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Seismic attenuation due to wave-induced flow

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Abstract. Analytical expressions for three P-wave attenuation mechanisms in sedimentary rocks are given a unified theoretical framework. Two of the models concern wave-induced flow due to heterogeneity in the elastic moduli at “mesoscopic” scales (scales greater than grain sizes but smaller than wavelengths). In the first model, the heterogeneity is due to lithological variations (e.g., mixtures of sands and clays) with a single fluid saturating all the pores. In the second model, a single uniform lithology is saturated in mesoscopic “patches” by two immiscible fluids (e.g., air and water). In the third model, the heterogeneity is at “microscopic” grain scales (broken grain contacts and/or micro-cracks in the grains) and the associated fluid response corresponds to “squirt flow.” The model of squirt flow derived here reduces to proper limits as any of the fluid bulk moduli, crack porosity, and/or frequency is reduced to zero. It is shown that squirt flow is incapable of explaining the measured level of loss \(10^{-2} < Q^{-1} < 10^{-1}\) within the seismic band of frequencies (1 to 10^4 Hz); however, either of the two mesoscopic scale models easily produce enough attenuation to explain field data.

1. Introduction

The physics controlling the intrinsic seismic attenuation of sedimentary rock throughout the seismic band of frequencies (say 1 to 10^4 Hz) is still not entirely understood. In particular, seismic data from sedimentary regions often exhibits more intrinsic attenuation than can be explained using existing theoretical models. The principal goal of this paper is to provide models that can help explain the levels of loss determined from seismograms.

Intrinsic loss is often quantified using the inverse quality factor \(Q^{-1}\) which represents the fraction of wave energy lost to heat in each wave period. For seismic-transmission experiments (earthquake recordings, VSP, cross-well tomography, sonic logs), the total attenuation inferred from the seismograms can be decomposed as \(Q_{\text{total}}^{-1} = Q_{\text{scat}}^{-1} + Q^{-1}\), where both the scattering and intrinsic contributions are necessarily positive. Multiple scattering transfers energy from the coherent first-arrival pulse into the coda and into directions that will not be recorded on the seismogram, and is thus responsible for the effective “scattering attenuation” \(Q_{\text{scat}}^{-1}\). Techniques have been developed that attempt to separate the intrinsic loss from the scattering loss in transmission experiments [e.g., Wu and Aki, 1988, and Sato and Fehler, 1998]. In seismic-reflection experiments, back-scattered energy from the random heterogeneity can sometimes act to enhance the amplitude of the primary reflections. At the present time, techniques that can reliably separate the total inferred loss into scattering and intrinsic portions are generally not available.

Crosswell experiments in horizontally-stratified sediments produce negligible amounts of scattering loss so that essentially all apparent loss (except for easily corrected spherical spreading) is attributable to intrinsic attenuation. Quan and Harris [1997] use tomography to invert the amplitudes of crosswell P-wave first arrivals to obtain the \(Q^{-1}\) for the layers of a stratified sequence of shaly sandstones and limestones (depths ranging from 500-900 m). The center frequency of their measurements is roughly 1750 Hz and they find that \(10^{-2} < Q^{-1} < 10^{-1}\) for all the layers in the sequence. Sams et al. [1997] also measure the intrinsic loss in a stratified sequence of water-saturated sandstones, siltstones and limestones (depths ranging from 50-250 m) using VSP (30-280 Hz), crosswell (200-2300 Hz), sonic logs (8-24 kHz), and ultrasonic laboratory (500-900 kHz) measurements. Sams et al. [1997] calculate (with some inevitable uncertainty) that in the VSP experiments, \(Q^{-1}/Q_{\text{scat}}^{-1} \approx 4\), while in the sonic experiments, \(Q^{-1}/Q_{\text{scat}}^{-1} \approx 19\); i.e., for this sequence of sediments, the intrinsic loss dominates the scattering loss at all frequencies. Sams et al. [1997] also find \(10^{-2} < Q^{-1} < 10^{-1}\) across the seismic band.

It will be demonstrated here that wave-induced fluid flow generates enough heat to explain these measured levels of intrinsic attenuation. Other attenuation mechanisms need not be considered, although they may in fact be present, since they are likely contributing much smaller percentages to the overall observed attenuation. The induced flow occurs at many different spatial scales that can broadly be categorized as “macroscopic,” “mesoscopic,” and “microscopic.”

The macroscopic flow is the wavelength-scale equilibrium occurring between the peaks and troughs of a P-wave. This mechanism was first treated by Biot [1956a, b] and is often simply called “Biot loss.” However, the flow at such macro-scales drastically underestimates the measured loss in the seismic band (by as much as 5 orders of magnitude). Two possible alternatives to Biot loss were therefore proposed in the mid-seventies.

First, Mavko and Nur [1975, 1979], Budiansky and O’Connell [1976], and O’Connell and Budiansky [1977] proposed a microscopic mechanism due to microcracks in the grains and/or broken grain contacts. When a seismic wave squeezes a rock having such grain-scale damage, the cracks
respond with a greater fluid pressure than the main porespace result in a flow from crack to pore that Masko and Nur [1975] named "squirt flow". Dvorak et al. [1995] have also presented a squirt-flow model applicable to liquid-saturated rocks. Although squirt flow seems capable of explaining much of the measured attenuation in the laboratory at ultrasonic frequencies, it may also turn out to be important for propagation in ocean sediments at ultrasonic frequencies [Williams et al., 2002]. We show here that this mechanism cannot explain the attenuation in the seismic band.

Second, White [1975] and White et al. [1975] modeled the wave-induced flow created by mesoscopic-scale heterogeneity. Mesoscopic length scales are those larger than grain sizes but smaller than wavelengths. Heterogeneity across these scales may be due to lithological variations or to patches of different immiscible fluids. When a compressional wave squeezes a material containing mesoscopic heterogeneity, the effect is similar to squirt with the more compliant portions of the material responding to lower fluid pressure than the stiffer portions. There is a subsequent flow of fluid capable of generating significant loss in the seismic band.

