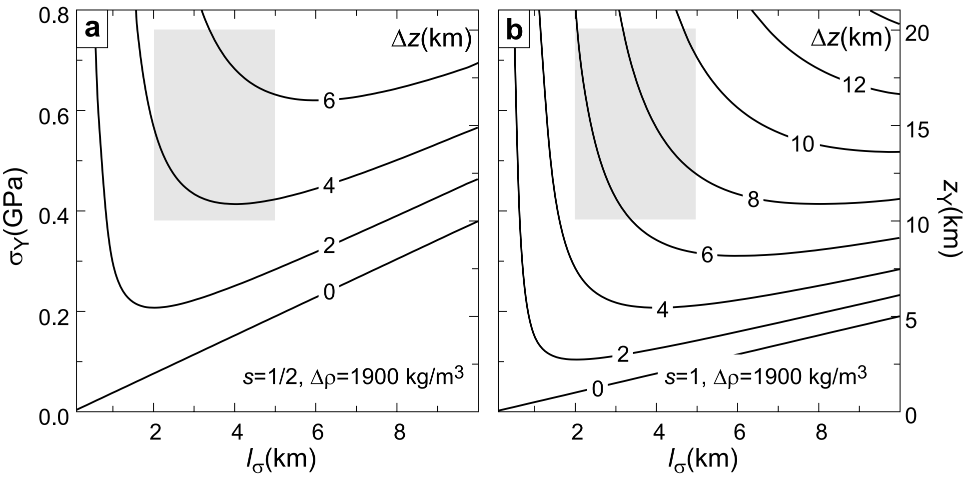

Figure 5. Vertical extent of downward directed fluid flow (equation (11)), for $\Delta\rho = 1900$ kg m^{-3}, beneath the brittle-ductile transition as a function of the length scale for ductile relaxation of the compressional stress l_σ (Figure 3) and the yield stress at the brittle-ductile transition in (a) vertically elongated ($s = 1$) and (b) spherical ($s = 1/2$) hydraulic domains. The right hand axes relates brittle yield strength to depth assuming the Mohr-Coulomb failure criterion as described for Figure 4 (equation (12)).

flow in such domains would have an upward compaction-driven component irrespective of tectonic forcings. These observations suggest an antagonistic relationship between the hydraulic potential and conductivity, such that conductive vertically oriented structural features would tend to evolve to less conductive horizontal structures beneath the brittle ductile transition. If it is assumed that the dynamic evolution of the domain is limited by the same deformation mechanism responsible for the ductile deformation of the crust, then for conditions such that $l_\varepsilon < \delta$, the local velocity scale [*Connolly and Podladchikov*, 2000] for the propagation of a horizontal domain is of the order

$$v \approx l_\varepsilon/\tau, \qquad (13)$$

where the local compaction time τ is [*Connolly*, 1997]

$$\tau \approx \exp\left(\frac{Q}{RT}\right) \Big/ A(l_\varepsilon \Delta\rho g)^n. \qquad (14)$$

For experimentally determined quartzite rheologies, the strong depth dependence and experimental uncertainty of these scales (Figure 6) precludes any broad statement about the efficiency of the ductile compaction mechanism in the depth range of interest. What can be stated is that for either rheological parameterization at shallow depths ($z \sim 10$ km), equation (13) implies a horizontal domain would propagate upward at <10 m Myr^{-1}. Thus in the absence of a more effective deformation mechanism, once formed, such domains would remain as long term features of the midcrust.

