ISSN No: 2348-4845
International Journal & Magazine of Engineering, Technology, Management and Research
A Peer Reviewed Open Access International Journal

Compressional sensing based fluid dynamics in weather forecasting

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Abstract
The standard Biot-Gassmann theory of poroelasticity fails to explain strong compressional wave velocity dispersion experimentally observed in 12 tight siltstones with clay-filled pores. In order to analyze and understand the results, we developed a new double-porosity model of clay squirt flow where wave-induced local fluid flow occurs between the micropores in clay aggregates and intergranular macropores. The model is validated based on the combined study of ultrasonic experiments on specimens at different saturation conditions and theoretical predictions. 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

I. Introduction
Turbulence in fluids has been an active research topic due to its impact on a wide variety of applications, including those in aeronautics, transportation, energy generation systems and weather forecasting. Several experimental studies of turbulent flows in various canonical and practical cases have enhanced the understanding of turbulent behavior, leveraging it to design more efficient systems. Although experiments have been invaluable, complementary computational fluid dynamics (CFD) efforts have the ability to gain a more detailed insight into the physics of turbulent flows, especially in situations where experimentation had been too expensive and/or impractical. With the availability of increased computing power in the recent years, high-fidelity CFD techniques like Large Eddy Simulation (LES) and Direct Numerical Simulation (DNS) have made it feasible to study turbulence with an unprecedented level of detail.

Cite this article as: Naresh Kagula, Devendra Uppara, Geetha Goparaju & Chakali Harish, "Compressional sensing based fluid dynamics in weather forecasting", International Journal & Magazine of Engineering, Technology, Management and Research, Volume 5 Issue 2, 2018, Page 233-243.

However, with an increase in fidelity there arise some significant challenges and opportunities. LES/DNS computations are very time intensive and expensive, therefore they are primarily used as a research tool to study the fundamental physics of turbulence. Additionally, these computations generate extremely high dimensional large datasets typically contain millions of degrees of freedom, which are often intractable to efficiently handle and analyze. As a result, techniques for modeling the high-fidelity dynamics of turbulence, while significantly minimizing the steep computation and data storage costs associated with LES/DNS have been a subject of active research [1]. Such efforts to model spatio-temporal dynamics of turbulence in a low dimensional space are generally referred to as Reduced Order Models (ROMs).

ROMs have two primary objectives: a) The ability to model the key dynamics/coherent features of the turbulent flow, and b) Provide an efficient means of data compression for LES/DNS datasets. A major application of such ROMs is to design flow control systems for turbulence, where their low computational cost and efficient models make them ideal candidates for building control logic for actuators [2]. Indeed, this capability of ROMs has been much sought after and has been a subject of considerable research, due to its wide-ranging applications in aerospace and mechanical engineering. Various methods of building ROMs exist, the common underlying theme being extracting the key features in the flow-field, preferably from a high fidelity experimental or computational data source. These extracted features are carefully chosen such that they represent dominant spatio-temporal dynamics computed by the Navier-Stokes equations. Mathematically, the goal here is model reduction i.e., representing high dimensional data in a low-dimensional subspace, which is essential to reduce computational and data handling costs [3]. The various types of ROMs have been demonstrated for canonical problems. The ROM based control for fluid flows. The luster based modeling to extract features and build adaptive ROMs [4].

ROM to build a Machine learning based control for various nonlinear dynamical systems. Generally, the most commonly adopted model reduction technique is the Proper Orthogonal Decomposition (POD) [5-7], as POD reduced basis (or modes) are mathematically optimal [8] for any given dataset. After model reduction, the next step is to use the reduced basis for modeling the flow at future time instants. A highly popular technique is the Galerkin projection (GP) approach, which has been documented extensively in literature [9–11]. The crux of the GP method lies in the use of spatio-temporal dynamics captured by the reduced basis (such as POD) which can then be evolved them in time, instead of the full Navier-Stokes equations. The use of the reduced basis with ODEs ensures that the computation is much cheaper, since they contain considerably fewer degrees of freedom. On the other hand, Navier-Stokes simulations utilize PDEs and considerably large mesh sizes which drastically increase computational costs. Therefore, GP based approaches have seen widespread popularity and have been demonstrated for several canonical problems [12].

However, GP models do not typically account for spatial variations in the flow and are known to become unstable under different conditions, even for canonical cases [13–16]. As such, considerable research efforts are being devoted to improving the stability of GP models, with some efforts focusing on Galerkin-free formulations. As a potential alternative, this paper explores a non-Galerkin projection-based approach to ROM using deep learning. Deep learning is a subset of machine learning which refers to the use of highly multilayered neural networks to understand (or "learn") a complex dataset with the intention of predicting some features of that dataset. The neural network does so by internally extracting a set of patterns or associations between different variables/parameters in the dataset, and subsequently using these patterns to predict a variable of interest (the output) given some input variable(s).