White [1975] considered the flow in a concentric porous sphere model in which the inner sphere is saturated by one fluid type (say gas), the outer shell is saturated by another fluid type (say liquid), and the porous frame properties are everywhere uniform. This is the first so-called "patchy saturation" model. White had the insight to use the Biot [1956] theory as the local model for the mesoscopic flow between the spheres. Dutta and Ole [1979a,b] and Dutta and Zerff [1979] went on to make several important corrections to the initial White [1975] model, adding to our understanding of the low-frequency and high-frequency limits. White's [1975] prediction of enhanced attenuation in the presence of even small volume fractions of gas phase has been experimentally confirmed [e.g., Murphy, 1982, 1984; Cadomot et al., 1988].

White et al. [1975] considered the wave-induced flow between the mesoscopic-scale layers in a sedimentary basin. Here the mesoscopic heterogeneity is in the frame properties of the porous rocks with a single fluid saturating all layers. Again, Biot theory was used as the local model for the mesoscopic flow. A host of theoretical refinements have subsequently been added to White's initial model of mesoscopic flow in finely-layered media [e.g., Norris, 1993; Garvenich and Lapatnikov, 1995; Gelnik and Shapiro, 1997].

More recent work by Johnson [2001] has treated wave-induced mesoscopic flow due to patchy saturation without placing restrictions on the patch geometries. The present study also seeks to model the wave-induced flow for arbitrary mesoscopic geometry due either to lithological variations or to patchy saturation, albeit under the restriction that only two porous phases are mixed together in each averaging volume. Furthermore, our same formalism is shown to produce new results at both low and high frequencies for the Dvorak et al. [1995] squirt-flow model.

In section 2, we review the recent theory of Pride and Berryman [2003a,b] treating the mesoscopic loss created by lithological patches having, for example, different degrees of consolidation. This so-called "double-porosity" model provides the theoretical framework that will be used throughout. In section 3, we re-analyze the patchy-saturation model of Johnson [2001] and demonstrate numerically that our double-porosity approach to the problem is asymptotically identical to Johnson's result in the limits of low and high frequencies (both analyses are exact for the model in these two limits, but may differ somewhat at intermediate frequencies). In section 4, we provide a new analysis of the Dvorak et al. [1995] squirt-flow model that is numerically compared to the approximate analysis of Dvorak et al. [1995]. Finally, in the concluding section 5, we summarize what has been learned from these models.

2. Review of the Double-Porosity Theory

In this theory, the mesoscopic heterogeneity is modeled as a mixture of two porous phases saturated by a single fluid.

Various scenarios can be envisioned for how two porous phases might come to reside within a single geological sample. For example, even within an apparently uniform sandstone formation, there can remain a small volume fraction of less-consolidated (even non-cemented) sand grains. This is because diagonal is a transport process sensitive to even subtle heterogeneity in the initial grain pack resulting in spatially variable mineral deposition [e.g., Thompson et al., 1987] and, supposedly, spatially variable elastic moduli. Alternatively, the two phases might correspond to interwoven lenses of detrital sands and clays; however, any associated anisotropy in the deviatory seismic response will not be modeled in the present paper. Jointed rock is also well modeled as a double-porosity material. The joints or macroscopic fractures are typically more compressible and have a higher intrinsic permeability than the background host rock they reside within.

2.1. Local Governing Equations

Each porous phase is locally modeled as a porous continuum and obeys the laws of poroelasticity [e.g., Biot, 1962]

\[ \nabla \cdot \tau_D^p - \nabla p_{\text{ci}} = \rho \ddot{u}_i + \rho_f \ddot{Q}_i, \]

(1)

\[ Q_i = -\frac{1}{\eta} \left( \nabla p_{fi} + \rho \ddot{u}_i \right), \]

(2)

\[ \frac{\nabla \cdot u_i}{\nabla \cdot Q_i} = -\frac{1}{K_i} \left[ \frac{1}{\alpha_i} \frac{p_{\text{ci}}}{p_{fi}} - \frac{1}{\alpha_i} \frac{p_{\text{ci}}}{p_{fi}} \right], \]

(3)

\[ \tau_D^p = G_i \left( \nabla u_i + \nabla u_i^T - \frac{2}{3} \nabla \cdot u_i \mathbf{1} \right), \]

(4)

where the index \( i \) represents the two phases (\( i = 1, 2 \)). The response fields in these equations are themselves local volume averages taken over a scale larger than the grain sizes but smaller than the mesoscopic extent of either phase. The local fields are: \( u_i \), the average displacement of the framework of grains; \( Q_i \), the Darcy filtration velocity; \( p_{fi} \), the fluid pressure; \( p_{\text{ci}} \), the confining pressure (total average pressure); and \( \tau_D^p \), the deviatoric (or shear) stress tensor. In the linear theory of interest here, the overdots on these fields denote a partial time derivative. In the local Darcy law (2), \( \eta \) is the fluid viscosity and the permeability \( k_i \) is a linear time-convolution operator whose Fourier transform \( k_i(\omega) \) is called the "dynamic permeability" and can be modeled using the theory of Johnson et al. [1987] (see the appendix).

In the local compressibility law (3), \( K_i^p \) is the drained bulk modulus of phase \( i \) (confining pressure change divided by sample dilatation under conditions where the fluid pressure does not change). \( B_i \) is Skempton's [1954] coefficient of phase \( i \) (fluid pressure change divided by confining pressure change for a sealed sample), and \( \alpha_i \) is the Biot and Willis [1957] coefficient of phase \( i \) defined as

\[ \alpha_i = \left( 1 - K_i^p / K_i^w \right) / B_i, \]

(5)

where \( K_i^w \) is the undrained bulk modulus (confining pressure change divided by sample dilatation for a sealed sample). In the present work, no restrictions to single-mineral isotropic grains will be made. Finally, in the deviatoric constitutive law (4), \( G_i \) is the shear modulus of the framework of grains. At the local level, all these poroelastic constants are taken to be real constants. In the appendix we give the Gassmann [1951] fluid-substitution relations that allow \( B_i \) and \( \alpha_i \) to
2.2. Double-Porosity Governing Equations