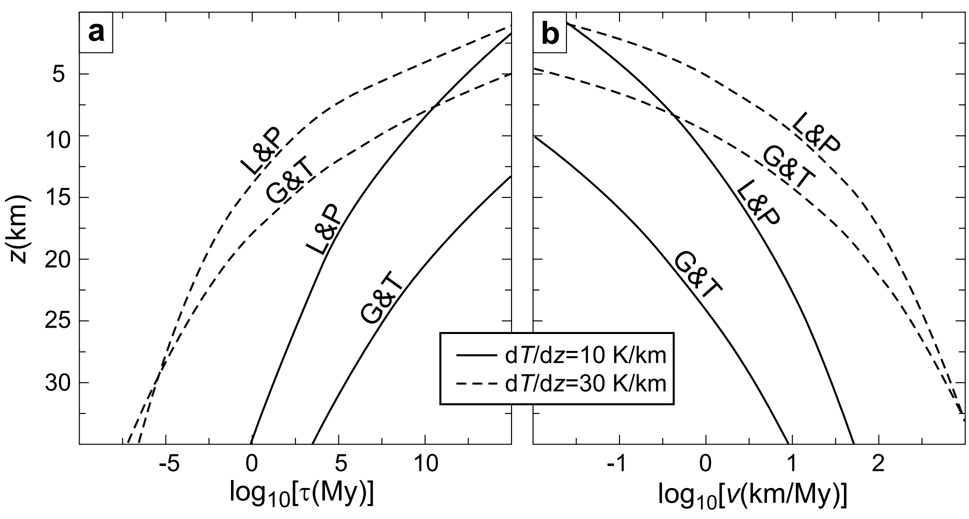

Figure 6. Local time (equation (14)) and velocity (equation (13)) scales for rheologically limited ($l_\varepsilon < \delta$, $v < q/\phi$) compaction. Curves labeled L&P and G&T correspond to the experimentally determined quartzite rheologies of *Paterson and Luan* [1990] and *Gleason and Tullis* [1995], respectively, with parameters as given in the Figure 3 caption.

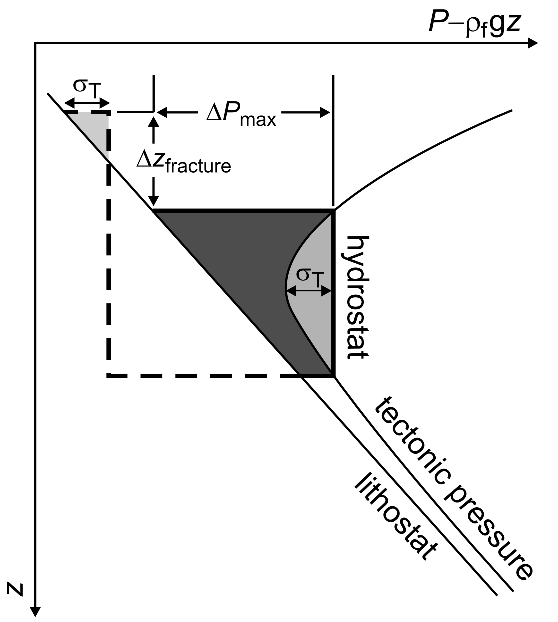

Figure 7. Schematic model of hydrofracturing induced by relaxation of compressional tectonic stress. During compression fluids accumulate within the stagnant zone with a maximum overpressure (lightly shaded field) limited by the tensile strength at the center of the stagnant zone. With relaxation of tectonic pressure toward the lithostat, the depth of maximum overpressure (heavily shaded field) shifts to the top of the stagnant region. The resulting overpressure causes hydrofracturing to propagate upward over the interval $\Delta z_{\mathrm{fracture}}$ until the overpressure decays to the tensile strength (profile indicated by the heavy dashed curve). Equation (17) gives an upper limit on $\Delta z_{\mathrm{fracture}}$ of ~5–7 km.

[13] Collection of fluids beneath the brittle-ductile transition must lead to the development of a depth interval (about z_0, Figure 4b) at which the fluid is overpressured but stagnant. If these fluids are accommodated by ductile dilational deformation, the depth interval would develop on the length scale $l_\varepsilon \sim 1$ km. In the event that fluid accumulation occurs on a timescale that is much shorter than the compaction timescale (Figure 6a), then the extent of this interval would only become limited once fluid overpressures became sufficient to induce hydrofracturing. Since the maximum overpressures would occur at depth z_0, fracturing would tend to localize within, rather than at the margins of, the overpressured interval. If the maximum sustainable fluid overpressure is equated to rock tensile strength σ_T and the curvature of the mean stress profile estimated from the derivative of equation (10) with respect to depth at z_0, then the interval over which fluids are trapped is

$$\delta z \approx 2^{3/2}\sqrt{\frac{\sigma_T l_\sigma}{\Delta\rho g}}. \tag{15}$$

For rock tensile strengths typical of those measured experimentally and inferred from structural studies (5–20 MPa [*Gueguen and Palciauskas*, 1994; *Simpson*, 1998]), equation (15) implies aqueous fluids would be trapped over a maximum depth interval of 2–5 km. The extent of this interval would also be constrained by the brittle-ductile transition, such that for conditions where $\Delta z = \delta z/2$ fluid the interval would breach the brittle-ductile transition and permit fluid to drain into the brittle crust. If this drainage perturbs fluid pressures within the upper crust, the consequent lowering of the yield strength at the brittle-ductile transition creates a feedback mechanism by weakening the stress-induced barrier to fluid flow, an effect that would cause episodic flow across the transition.