In recent years, deep learning has shown considerable promise in modeling complex datasets in diverse fields like imaging, finance, robotics etc. Application of machine learning (and deep learning) in fluid mechanics is an emerging area of research, with several efforts by Duraisamy Wang and Ling focusing on improving the accuracy of CFD simulations using data-driven approaches. Apart from CFD, reduced order modeling appears to be an ideal candidate to be approached using machine learning algorithms, due to the sheer data-driven nature of the problem.

The focus of this work is to exploit neural networks to "learn" the key dynamics of turbulent flows from high-fidelity simulation databases and use them to generate ROMs for flow control applications. These ROMs can then be employed to model the flow-field at future time instants. Different types of neural networks (NNs) exist for various types of data, and a choice must be made based on the application.

II. Fluid Pressure, Mean Stress, and Hydraulic Domains

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, and ubiquitous textural and structural evidence of deformation at high fluid pressures 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 and geo-mechanical models 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, 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. Because fluid-saturated rocks have little strength, such domains would behave analogously to weak inclusions within a stressed solid, 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.

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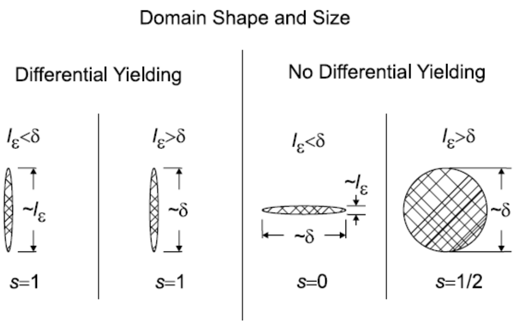

Figure 1. Dependence of the morphology and size of lower crustal hydraulic domains, in the absence of far-field tectonic stress, on deformation style. ‘‘Differential yielding’’ refers to behavior such that the rock matrix is drastically weakened at negative effective pressure (i.e., Pfluid > Ptotal) 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, it appears probable that in the absence of tectonic stress, vertically elongated hydraulic domains predominate beneath the brittle-ductile transition. Elsewhere 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 d > le, then velocities are on the order of Darcyian fluid velocity required to drain metamorphic fluid production by steady state pervasive flow (_10 km Myr_1); however, if d < le, 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; here we also assume that any effects that might arise due to nonlinear viscous lower crustal rheology are negligible. 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’’. 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 or perturbations caused by metamorphic reactions.

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. 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 d and the scale for variation in the rheology due to the geothermal gradient le. The horizontal scale is determined by whether or not the matrix strength is drastically reduced at negative effective pressure by processes such as hydro fracture, a phenomenon we designate as differential yielding. When differential yielding is

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suppressed, then the horizontal scale is the viscous compaction length, and the domains are spherical or oblate ellipsoids. 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 channelled into vertically elongated structures with spacing comparable to the viscous compaction length. To quantify the operative length scales in the nominally ductile portion of the crust, we assume a constitutive relationship of the general form

$$\dot{\varepsilon} = A \exp\left(\frac{-Q}{RT}\right)\Delta\sigma^{n}, \qquad (1)$$

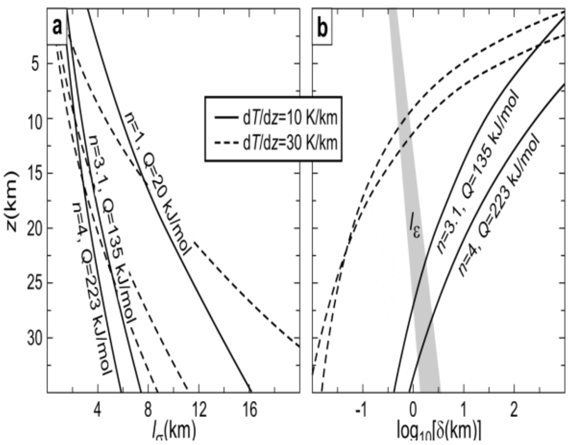

Figure 2. Characteristic length scales as a function of crustal depth for relaxation of (a) differential stress, l_s (equation (2)) and (b) hydraulic domains maintained by ductile fluid expulsion, $l_e = l_s/n$ and d (equation (6)). Solid and dashed curves correspond to geothermal gradients of 10 and 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 (A = 1.1 10^{26} Pa^n, n = 4).