In the double-porosity theory, the goal is to determine the average fluid response in each of the porous phases in addition to the average displacement of the solid grains [Berryman and Wang, 1995]. The averages are taken over regions large enough to significantly represent both porous phases, but smaller than wavelengths. Assuming an $e^{-ict}$ time dependence, Pride and Berryman [2003a] have used volume averaging of the local laws (1)-(4) to obtain the macroscopic "double-porosity" governing equations in the form

$$\nabla \cdot \tau^D - \nabla P_c = -i\omega (\rho v + \rho_f q_1 + \rho_f q_2),$$

and

$$\begin{bmatrix} q_1 \\ q_2 \end{bmatrix} = -\frac{1}{\eta} \begin{bmatrix} \kappa_{11} & \kappa_{12} \\ \kappa_{12} & \kappa_{22} \end{bmatrix} \begin{bmatrix} \nabla P_f - i\omega \rho_f v \\ \nabla P_f - i\omega \rho_f v \end{bmatrix}.$$  

$$\begin{bmatrix} \nabla \cdot v \\ \nabla \cdot q_1 \\ \nabla \cdot q_2 \end{bmatrix} = i\omega \begin{bmatrix} a_{11} & a_{12} & a_{13} \\ a_{12} & a_{22} & a_{23} \\ a_{13} & a_{23} & a_{33} \end{bmatrix} \begin{bmatrix} P_c \\ \nabla P_f + i\omega \zeta_{\text{int}} \\ -\zeta_{\text{int}} \end{bmatrix}.$$  

$\gamma(\omega) (\nabla P_f - \nabla P_f^*) = -i\omega \zeta_{\text{int}} (i\omega \gamma/\omega),$  

and

$$-i\omega \tau^D = [G(\omega) - i\omega g(\omega)] \left( \nabla v + (\nabla v)^T \right) - \frac{2}{3} \nabla \cdot v I.$$  

The macroscopic fields are: $v$, the average particle velocity of the solid grains throughout an averaging volume of the composite; $q_i$, the average Darcy flux across phase $i$; $P_c$, the average total pressure in the averaging volume; $\tau^D$, the average deviatoric stress tensor; $P_f$, the average fluid pressure within phase $i$; and $-i\omega \zeta_{\text{int}}$, the average rate at which fluid volume is being transferred from phase 1 to phase 2 as normalized by the total volume of the averaging region. The dimensionless increment $\zeta_{\text{int}}$ represents the "mesoscopic flow."  

Equation (7) is the generalized Darcy law allowing for fluid cross-coupling between the phases [c.f., Pride and Berryman, 2003b]; equation (8) is the generalized compressibility law where $\nabla \cdot q_1$ corresponds to fluid that has been depleted from phase 1 due to transfer across the external surface of an averaging volume, and equation (9) is the transport law for internal mesoscopic flow (fluid transfer between the two porous phases).  

The coefficients $a_{ij}$ and $\gamma$ in these equations have been modeled in detail by Pride and Berryman [2003a,b]. Before presenting these results in the following subsections, the nature of the waves implicitly contained in these laws is briefly commented upon. If plane-wave solutions for $v$, $q_1$ and $q_2$ are introduced, there is found to be a single transverse wave, and three longitudinal responses: a fast wave and two slow waves [Berryman and Wang, 2000]. The fast wave is the usual P-wave identified on seismograms, while the two slow waves correspond to fluid-pressure diffusion in phases 1 and 2. The only problem with analyzing the fast compressional wave in this manner is that the characteristic equation for the longitudinal slowness $s$ is cubic in $s^2$ and, therefore, somewhat inconvenient for analysis.

2.3. Reduction to an Effective Biot Theory

The approach that we take instead is to reduce these double-porosity laws (6)-(10) to an effective single-porosity Biot theory having complex frequency-dependent coefficients. The easiest way to do this is to assume that phase 2 is entirely embedded in phase 1 so that the average flux $q_2$ into and out of the averaging volume across the external surface of phase 2 is zero. By placing $\nabla \cdot q_2 = 0$ into the compressibility laws (8), the fluid pressure $P_f$ can be entirely eliminated from the theory. In this case the double-porosity laws reduce to effective single-porosity poroelasticity governed by laws of the form (3) but with effective poroelastic moduli given by

$$\frac{1}{K_D} = a_{11} - \frac{a_{13}^2}{a_{33} - \gamma/\omega},$$

$$B = -a_{12} (a_{33} - \gamma/\omega) + a_{13} (a_{23} + \gamma/\omega),$$

and

$$\frac{1}{K_U} = \frac{1}{K_D} + B \left( \frac{a_{12}}{a_{33} - \gamma/\omega} \right).$$

Here, $K_D(\omega)$ is the effective drained bulk modulus of the double-porosity composite, $B(\omega)$ is the effective Skempton’s coefficient, and $K_U(\omega)$ is the effective undrained bulk modulus. An effective Biot-Willis constant can then be defined using $\alpha(\omega) = \{1 - K_D(\omega)/K_U(\omega)\}/B(\omega)$.  

The complex frequency dependent “drained” modulus $K_D$ defines the total volumetric response when the average fluid pressure throughout the host phase 1 is unchanged. Due to the fluid pressure differences between the two phases, fluid-pressure equilibration ensues which results in $K_D$ being complex and frequency dependent. Similar interpretations hold for the undrained moduli $K_U$ and $B$. An undrained response is when no fluid can escape or enter through the external surface of an averaging volume; however, there can be considerable internal exchange of fluid between the two phases resulting in the complex frequency-dependent nature of both $K_U$ and $B$.

2.4. Double-Porosity $a_{ij}$ Coefficients

The constants $a_{ij}$ are all real and correspond to the high-frequency response for which no internal fluid-pressure relaxation can take place. They are given exactly as [Pride and Berryman, 2003a]

$$a_{11} = 1/K,$$

$$a_{22} = v_1 a_1 K_1^4 \left( \frac{1}{K_1} - \frac{a_1 (1 - Q_1)}{K_1^4} \right),$$

$$a_{33} = v_2 a_2 K_2^4 \left( \frac{1}{K_2} - \frac{a_2 (1 - Q_2)}{K_2^4} \right),$$

$$a_{12} = -v_1 Q_1 a_1 K_1^4,$$

$$a_{13} = -v_2 Q_2 a_2 K_2^4,$$

$$a_{23} = -a_1 a_2 K_1^4 K_2^4 \left( \frac{1}{K_1} - \frac{v_1}{K_1^4} + \frac{v_2}{K_2^4} \right),$$

where the $Q_i$ are auxiliary constants given by

$$v_1 Q_1 = \frac{1 - K_2^4/K_1}{1 - K_2^4/K_1^4},$$

and

$$v_2 Q_2 = \frac{1 - K_1^4/K_2}{1 - K_1^4/K_2^4}.$$  

Here, $v_1$ and $v_2$ are the volume fractions of each phase within an averaging volume of the composite.  