[14] The relaxation of compressional tectonic stresses would cause the locus of maximum fluid pressure to migrate upward from the middle toward the top of the stagnant domain providing a mechanism by which hydrofractures might be propagated upward (Figure 7). This process can be constrained given that the maximum fluid overpressure, relative to fully relaxed conditions, i.e., lithostatic pressure, in the stagnant region is

$$\Delta P_{\max} = l_\sigma \Delta\rho g \exp\left(\sqrt{\frac{\sigma_T}{l_\sigma \Delta\rho g}}\right). \tag{16}$$

If the effects of elastic dilation during hydrofracturing are disregarded, the vertical extent of the hydrofracturing necessary to cause fluid pressures to fall below the conditions for tensile failure is

$$\Delta z_{\mathrm{fracture}} = (\Delta P_{\max} - \sigma_T)/\Delta\rho g. \tag{17}$$

Taking the parameter values used previously, equation (17) implies this mechanism could propagate hydrofractures 5–7 km above the top of the initial zone of stagnation, providing a second mechanism by which deep crustal fluids might breach the brittle-ductile transition zone.

5. Downward Fluid Flow Through Ductile Rocks

[15] Inverted stress gradients create the potential for downward fluid flow in ductile rocks. For any particular yield stress at the brittle-ductile transition, the depth interval where these conditions are attained is fixed (equation (10)) and limits the extent of downward flow (Figure 5). However, in dynamic tectonic settings, increases in the intensity of crustal deformation cause the brittle-ductile transition to shift downward, giving rise to a mechanism by which upper crustal fluids might be swept into the lower crust. To quantify the scale of this effect, we note that a Taylor expansion of the expression derived by equating the ductile flow stress (equation (2)) to the brittle strength (equation (12)) shows that the depth of the brittle-ductile transition varies with strain rate as l_σ ln $(\dot{\varepsilon})$/n, provided $A'(2^m l_\sigma \Delta\rho g)^n < 1$, as generally the case. This strain rate dependence is identical to that obtained for the maximum depth of downward flow potential gradient

$$z_0 = \frac{l_\sigma}{n}\ln\left(\frac{\dot{\varepsilon}}{A'}\left(\frac{s}{l_\sigma \Delta\rho g}\right)^n\right). \tag{18}$$

Thus strain rate variations will not have a major effect on the width of the interval of downward fluid flow, but will cause the interval to shift together with the brittle-ductile transition. Given the parameters estimated for power law crustal rheology (Figure 3), this shift would be ~3 km downward per order of magnitude increase in strain rate, whereas for linear viscous rheology an order of magnitude increase in strain rate would be sufficient to cause aqueous fluids to flow to the base of the crust. The effectiveness of this mechanism is critically dependent on uncertain rheological properties of the crust and imposed strain rate. Applying the parameter values used previously (Figure 2) to equations (3) and (18) with strain rates of 10^{-15} s^{-1}, the depth of downward fluid flow varies from 9–13 km to 23–35 km with geothermal gradients of 10–30 K km^{-1}. Thus, quantitatively, it appears plausible that this mechanism could cause infiltration of upper crustal fluids into the lower crust, particularly in cold tectonic settings. The vertical flow channeling mechanism caused by differential yielding [*Connolly and Podladchikov*, 1998] is insensitive to the direction of fluid flow. Such a mechanism causes flow to focus spontaneously into vertical channels or to exploit weak preexisting structural features such as ductile shear zones. Therefore a virtue of this hypothesis is that it can explain the apparent association of fluids with ductile shear zones [*McCaig et al.*, 1990; *Upton et al.*, 1995; *Cartwright and Buick*, 1999; *Read and Cartwright*, 2000] without appealing to brittle dilatancy. In contrast to the formulation in the previous section that assumes knowledge of the depth dependence of the brittle strength to estimate the depth interval of downward fluid flow, no assumptions about the brittle rheology are involved in the prediction of the maximum depth of downward fluid flow from equation (18), which is valid provided the ductile rheology is dominant at the indicated depth.