The length scale for hydraulic domains is defined by the shorter of l_e (shaded field in Figure 3b) or d, 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. In Figure 3b, d is computed with f = 0.01and q = 10^9 m s^1, a plausible value for regional metamorphism. For n 3, d increases by a factor of 2 for an order of magnitude increase in f or q. Although d is uncertain, the estimates for d in conjunction with the much narrower range for l_e suggest hydraulic domains will develop on a length scale of 1 km.

Because of its strong depth dependence, d 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, natural compaction profiles, and experiment. These values are likely to provide an upper limit on l_s for the linear-viscous rheologies that might operate episodically during crustal metamorphism

III. Stress Gradients During Tectonic Compression

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. 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, r, with depth is insignificant;

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The 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. 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 brittleductile transition. To assess the importance of this effect on fluid flow, we characterize the differential stress by equation (3), assuming that l_s 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_s is seemingly large (Figure 3a), the effect of such variations is dampened by the logarithmic dependence of differential stress on l_s in equation (3); thus the accuracy of our analysis is not reduced significantly by this approximation. With this simplification, equation (3) can be written in terms of the differential stress s_Y at the depth of the brittle-ductile transition z_Y

$$\Delta\sigma = \sigma_Y \exp\left(\frac{z_Y - z}{l_\sigma}\right). \tag{2}$$

If the ductile deformation occurs by homogeneous pure shear (Figure 1), then from equations (2) the potential for vertical fluid flow (P^ = P $r_f$gz) is

$$\hat{P} = \Delta\rho g z + s\sigma_Y \exp\left(\frac{z_Y - z}{l_\sigma}\right), \tag{3}$$

where Dr = r _ rf. Because the first term in equation (2) grows linearly with depth, while the second term decays exponentially, there must be a depth z0 at which the gradient in the hydraulic potential is zero, i.e.,

$$\frac{\partial \hat{P}}{\partial z} = 0 = \Delta\rho g + \frac{s\sigma_Y}{l_\sigma} \exp\left(\frac{z_Y - z_0}{l_\sigma}\right). \tag{4}$$

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. The existence of this point is only relevant if it lies beneath the brittle-ductile transition at zY.

Since the rheological parameter l_s is well constrained (Figure 3a) and the variation in Dr (1900 kg m^3) would be minor for aqueous fluids along the cool geotherms characteristic of compressive tectonic settings, 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, a model supported by in situ stress studies. 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 p/6, then the Mohr-Coulomb criterion gives the yield strength as a function of depth as

s_Y ¼ 2^mDrgz;   ð12Þ (5)

where the exponent m is introduced to distinguish the MohrCoulomb (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_s = 2–5 km), imply yield stresses at the base of the

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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_s 8 km is appropriate. 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. Increases the depth of the neutral buoyancy by el_s (10 km) over those estimated for aqueous fluid.

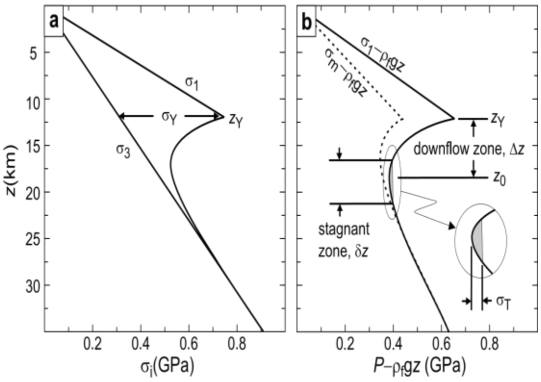

Figure 3. (a) Principal stress and (b) hydraulic potential depth profiles for a two-dimensional pure shear model of the crust. The vertical principal stress is the lithostatic load (r = 2700 kg m3). In the brittle crust, which extends to depth z_Y, the differential stress is computed from Byerlee's law for an internal angle of friction of p/6 and assuming the presence of hydrostatically pressured pore fluid with density r_f = 800 kg m3. The hydraulic potential for fluid flow at rock pressure is computed from equation (9) with Dr = 1900 kg m3 and l_s = 3000 m (Figure 3), values appropriate for aqueous fluids in a quartzdominated 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.

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''. 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.

IV. Stress-Induced Fluid Stagnation and Hydrofracture

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 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.

For experimentally determined quartzite rheologies, the strong depth dependence and experimental uncertainty of these scales 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 10 km), equation (3) implies a horizontal domain would propagate upward at <10 m Myr1. Thus, in

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the absence of a more effective deformation mechanism, once formed, such domains would remain as long-term features of the mid crust.