The one constant that has not yet been determined is the overall drained modulus $K = 1/a_{11}$ of the two-phase composite (the modulus defined in the quasi-static limit where the local fluid pressure throughout the composite is everywhere unchanged). It is through $K$ that the $a_{ij}$ acquire their dependence on both the mesoscopic geometry and shear properties of each porous phase. Having expressions for how $K$ depends on the properties of the two constituents is quite useful even though an exact analytical model applicable to any given double-porosity scenario may not be known.  

The Hashin and Shtrikman [1963] bounds for the overall low-frequency drained bulk modulus $K$ and shear modulus
2.5. Double-Porosity Transport

Pride and Berryman [2003b] obtain the internal transport coefficient \( \gamma \) of equation (9) as

\[
\gamma(\omega) = \gamma_m \sqrt{1 - \frac{i \omega}{\omega_m}}
\]

(24)

where \( \gamma_m \) and \( \omega_m \) are parameters dependent on the constituent properties and the mesoscopic geometry. To obtain useful analytical results, some type of approximation is required.

Normally, the double-porosity model is useful (or necessary) only in situations where the two phases have strong contrasts in their physical properties. When the embedded phase 2 is much more permeable than the host phase 1, Pride and Berryman [2003b] find that

\[
\gamma_m = -\frac{k_1 K_i^d}{\eta L_i^4} \left( \frac{a_{12} + B_o (a_{22} + a_{33})}{R_1 - B_o / B_1} \right) \left[ 1 + O(k_1/k_2) \right],
\]

(25)

where the \( a_{ij} \) are given by equations (14)–(19) and where the remaining terms \( B_o \), \( L_1 \) and \( R_1 \) are now defined.

The dimensionless quantity \( B_o \) is the static Skempton’s coefficient for the composite and is given exactly by

\[
B_o = \frac{(a_{12} + a_{13})}{a_{22} + 2a_{23} + a_{33}}
\]

(26)

regardless of the mesoscopic geometry.

The length \( L_1 \) characterizes the average distance in phase 1 over which the fluid-pressure gradient still exists in the final approach to equilibration and has the formal mathematical definition

\[
L_1^2 = \frac{1}{V_1} \int_{\Omega_1} \frac{dV}{\nabla \cdot \Phi_1}
\]

(27)

where \( \Omega_1 \) is the region of an averaging volume occupied by phase 1 and has a volume measure \( V_1 \). The potential \( \Phi_1 \) has units of length squared and is a solution of an elliptic boundary-value problem that under conditions where the harmonic mean is a good approximation for the overall drained modulus and where the permeability ratio \( k_1/k_2 \) can be considered small, reduces to

\[
\nabla \cdot \Phi_1 = -1 \quad \text{in} \quad \Omega_1,
\]

(28)

\[
\nabla \cdot \Phi_1 = 0 \quad \text{on} \quad \partial E_1,
\]

(29)

\[
\Phi_1 = 0 \quad \text{on} \quad \partial \Omega_12,
\]

(30)

where \( \partial E_1 \) is the external surface of the averaging volume coincident with phase 1, and where \( \partial \Omega_12 \) is the internal interface separating phases 1 and 2. Multiplying equation (28) by \( \Phi_1 \) and integrating over \( \Omega_1 \), establishes that second integral of equation (27).

The dimensionless quantity \( R_1 \) is the ratio of the average static confining pressure in phase 1 to the pressure applied to the external surface of a sealed sample of the composite. Pride and Berryman [2003a] derive this ratio to be

\[
R_1 = \frac{Q_1 + \alpha_1 (1 - Q_1) B_o}{1 - K_i^{d2}/K_i^d} - \frac{v_2 \alpha_2 (1 - Q_2) B_o}{v_1 - 1 - K_i^{d2}/K_i^d}
\]

(31)

where the \( Q_i \) are given by equation (20). Thus, once the overall drained modulus \( K \) is chosen (e.g., using the Hashin and Shtrikman [1963] lower bound), \( \gamma_m \) can now be determined from equation (25).

If it is more appropriate to consider the host phase 1 as being more permeable than the embedded phase 2 (\( k_2/k_1 \ll 1 \)), one needs only to exchange indices 1 and 2 throughout all of equations (25)–(31).

In passing, if it is assumed that the harmonic mean is a reasonable approximation for the drained modulus of the composite (i.e., \( 1/K = v_1/K_i^d + v_2/K_i^d \)) then \( Q_1 = 1 \), \( a_{22} = 0 \), \( R_1 = 1 \) and all of the above expressions exactly reduce to

\[
\gamma_m = \frac{v_1 k_1}{\eta L_1^4} \left[ 1 + O(k_1/k_2) \right].
\]

(32)

However, the harmonic mean for \( K \) is special, and not always appropriate, so we consider the lower Hashin and Shtrikman [1963] bound as the preferable modeling choice for most geological situations of interest.

The transition frequency \( \omega_m \) corresponds to the onset of a high-frequency regime in which the fluid-pressure-diffusion penetration distance between the phases becomes small relative to the scale of the mesoscopic heterogeneity. It is given by Pride and Berryman [2003b] to be

\[
\omega_m = \frac{\eta B_i K_i^d}{k_1 \alpha_1} \left( \frac{V}{S} \right)^2 \left( 1 + \frac{k_1 B_i K_i^d \alpha_1}{k_2 B_i K_i^d \alpha_2} \right)^2.
\]

(33)
The length $V/S$ is is the volume-to-surface ratio where $S$ is the area of $\partial\Omega_2$ in each volume $V$ of composite.