6. Discussion

[16] The conflict between the existence of high metamorphic fluid pressures simultaneously with significant strength and ductile deformation style in the lower crust can be reconciled by the ‘‘layer cake’’ model proposed by *Etheridge et al.* [1984]. In the layer cake model the ductile crust is composed of alternating layers of strong, and relatively impermeable, fluid-poor, rock alternating with weak, permeable, fluid-rich rock. Within the weak layers the fluid pressure gradient is near hydrostatic, but the absolute pressure near the confining pressure. The ‘‘Swiss cheese’’ model advocated here is an extension of the layer cake model in which the fluid-rich domains are envisioned as weak self-propagating holes within ductile lower crustal rocks. This model accounts for the inherent flow instability caused by the divergence of the fluid and rock pressure gradients within the high-permeability domains [*Connolly*, 1997], an instability that explains both the nucleation and propagation of hydraulic domains within porous ductile rocks [e.g., *Scott and Stevenson*, 1984; *Richter and McKenzie*, 1984; *Connolly and Podladchikov*, 1998]. Analysis of the length scales for these domains (Figures 2 and 3) suggests that the vertical length scale is likely to be on the order of a kilometer and controlled by thermal activation of the ductile rheology. The horizontal length scale is dependent on the nature of the yield mechanism at high fluid pressure, such that differential yielding will favor the formation of high aspect ratio domains with a characteristic spacing comparable to the viscous compaction length (equation (5)). In the absence of yielding the viscous compaction length dictates the horizontal length scale, leading to the formation of sill-like domains. Because the direction of compaction driven fluid flow is dictated by the mean stress gradient [e.g., *Morgan*, 1987; *Spiegelman and McKenzie*, 1987; *Connolly and Podladchikov*, 2000], the orientation of hydraulic domains with respect to a tectonically imposed far-field stress field can have profound consequences. In extensional settings this effect may influence the magnitude of the hydraulic potential responsible for fluid expulsion but cannot affect its sign. However, in compressional settings the relaxation of brittle yield stress within the ductile portion of the crust leads to a depth interval characterized by a negative gradient in the compressive stress [*Petrini and Podladchikov*, 2000]. We have shown here that the negative stress gradient gives rise to a depth of neutral buoyancy for fluids within vertically elongated and equant hydraulic domains. Above the depth of neutral buoyancy such domains will propagate downward, whereas domains propagating upward from greater depth will become trapped at the depth of neutral buoyancy.

[17] Several lines of geological evidence lend credence to the mechanism proposed here for trapping fluids beneath the brittle-ductile transition. Geochemical evidence for lateral fluid flow within ductile rocks at midcrustal levels is common in metamorphic rocks that appear to record elevated fluid pressures [e.g., *Ferry*, 1994; *Witt et al.*, 1997]. Since the lateral pressure gradients that can be supported in ductile rocks are weak in comparison to the vertical hydraulic gradient under lithostatic conditions, large-scale lateral fluxes are difficult to reconcile with elementary mechanical principles unless the vertical hydraulic gradient is reduced as would be the case within zones of tectonically induced fluid stagnation. The simultaneous upward and downward propagation of hydrofractures at a depth of 12–14 km in the Kodiak accretionary prism [*Fisher et al.*, 1995] is explicable in terms of the accumulation of fluid at conditions of vanishing vertical hydraulic gradient. Alternating episodes of hydrofracturing and ductile folding in low-grade metamorphic settings [*Fisher et al.*, 1995; *Simpson*,1998] could be related to fluctuations in far field stresses or to breaching of the brittle-ductile transition, which would permit periodic drainage of the fracture systems as described in the previous section.

[18] Anomalous deep crustal seismic reflectivity in active orogenic settings that is attributed to accumulations of fluid offers intriguing, albeit indirect, evidence for stress-induced fluid stagnation [e.g., *Hammer and Clowes*, 1996; *Ozel et al.*, 1999; *Liotta and Ranalli*, 1999; *Makovsky and Klemperer*, 1999; *Vanyan and Gliko*, 1999; *Stern et al.*, 2001]. Although the midcrustal stress state in some examples is ambiguous [*Liotta and Ranalli*, 1999; *Makovsky and Klemperer*, 1999; *Vanyan and Gliko*, 1999], in all cases the reflectors are invariably associated with, but lie below, the seismogenic zone, an observation consistent with the analysis here, which suggests stress-induced fluid stagnation should occur over a narrow depth interval roughly centered about a point 2–4 km below the brittle-ductile transition (Figure 5). If fluid accumulation in this interval is accommodated by ductile dilation, its thickness should be on the