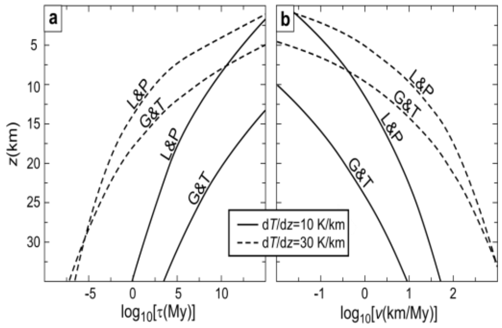

Figure 4. Local time (equation (14)) and velocity (equation (13)) scales for rheologically limited (l_e < d, v < q/f) compaction. Curves labeled L&P and G&T correspond to the experimentally determined quartzite rheologies of Paterson and Luan [1990] and Gleason and Tullis, respectively, with parameters as given in the Figure 3 caption.

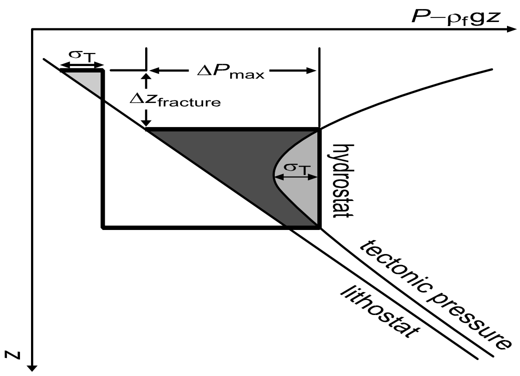

Figure 5. Schematic model of hydrofracturing induced by relaxation of compressional tectonic stress. During compression fluids accumulate within the stagnant zone with a maximum overpressure (lightly shaded field) limited by the tensile strength at the center of the stagnant zone.

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 $Dz_{fracture}$ until the overpressure decays to the tensile strength (profile indicated by the heavy dashed curve). Equation (3)gives an upper limit on $Dz_{fracture}$ of 5–7 km.

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_e 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 overpressure interval.

For rock tensile strengths typical of those measured experimentally and inferred from structural studies (5– 20 MPa) 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 Dz = dz/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.

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 hydro fractures might be propagated upward. This process can be constrained given that the maximum fluid overpressure, relative to fully relaxed conditions, i.e., lithostatic pressure, in the stagnant region.

V. Downward Fluid Flow Through Ductile Rocks

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 (2)) shows that the depth of the brittleductile transition varies with strain rate as l_s ln (e_)/n, provided $A^0(2^m l_s Drg)^n < 1$, as generally the case. This strain rate dependence is identical to that obtained for the maximum depth of downward flow potential gradient

$$z_0 = \frac{l_\sigma}{n}\ln\left(\frac{\dot{\varepsilon}}{A'}\left(\frac{s}{l_\sigma \Delta\rho g}\right)^n\right). \tag{6}$$

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 (5) 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 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 (8), which is valid provided the ductile rheology is dominant at the indicated depth.

VI. Discussion

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. 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, an instability that explains both the nucleation and propagation of hydraulic domains within porous ductile rocks. 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. 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, 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. 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. 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.

Conclusions

This paper proposes a deep learning based ROM for turbulent flows for flow control applications using the Long Short Term Memory (LSTM) neural network , since they have demonstrated immense potential in modeling complex sequential data in other domains. We now outline some merits, limitations for the LSTM-ROM and avenues for further improvement based on our analysis so far. One of the more interesting observations in this work was that Bidirectional LSTM consistently performed worse than the traditional LSTM, despite its theoretical formulation intending otherwise. We surmise that this is likely because it over-fits data, by assuming long range memory that may not have actually existed. While LSTM was more accurate, it was seen that its accuracy deteriorated with an increase in horizon. While we made every effort in this work to tune the neural network hyper-parameters to improve accuracy, there is a possibility that a further improvement could have been obtained. However, we believe that any such gains would have been marginal and the qualitative trends would hold. Furthermore, our accuracy may also be theoretically restricted due to the Lyapunov exponent theory for dynamical systems. Intuitively, in a dynamical system the likelihood of making accurate predictions for time series farther away from the origin, drops exponentially. However, this merely indicates the accurate, pure data-driven approaches to prediction of long time horizons may have several difficulties. Developing governing equation, physics-based models (Navier Stokes equations) with a data-driven approach (LSTM) may still provide us with an accurate and efficient prediction scheme that complements the strengths of both approaches, while not being constrained by the Lyapunov exponent. Additionally, we must address the implicit assumption made here, that the dominant POD spatial modes are consistent within the same regime.

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