The geometry of the phase 2 inclusion is affecting the length parameters $L_1$ and $V/S$ as well as the drained modulus $K$. Putting in a highly complicated multi-scale distribution of phase 2 (even a fractal distribution) changes the values of these three numbers but does not change the analytic structure of the above results for $\gamma_m$ and $\omega_m$.

For complicated geometry, the length $L_1$ can only be determined numerically or inverted for from data. For idealized geometries it can be analytically estimated. For example, in a concentric-sphere geometry with $k_2/k_1 \ll 1$, Pride and Berryman [2003b] obtain

$$L_2^2 = \frac{9}{14} R^2 \left[ 1 - \frac{7}{6} \frac{a}{R} + O(a^3/R^3) \right],$$

where $a$ is the radius of each sphere of phase 2 embedded within each sphere $R$ of composite. The volume fraction $v_2$ of embedded spheres is $v_2 = (a/R)^3$ in this case so that $R$ can be eliminated using $R = a/v_2^{1/3}$. In the alternative case where $k_2/k_1 \ll 1$, the length $L_2$ for this same concentric-sphere geometry is [e.g., Johnson, 2001] $L_2^2 = a^2/3$.

In the scenario of interest in which phase 2 is taken to be penny-shaped lenses of one more compliant material mixed into a stiffer phase-1 host, the length parameter $L_1$ can at least be approximately estimated. Assuming that each penny-shaped inclusion has a radius $a$ and a thickness $\varepsilon a$ where $\varepsilon$ is the aspect ratio of the inclusion, one can estimate $\Phi_1$ using a simple slab geometry. With the volume fraction $v_2$ and both $a$ and $\varepsilon$ treated as user-controlled parameters, one finds that $V/S = a^2/(2v_2)$ and $L_1^2 = a^2/12$. These estimates for $L_1$ and $V/S$ along with the Hashin and Shtrikman [1963] lower bound for $K$ and $G$ will be the model treated in the numerical examples that follow.

The coefficient $G(\omega) - i\omega(g(\omega)$ governing shear generally has a non-zero “viscosity” $g(\omega)$ associated with the mesoscopic fluid transport between the compressional lobes surrounding a sheared phase 2 inclusion. Both of the frequency functions $G(\omega) - i\omega(g(\omega)$ are real and are Hilbert transforms of each other. The frequency dependence of $g(\omega)$ was not modeled by Pride and Berryman [2003b], but is presently being analyzed further by these authors. Here, we continue to ignore any possible dispersion in the shear properties and take $G$ to be a real constant given by the Hashin and Shtrikman [1963] lower bound.

Finally, the dynamic permeability $k(\omega)$ to be used in the effective Biot theory can be modeled in several ways. The appropriate modeling choice when phase 2 is modeled as small inclusions embedded in phase 1 is the harmonic mean $1/k(\omega) = v_1/k_1(\omega) + v_2/k_2(\omega) \approx v_1/k_1(\omega)[1 + O(v_2k_2/k_1)]$.

2.6. Phase Velocity and Attenuation

With all of the double-porosity coefficients now defined, the compressional phase velocity and attenuation can be determined by inserting a plane-wave solution into the effective single-porosity Biot equations [of the form (1)-(4)]. This gives the standard complex longitudinal slowdown $s$ of Biot theory

$$s^2 = b \mp \sqrt{b^2 - \frac{\rho^2 \rho_0^2}{MH - C^2}},$$

where

$$b = \frac{\rho_0 M + \rho H - 2\rho_0 C}{2(MH - C^2)}$$

is simply an auxiliary parameter, and where $H$, $C$, and $M$ are the Biot [1962] poroelastic moduli defined in terms of the complex frequency-dependent parameters of equations (11)-(13) as

$$H = K_U + 4G/3,$$

$$C = BK_U,$$

$$M = \frac{b^2}{1 - KD/K_U} K_U.$$

Taking the minus sign in equation (34) gives an $s$ having an imaginary part much smaller than the real part and that thus corresponds to the normal P-wave. Taking the positive sign gives an $s$ with real and imaginary parts of roughly the same amplitude and that thus corresponds to the slow P-wave (a pure fluid-pressure diffusion across the seismic band of frequencies). We are only interested here in the properties of the normal P-wave.

The P-wave phase velocity $v_p$ and the attenuation measure $Q_p^{-1}$ are related to the complex slowdown $s$ as

$$v_p = 1/Re\{s\},$$

$$Q_p^{-1} = \text{Im}\{s^2\}/Re\{s^2\}.$$
Figure 2. A comparison of modeling the embedded phase 2 as either penny-shaped lenses or spheres. All curves have identical phase 1 and phase 2 material properties and identical phase 2 volume fractions $v_2 = 2\%$. The only difference is the assumed shape of the phase 2 inclusion, which has a strong influence on the overall drained bulk modulus of the composite (the Hashin and Shtrikman [1963] upper bound holds in the case of spheres, while the lower bound holds in the case of penny-shaped lenses).

Figure 3. Attenuation as predicted by the double-porosity model of Pride and Boros [2003a] (the solid curves) as compared to the data of Sams et al. [1997] (rectangular boxes). The number of $Q^{-1}$ estimates determined by Sams et al. [1997] falling within each rectangular box are: 40 VSP, 69 crosswell, 854 sonic log and 46 ultrasonic core measurements. A similar number of velocity measurements were made. These various measurements come from different depth ranges at their test site.

In this way, the moduli $K^2_{ij}$ and $G_2$ are functions of the background effective-stress level $P_e$. The host phase 1 is modeled using $\phi_1 = 0.20$ and $c = 2$ in the model given in the appendix. All mineral moduli are taken to be that of quartz $K_s = 38$ GPa and $G_s = 44$ GPa and the permeability of the host phase is $k_1 = 10$ mD. The drained properties of the composite were modeled using the Hashin and Shtrikman [1963] lower bounds given in equations (21)–(22). The penny-shaped inclusion of phase 2 have the following geometrical properties: $a = 3$ cm, $c = 10^{-2}$, $v_2 = 3\%$, $L_1 = 8.6$ mm, and $V/S = 5$ mm. The specific shape of the attenuation curve is highly sensitive to whether $L_1$ is greater than or less than $V/S$. The invariant peak near $10^{6}$ Hz is due to the Biot loss (fluid equilibration at the scale of the seismic wavelength), while the broad main peak that changes with the effective pressure $P_e$ is that due to mesoscopic-scale equilibration. All dependence on $P_e$ in this example comes from how $K^2_{ij}$ and $G_2$ vary with $P_e$.