order of $l_\varepsilon \sim 1$ km, whereas accommodation by hydrofracture would lead to slightly greater thicknesses of 2–5 km (equation (15)). The vertical extent of midcrustal reflectors is often not well constrained, but the thicknesses in the range of a few hundred meters to 1.6 km as inferred for the Tibetan Plateau [*Makovsky and Klemperer*, 1999; *Li et al.*, 2003] and Mount Cayley bright spots [*Hammer and Clowes*, 1996] are comparable to those anticipated by our analysis. An oft cited objection to hypotheses that advocate a fluid-related origin for deep crustal seismic reflectors is that the high fluid connectivity and pressures required to explain associated electrical conductivity anomalies [e.g., *Wannamaker et al.*, 2002; *Li et al.*, 2003] is inconsistent with the low permeability required to prevent upward escape of the fluids [e.g., *Hyndman et al.*, 1993]. While various mechanisms of creating permeability barriers have been proposed [*Thompson and Connolly*, 1990; *Bailey*, 1994; *Connolly*, 1997], stress-induced stagnation is distinct in that it is a consequence of the absence of a potential gradient for fluid flow rather than modification of wetting properties or permeability. In this respect, stress-induced fluid stagnation would operate independently of the chemical and hydraulic properties of the rock-fluid systems as is consistent with the existence of reflectors that crosscut lithologic structures.

[19] Far-field compressional stress is expected to provoke a dynamic evolution in the morphology of lower crustal hydraulic domains from conductive vertically oriented structures to equant or tabular domains. This effect would amplify the tendency of thermally activated rheology, regardless of tectonic setting, to cause self-propagating hydraulic domains (i.e., porosity waves) to spread laterally (Figure 6b) and slow (Figure 7b) during upward propagation. In ductile rocks, these compaction effects cause domains to coalesce into quasi-static fluid-bearing horizons controlled by the rheology [*Connolly and Podladchikov*, 1998], a possible explanation for some lower crustal geophysical anomalies in extensional or posttectonic settings [e.g., *Marquis and Hyndman*, 1992; *Hyndman et al.*, 1993, *Vanyan et al.*, 2001]. While acknowledging that the majority of lower crustal geophysical anomalies occur in such settings, we have focused on compressive environments because of our interest in the influence of tectonic stress on lower crustal fluid flow. In this regard, it is pertinent to observe that elastic deformation in supposedly posttectonic continental shields generally maintains near critical compressive stresses to ~12 km depth as a consequence of plate boundary forces [*Zoback*, 1992; *Zoback and Townend*, 2001]. Viscous relaxation of these stresses is exactly analogous to the relaxation of brittle yield stress developed in active compressive settings and may therefore also induce lower crustal fluid stagnation. In drawing attention to these phenomena, it is not our intention to advocate a universal genetic relationship between fluids and lower crustal geophysical anomalies, but rather to suggest that such relationships are more palatable than sometimes supposed.

[20] Evidence for the involvement of meteoric fluids in metamorphism of ductile lower crustal rocks is widespread [e.g., *McCaig et al.*, 1990; *Wickham et al.*, 1993; *Upton et al.*, 1995; *Cartwright and Buick*, 1999; *Read and Cartwright*, 2000; *Gleeson et al.*, 2000; *Yardley et al.*, 2000, *Munz et al.*, 2002; *Gleeson et al.*, 2003] but contradicts the conventional wisdom that metamorphic fluid fluxes in ductile rocks are upward due to the compaction process ultimately responsible for fluid expulsion [e.g., *Walther and Orville*, 1982]. This contradiction is usually explained by a model after *Sibson* [1986] in which dilational deformation in large-scale shear zones results in lowered fluid pressures that cause upper crustal fluids to be drawn down into ductile lower crustal rocks. Progressive pressurization of inclusions that are inferred to represent originally shallow fluids supports this model [*Gleeson et al.*, 2000]. The weakness of this model is that the downward fluid fluxes that can be effected by this mechanism are localized and limited by the magnitude of the dilational deformation which must approach zero in the limit of a ductile shear zone. These limitations are at odds with the pervasive nature or deformation style in many of the fracture systems and shear zones through which downward fluid flow is thought to have occurred. The proposition proffered here that tectonic stresses may induce downward fluid flow does not suffer these limitations, but does require that the downward propagation of the brittle-ductile transition in response to increasing compressional stresses, which we argue is responsible for sweeping upper crustal fluids into the lower crust, must occur on a timescale comparable to that required for compaction to pressurize the fluids (Figure 7a). To the extent that power law quartzite rheologies are a valid analog for the ductile portion of the crust, what is less equivocal is that compressional tectonic strain rates are adequate to generate a zone of downward fluid flow that extends to the base of crust (equations (11) and (18)).

