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On mass transport in porosity waves

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ABSTRACT

Porosity waves arise naturally from the equations describing fluid migration in ductile rocks. Here, we show that higher-dimensional porosity waves can transport mass and therefore preserve geochemical signatures, at least partially. Fluid focusing into these high porosity waves leads to recirculation in their center. This recirculating fluid is separated from the background flow field by a circular dividing streamline and transported with the phase velocity of the porosity wave. Unlike models for one-dimensional chromatography in geological porous media, tracer transport in higher-dimensional porosity waves does not produce chromatographic separations between relatively incompatible elements due to the circular flow pattern. This may allow melt that originated from the partial melting of fertile heterogeneities or fluid produced during metamorphism to retain distinct geochemical signatures as they rise buoyantly towards the surface.

1. Introduction

Fluid migration in ductile rocks controls important geological processes such as melt segregation and fluid expulsion during regional metamorphism. Fluid production by partial melting and devolatilization leads to a percolating fluid network that allows for the segregation of fluid by porous flow at very low porosities (von Bargen and Waff, 1986; Cheadle, 1989; Wark and Watson, 1998; Miller et al., 2014; Ghanbarzadeh et al., 2014). Fluid segregation is driven by the buoyancy of the fluid and resisted by viscous compaction of the solid matrix (McKenzie, 1984; Scott and Stevenson, 1984; Fowler, 1985a). Fluid flow in rocks is predominantly vertical, because the segregation velocity of the fluid is significantly faster than the solid state creep velocity of the ductile rocks (Phipps Morgan, 1987; Sparks and Parmentier, 1991; Katz, 2008).

Fluid production in heterogeneous rocks leads to spatial variations in fluid content that may evolve into porosity waves, which migrate upwards at a velocity greater than the segregation velocity of the buoyantly rising background fluid. Porosity waves are an ubiquitous feature of the equations governing melt migration by porous flow (Spiegelman, 1993c). Porosity waves are also thought to arise from fluid expulsion during regional metamorphism (Bailey, 1990; Thompson and Connolly, 1990; Connolly, 1997, 2010; Tian and Ague, 2014; Skarbek and Rempel, 2016) and in the context of brine and hydrocarbon migration in sedimentary basins (McKenzie, 1987; Connolly and Podladchikov, 2000; Appold and Nunn, 2002; Joshi and Appold, 2016). In the aforementioned applications it is important to understand if solitary waves are effective carriers of energy, mass and geochemical signals. Here we revisit the viability of transport by porosity waves.

An idealized limit of compaction-driven porosity waves are so-called solitary porosity waves, which propagate at constant phase velocity, λ, without change in shape (Fig. 1*a*). In solitary waves the decompaction due to fluid overpressure at the front is perfectly balanced by compaction due to fluid underpressure in the back (McKenzie, 1984; Scott and Stevenson, 1984, 1986; Barcilon and Richter, 1986; Wiggins and Spiegelman, 1995; Simpson and Spiegelman, 2011). In one dimension, the fluid velocity within the solitary wave is increased relative to the background, but always remains lower than the phase speed of the solitary porosity wave (Fig. 1*b*). Therefore, no sustained mass transport occurs in one-dimensional solitary porosity waves (Richter and Daly, 1989; Barcilon and Lovera, 1989; Watson and Spiegelman, 1994; Spiegelman, 1994; Liang, 2008; Solano et al., 2014). This analysis of the one-dimensional case has led to the assumption that porosity waves in general cannot transport mass.

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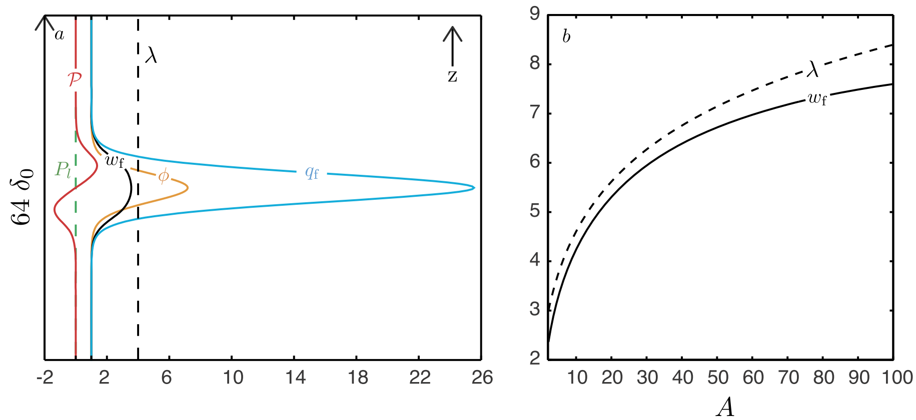

Fig. 1. One dimensional solitary porosity wave with phase speed, $\lambda = 4$. *a)* A high accuracy numerical solution for a dimensionless, one dimensional solitary porosity wave from Simpson and Spiegelman (2011): Porosity, ϕ, is scaled to the background porosity, $\phi_0 = 0.001$. Fluid pressure, $\mathcal{P}$ is scaled by the pressure due to buoyancy over the characteristic length scale, $\Delta\rho g \delta_0$. In the ambient background $\mathcal{P}$ is the lithostatic pressure, P_l. The upward volumetric flux of the fluid, q_{f}, and its vertical velocity w_{f} are scaled to the background separation flux, q_0. Both q_{f} and $w_{\mathrm{f}} = q_{\mathrm{f}}/\phi$ are elevated within the solitary porosity wave. *b)* Phase and vertical fluid velocities as functions of amplitude, A, of the porosity increase at the center of the solitary porosity wave. All calculations use the constitutive exponents $(n, m) = (2, 1)$, see Section 2.1 for definition.