The level of attenuation in the double-porosity theory is controlled by the factors that allow phase 2 to develop a different fluid pressure response as compared to phase 1. In figure 2, this is demonstrated by comparing phase 2 modeled as spheres to phase 2 modeled as penny-shaped lenses. Both examples have identically the same volume fractions of phase 2 as well as phase 1 and 2 material properties. The difference is that in the sphere model, the Hashin and Shtrikman [1963] upper bound is used for $K$ and $G$, while the lower bound is used in the penny-shaped lens model. A compliant sphere of phase 2 is protected from an applied compression by the rigidity of the phase 1 host that surrounds it. Accordingly, not much fluid pressure difference is created between the two phases and so there is only a small amount of mesoscopic loss.

In modeling the penny-shaped inclusions in figure 2, we have used the parameter values $a = 3$ cm (inclusion radius) and $\varepsilon = 10^{-1}$ to obtain $V/S = 5$ cm and $L_1 = 0.9$ cm. In this case, $V/S > L_1$ which has changed considerably the look of the attenuation curve as compared to figure 1 where $V/S < L_1$. What is happening can be seen in the effective moduli of equations (11)–(13). The principal relaxation in the effective moduli occurs whenever $\omega = \gamma/\alpha_{ij}$. But there is also a relaxation in $\gamma(\omega)$ when $\omega = \omega_m$. For situations where $V/S \gg L_1$, the effective moduli relax at a frequency much less than $\omega_m$ (with $\gamma(\omega) = \gamma_m$) and this is the case in figure 2. When $V/S < L_1$, the relaxation in $\gamma(\omega)$ can begin to occur prior to the principal relaxation as is seen in figure 1.

Finally, in figure 3, we compare the double-porosity model to the data of Sams et al. [1997] who used different seismic measurements (VSP, crosswell, sonic log, and ultrasonic lab) to determine $Q^{-1}$ and $P$-wave velocity over a wideband of frequencies at their test site in England. The variance of the measurements falling within each rectangular box are due to the various rock layers present at this site. Data collection

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Data collection
was between four wells that are a few hundred meters deep. The geology at the site is a sequence of layered limestones, sandstones, siltstones, and mudstones. We modeled phase 2 as unconsolidated penny-shaped inclusions in which $a = 5 \text{ cm}$ (inclusion radius), $\varepsilon = 6 \times 10^{-3}$, $v_2 = 1.2 \%$, $k_1 = 80 \text{ mD}$, $V/S = 1.25 \text{ cm}$, and $L_1 = 1.45 \text{ cm}$. The phase 1 host is again taken to be a well-consolidated sandstone.

2.8. Discussion

The overall magnitude of attenuation in the double-porosity model is dominantly controlled by the contrast of compressibilities between the two porous phases and the assumed shape of the embedded phase. Certain assumed shapes, such as spherical inclusions, allow the rigidity of the host phase to protect even a very soft inclusion from being compressed much and this results in minimal mesoscopic loss for such a geometry. Less compact and more elongated or even dendritic mesoscopic geometries are what potentially allow the mesoscopic-loss to be important. However, even in the presence of such structure, a strong contrast in the drained properties of the two phases is also required in order to generate a significant mesoscopic fluid-pressure gradient and mesoscopic loss. A contrast in permeability alone would generate no such mesoscopic-scale fluid pressure gradients.

The relaxation frequency at which the mesoscopic loss per cycle is maximum is proportional to $\eta k_1 / L_2^2$. Far below this relaxation frequency, $Q^{-1}$ always increases linearly with frequency as $f n / k_1$. Thus, the permeability information in the double-porosity attenuation is mainly in the frequency dependence of $Q^{-1}$, not in the overall magnitude of $Q^{-1}$, and involves (most importantly) the permeability $k_1$ of the host phase, not the overall permeability of the composite (see Berrymann [1988] for a related discussion). If phase 2 is well modeled as being small penny-shaped inclusions embedded in phase 1, then $k_1$ is controlling the overall permeability. If phase 2 corresponds to through-going connected joints, then although $Q^{-1}(\omega)$ contains information about $k_1$, it does not contain information about the overall permeability which is being dominated by $k_2$ in this case (i.e., $k_2$ has no significant influence on the mesoscopic-loss process).

In the case of through-going joints, the equilibrium at the scale of the wavelength (the Biot loss) has a chance of being shifted to lower frequencies. The only way to determine the proper attenuation curve in this case is to solve the cubic characteristic equation for $s^2$ [the characteristic equation is obtained by inserting a plane-wave solution into the complete double-porosity equations (6)-(10) as discussed earlier].

3. Patchy-Saturation Model

Another important source of mesoscopic-scale heterogeneity having an important influence on seismic properties is patchy fluid saturation [e.g., Knight et al., 1998]. All natural hydrological processes by which one fluid non-miscibly invades a region initially occupied by another result in a patchy distribution of the two fluids. The patch sizes are distributed across the entire range of mesoscopic length scales and for many invasion scenarios are expected to be fractal. As a compositional wave squeezes such a material, the patches occupied by the less-compressible fluid will respond with a greater fluid-pressure change than the patches occupied by the more-compressible fluid. The two fluids will then equilibrate by the same type of mesoscopic flow already modeled in the double-porosity model.

An analysis almost identical to that of Pride and Berrymann [2003a,b] can be carried out that leads to the same effective poroelastic moduli given by equations (11)-(13) but with different definitions of the $a_{ij}$ constants and internal transport coefficient $\gamma(\omega)$. In the model, a single uniform porosity frame is saturated by mesoscopic-scale patches of fluid 1 and fluid 2. We define porous phase 1 to be those regions (patches) occupied by the less mobile fluid and phase 2 the patches saturated by the more mobile fluid: i.e., by definition $\eta_1 > \eta_2$. This most often (but not necessarily) corresponds to $K_1 > K_2$ and, therefore, to $B_1 > B_2$.