[21] Rigorous mechanical modeling of continental thickening has shown that the dynamic pressure can be represented by two principal components: flexural load and horizontal stress [*Petrini and Podladchikov*, 2000]. The flexural load acts to increase the dynamic pressure, but has little influence on the dynamic pressure gradient, which is largely controlled by the horizontal stress. The analysis presented here, based on a simplistic pure shear model for crustal deformation, is a first-order approximation that accounts only for horizontal stress. The pure shear model is therefore appropriate for establishing the depth intervals over which the potential gradient for fluid flow is upward or downward, but will tend to underestimate the magnitude of tectonically induced pressures.

[22] Conceptual models such as that posed here inevitably provoke criticism because they do not account for the heterogeneity of natural rocks. In anticipation of this criticism we observe that there are two limiting cases for lithological heterogeneity. If the heterogeneity occurs on a much smaller scale than the intrinsic length scales determined by the physical properties of an individual lithology, then the heterogeneities may create fine structure not accounted for in the homogeneous model, but will not alter the large-scale phenomenon. The problem in this context is not a limitation of the conceptual model but rather of defining effective physical properties for the composite lithology. Alternatively, if the heterogeneities are larger than the intrinsic length scales, then the relevant length scales must be adopted for each lithological unit. A likely effect of heterogeneity on the fine structure of our model would be to cause the zone of downward flow predicted for the homogeneous model to decompose into smaller zones, thereby

creating multiple depths at which tectonically induced fluid stagnation might occur.

Notation

A preexponential factor for $\Delta\sigma(T)$, equation (1) Pa^{-n}.
A′ preexponential factor for $\Delta\sigma(z)$, equations (2) and (4), Pa^{-n}.
k rock permeability, m^2.
l_ε vertical length scale l_σ/n for variation in rheology due to temperature (Figure 3b), m.
l_σ vertical length scale for stress relaxation during crustal compression (equation (3)), (Figure 3a), m.
m exponent distinguishing Mohr-Coulomb (m = 1) and Goetze's (m = 0) criteria (equation (12)).
n stress exponent (equation (1)).
P pressure (equation (7)), Pa.
$\hat{P}$ pieziometric pressure, $P - \rho_f g z$ (equation (9)), Pa.
Q activation energy for viscous creep (equation (1)), kJ mol^{-1}.
s hydraulic domain shape factor (Figure 1).
T temperature, K.
v hydraulic domain velocity (equation (13)), m s^{-1}.
z downward directed depth coordinate, m.
z_Y depth of brittle-ductile transition (Figure 4).
z_0 absolute depth of fluid stagnation (equation (18)) (Figure 4).
δ viscous compaction length (equation (5)) (Figure 3b).
δz maximum depth interval of fluid stagnation (equation (15)) (Figure 4b).
Δz relative depth of fluid stagnation, $z_0 - z_Y$ (equation (11)) (Figures 4 and 5).
$\dot{\varepsilon}$ pure shear strain rate, s^{-1}.
η = $\Delta\sigma^n/\dot{\varepsilon}$, Pa^n s.
ϕ porosity.
$\Delta\rho$ = $\rho - \rho_f$, ~1900 kg m^{-3}.
$\Delta\sigma$ differential stress, $\sigma_1 - \sigma_3$, Pa.
μ fluid viscosity, Pa s.
ρ bulk crustal density, ~2700 kg m^{-3}.
ρ_f fluid density, ~800 kg/m^3.
σ_Y differential yield stress in compression at the brittle-ductile transition.
σ_T yield stress in tension.
τ local compaction timescale (equation (14)), s.

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J. A. D. Connolly, Earth Sciences Department, Swiss Federal Institute of Technology, CH-8092 Zurich, Switzerland. (james.connolly@erdw.ethz.ch)

Y. Y. Podladchikov, Physics of Geological Processes, University of Oslo, Pb 1048, Blindern, 0316 Oslo, Norway.