In addition, fluid transport by porous flow in local chemical equilibrium leads to chromatographic separation of chemical elements according to their compatibility within the solid matrix (McKenzie, 1984; Navon and Stolper, 1987; Richter and Daly, 1989). A perfectly incompatible element travels at the velocity of the fluid, whereas the effective transport velocity of a trace element decreases relative to the fluid velocity with increasing compatibility. In the limit of perfect compatibility, the trace element travels with the solid. In one dimension, this chromatographic separation destroys any geochemical signature associated with the production of the fluid (Liang, 2008).

Fluid transport with porosity waves and chromatographic separations appear to make it impossible to preserve the distinct geochemical signature associated with the source region of the fluid. This is illustrated by the numerical simulation shown in Fig. 2. Here, fluid production has locally increased porosity and is initially co-located with two associated trace elements. Although the region of elevated porosity and trace element concentration are initially co-located, they become separated during fluid migration. As the trace element signatures abandoned by the porosity wave slowly migrate upwards, the continuous exchange between the fluid and solid separates tracers according to their compatibility. This implies that transport induced by the increase in fluid supply due to local fluid production carries with it no distinct geochemical signature.

However, the conclusion that solitary porosity waves do not transport mass is based upon one dimensional studies of melt transport. It is well known that one-dimensional porosity waves are unstable in two and three dimensions and break up into sets of cylindrical or spherical porosity waves (Scott and Stevenson, 1986; Wiggins and Spiegelman, 1995). Here we show that tracer transport in such higher dimensional porosity waves is dramatically different that in one dimension.

2. Fluid flow in two dimensional porosity waves

Models for fluid flow in ductile rocks assume a two phase mixture comprised of incompressible solid and melt phases. The flow of the fluid is described by Darcy's law and the solid matrix undergoes viscous deformation, often assumed to be Newtonian (McKenzie, 1984; Scott and Stevenson, 1984; Fowler, 1985a). Due to the intrinsic weakness of ductile rocks, porosities are very small. This allows significant simplifications to the governing equations that describe the two phase mixture. These simplified equations admit solutions in the form of solitary waves as shown in Figs. 1 and 2. The substantial literature on solitary wave solutions provides the ideal framework for discussing mass transport in porosity waves.

2.1. Governing equations in the small porosity limit

The dimensionless governing equations for the evolution of a porosity anomaly in a uniform background, in the limit of small porosities, are

$$\frac{\partial \phi}{\partial t} = \frac{\mathcal{P}}{\xi_\phi}, \tag{1a}$$

$$-\nabla \cdot K_\phi \nabla \mathcal{P} + \frac{\mathcal{P}}{\xi_\phi} = -\nabla \cdot K_\phi \hat{\mathbf{z}}, \tag{1b}$$

where $\mathcal{P}$ and ϕ are the dimensionless fluid pressure and porosity respectively and $\hat{\mathbf{z}}$ is the upward pointing unit vector. Here we write (1a) in terms of the partial derivative rather than the material derivative and assume no net translation of the solid. For the full dimensional governing equations see Appendix A.1.

The dimensionless permeability, K_ϕ, and effective viscosity, ξ_ϕ, are functions of porosity based on phenomenological laws,

$$K_\phi = \phi^n \quad \text{and} \quad \xi_\phi = \phi^{-m}, \tag{2a,b}$$

where the values of the exponents are typically $n \in [2, 3]$ and $m \in [0, 1]$ (Wark and Watson, 1998; Simpson and Spiegelman, 2011).

The porosity has been scaled to the characteristic porosity, ϕ_0, of the ambient background outside the porosity anomaly. The natural length scale that arises from the governing equations is the compaction length of the background, $\delta_0 = \sqrt{K_0 \xi_0/\mu}$, where K_0 and ξ_0 are permeability and effective viscosity of the background and μ is the fluid viscosity.

The fluid pressure, $\mathcal{P}$, is scaled by the pressure due to buoyancy over a compaction length, $\Delta\rho g \delta_0$, where $\Delta\rho = \rho_{\mathrm{s}} - \rho_{\mathrm{f}}$ is the density difference between solid and fluid, and g is the gravitational acceleration. The sign of $\mathcal{P}$ therefore indicates over and underpressure. Time is scaled by the segregation time δ_0/w_0, where the segregation velocity $w_0 = K_\phi \Delta\rho g/\phi_0 \mu$, is induced by the buoyancy of the fluid. The characteristic time scale is the time required for a percolating fluid to traverse a compaction length in the background.

The governing equations (1) admit solitary wave solutions in one, two and three dimensions. Fig. 3*a* shows porosity contours and the fluid pressure for a two-dimensional solitary porosity wave. Due to buoyancy, the fluid in the upper half of the solitary porosity wave is above lithostatic pressure and dilates the