Johnson [2001] has treated this model using a different coarse-graining argument while starting from the same local physics (however, he assumes the porous material is a Gaussmann-style mono-mineral material). Our final undrained bulk modulus is identical to the result of Johnson [2001] in the limits of high and low frequency and differs only slightly in the transition range of frequencies where the flow in either model is not explicitly treated.

3.1. Patchy-Saturation $a_{ij}$ Coefficients

To obtain the $a_{ij}$ for the patchy-saturation model, we note that, by model assumption, each patch has the same $\alpha$ and $K$. The poroelastic differences between patches is entirely due to $B_1$ being different than $B_2$. Upon volume averaging equation (3) and using $\nabla \cdot \mathbf{v} = \nabla \cdot (v_2 \mathbf{\tau}_2) + \nabla \cdot (v_1 \mathbf{\tau}_1)$, where an overline again denotes a volume average over the appropriate phase, and using the fact that the $a_{ij}$ are defined in the extreme high-frequency limit where the fluids have no time to traverse the internal interface $\partial B_{12}$ (i.e., the $a_{ij}$ are defined under the condition that $\zeta_{ab} = 0$), one has

$$\nabla \cdot \mathbf{v} = -\frac{\nu_1}{K} \mathbf{\tau}_{cl} - \frac{\nu_2}{K} \mathbf{\tau}_{c2} + \frac{\nu_1 \alpha}{K} \mathbf{\tau}_{f1} + \frac{\nu_2 \alpha}{K} \mathbf{\tau}_{f2},$$

$$\nabla \cdot \mathbf{q}_1 = \frac{\nu_1 \alpha}{K} \mathbf{\tau}_{c1} - \frac{\nu_1 \alpha}{K} \mathbf{\tau}_{f1},$$

$$\nabla \cdot \mathbf{q}_2 = \frac{\nu_2 \alpha}{K} \mathbf{\tau}_{c2} - \frac{\nu_2 \alpha}{K} \mathbf{\tau}_{f2}.$$

The average confining pressures $\mathbf{\tau}_{ci}$ in each phase are not a priori known; however, they are necessarily linear functions of the three independent applied pressures of the theory $P_i = v_1 \mathbf{\tau}_{ci} + v_0 \mathbf{\tau}_{c2}$, $\mathbf{\tau}_{f1}$, and $\mathbf{\tau}_{f2}$. It is straightforward to demonstrate that, if and only if the average confining pressures take the form

$$v_1 \mathbf{\tau}_{f1} = \nu_1 \mathbf{\tau}_{c1} + \beta \mathbf{\tau}_{f1} - \beta \mathbf{\tau}_{f2},$$

$$v_2 \mathbf{\tau}_{c2} = \nu_2 \mathbf{\tau}_{c2} - \beta \mathbf{\tau}_{f1} + \beta \mathbf{\tau}_{f2},$$

then equations (42)-(44) will produce $a_{ij}$ that satisfy the thermodynamic symmetry requirement of $a_{ij} = a_{ji}$ [i.e., these $a_{ij}$ constants are all second derivatives of a strain-energy function as demonstrated by Pride and Berrymann, 2003a]. Upon placing equations (45) and (46) into equations (42)-(44), we then have

$$a_{11} = 1/K,$$

$$a_{22} = (-\beta + v_1 / B_1) \alpha / K,$$

$$a_{33} = (-\beta + v_2 / B_2) \alpha / K,$$

$$a_{12} = -v_1 \alpha / K,$$

$$a_{13} = -v_2 \alpha / K,$$

$$a_{23} = \beta \alpha / K,$$

where $\beta$ is the single constant remaining to be determined. To obtain $\beta$, we note that in the high-frequency limit, each local patch of phase $i$ is undrained and thus characterized by an undrained bulk modulus $K_i^* = K_i / (1 - \alpha B_i)$ and a shear modulus $G$ that is the same for all patches. In this limit, the usual laws of elasticity (as opposed to those of poroelasticity) govern the response of the composite. Note that, even if the rock frame is spatially uniform, an exception to uniform $G$ can, in principle, occur if cracks are uniformly
present. In this case, it is known (see Berryman et al. [2002]) that the shear modulus in the regions containing dry cracks can be somewhat different from the shear modulus in the regions containing wet cracks. In reality, however, all cracks tend to be water wet in partially saturated rocks and it is a physically reasonable approximation to assume that $G$ is the same for each phase even when cracks are present.

Under these precise conditions (elasticity of an isotropic composite having uniform $G$ and all heterogeneity confined to the bulk modulus which in the present case corresponds to $K'$), we follow Johnson [2001] by invoking the theorem of Hill [1963] which states that the overall undrained-unrelaxed modulus of the composite $K_H$ is given exactly by

$$\frac{1}{K_H + 4G/3} = \frac{v_1}{K_1' + 4G/3} + \frac{v_2}{K_2' + 4G/3}.$$  \hspace{1cm} (53)

In terms of the $a_{ij}$, this same undrained-unrelaxed Hill modulus is given by

$$\frac{1}{K_H} = a_{11} + a_{12} \left( \frac{\delta p_{f1}}{\delta P_C} \right) + a_{13} \left( \frac{\delta p_{f2}}{\delta P_C} \right),$$  \hspace{1cm} (54)

where, upon using $\nabla \cdot q = 0$ and $\dot{\zeta}_{int} = 0$ in equation (8) and then using (47)–(52), the undrained-unrelaxed pressure ratios are

$$\frac{\delta p_{f1}}{\delta P_C} = \beta - \frac{v_1v_2}{B_2},$$  \hspace{1cm} (55)

$$\frac{\delta p_{f2}}{\delta P_C} = \frac{v_1}{B_1} + \frac{v_2}{B_2} \left[ \frac{\alpha - (1 - K/K_H)/(v_1B_1 + v_2B_2)}{\alpha - (1 - K/K_H)/(v_1B_1 + v_2B_2)} \right].$$  \hspace{1cm} (56)

Thus, after some algebra, equation (54) yields the exact result

$$\beta = \frac{v_1v_2}{B_2} \left[ \frac{\alpha - (1 - K/K_H)/(v_1B_1 + v_2B_2)}{\alpha - (1 - K/K_H)/(v_1B_1 + v_2B_2)} \right].$$  \hspace{1cm} (57)

with $K_H$ given by equation (53). All the $a_{ij}$ are now expressed in terms of known quantities.

3.2. Patchy-Saturation Transport

Next we must address the internal fluid-pressure equilibrium between the two phases with the goal of obtaining the internal transfer coefficient $\gamma$ of equation (9). The mathematical definition of the rate of internal fluid transfer is

$$\dot{\zeta}_{int} = \frac{1}{V} \int_{\partial\Omega_{12}} \mathbf{n} \cdot \mathbf{Q}_i \, dS,$$  \hspace{1cm} (58)

where $V$ is the volume occupied by the composite. A possible concern in the patchy-saturation analysis is whether capillary effects at the local interface $\partial\Omega_{12}$ separating the two phases need to be considered.

3.2.1. Local continuity conditions on $\partial\Omega_{12}$.

At the pore scale, the interface separating one fluid patch from the next is a series of menisci. Roughness on the grain surfaces keeps the contact lines of these menisci pinned to the grain surfaces. Pride and Fleckkoy [1999] argue that the contact lines of an air-water meniscus will remain pinned for fluid-pressure changes less than roughly $10^5$ Pa, which corresponds to the pressure range induced by linear seismic waves. So as a wave passes, the menisci will bulge and change shape but will not migrate away. This makes the problem much simpler to analyze.

One porous-continuum boundary condition is that all fluid volume that locally enters the interface $\partial\Omega_{12}$ from one side, must exit the other side so that $\mathbf{n} \cdot \mathbf{Q}_i = \mathbf{n} \cdot \mathbf{Q}_i = 0$. Another boundary condition is that the difference in the rate at which energy is entering and leaving the interface is entirely due to the work performed in changing the miniscus surface area. Before the wave arrives, each

miniscus has a mean curvature $H_0$ fixed by the static fluid pressures initially present: $p_{f1}' = p_{f2}' = \sigma H_0$, where $\sigma$ is the surface tension. During wave passage, one can demonstrate [Pride and Fleckkoy, 1999] that the mean curvature changes as $H = H_0 + \epsilon H_1 + O(\epsilon^2)$, where $H_1$ is of the same order as $H_0$, and where $\epsilon$ is a dimensionless number called the capillary number. The capillary number is defined $\epsilon = \eta |Q|/\sigma$, where $|Q|$ is some estimate of the wave-induced Darcy flux and that is thus bounded by the wave strain times phase velocity, i.e., $|Q| < \epsilon \times 10^{-3}$ m/s. For typical interfaces (like air and water), we have $\sigma > 10^{-7}$ Pa m and $\eta \approx 10^{-3}$ Pa s.

For linear wave problems, $\epsilon \ll 1$ and, thus $\epsilon$ can be considered a very small number.

By writing the fluid pressures as $p_{f1}' = p_{f1}'' + \delta p_{f1}$ and using the fact that $\mathbf{n} \cdot \mathbf{Q}$ is continuous, the conservation of energy at the interface may be expressed as

$$\mathbf{n} \cdot \left( \mathbf{Q} H_0 + [1 + O(\epsilon)] \right),$$  \hspace{1cm} (59)

The brackets on the left-hand side denote the jump in energy flux across the interface, while the right-hand side represents the rate at which work is performed in stretching the meniscus. Since conservation of momentum requires $\mathbf{n} \cdot \mathbf{u}$ to be continuous at the interface and since the assumption of the grains being welded together (or having an overburden effective pressure $(1 - \phi)(\rho_0 - \rho_f)\rho_0$ acting on them that is greater than the wave stress) requires that $\mathbf{u}$ is continuous, we find, to leading order in $\epsilon$, that

$$\delta p_{f1} = \delta p_{f2}$$  \hspace{1cm} (60)

along the interface $\partial\Omega_{12}$. This means that the fluid pressure equilibrium can be modeled using the standard displacement-stress continuity conditions along $\partial\Omega_{12}$ that were also employed in the double-porosity analysis; i.e., capillary effects can be neglected. In what follows, the fluid pressures correspond to the changes induced by the wave and so we cease to write the "$\delta$" explicitly in front of them.

3.2.2. Mesoscopic flow equations.

To obtain the transport law $-i\omega \zeta_{int} = \mathbf{n} \cdot \mathbf{Q}_i = \mathbf{n} \cdot \mathbf{Q}_i$, the mesoscopic flow is analyzed in the limits of low and high frequencies. These limits are then connected using a frequency function that respects causality constraints. The linear fluid response inside the patchy composite due to a seismic wave can always be resolved into two portions: (1) a vectorial response due to macroscopic fluid-pressure gradient fields across an averaging volume that generate a macroscopic Darcy flux $\mathbf{q}_i$ across each phase and that corresponds to the macroscopic conditions $\mathbf{p}_{f1} = 0$ and $\nabla \mathbf{p}_{f1} \neq 0$; and (2) a scalar response associated with internal fluid transfer and that corresponds to the macroscopic conditions $\mathbf{p}_{f1} = 0$ and $\nabla \mathbf{p}_{f1} = 0$. The macroscopic isotropy of the composite guarantees that there is no cross-coupling between the vectorial transport $\mathbf{q}_i$ and the scalar transport $\zeta_{int}$ within each sample ("Curie’s principle" which is, in fact, a theorem; c.f., de Groot and Mazur [1984]).

The mesoscopic flow problem that defines $\zeta_{int}$ is the internal equilibrium of fluid pressure between the patches when a confining pressure $P_f$ has been applied to a sealed sample of the composite. Having the external surface sealed is equivalent to the required macroscopic constraint that $\nabla \mathbf{p}_{f1} = 0$. Upon taking the divergence of (2) and using equation (3), the diffusion problem controlling the mesoscopic flow becomes

$$\frac{k}{\eta} \nabla^2 p_{f1} + i\omega \frac{\alpha}{K_{B1}} p_{f1} = i\omega \frac{\alpha}{K} p_{ext} \quad \text{in} \quad \Omega_i,$$

$$\left. p_{f1} \right|_{\partial\Omega_{12}} = \left. [\mathbf{n} \cdot \nabla p_{f1}] = 0 \right|_{\partial\Omega_{12}},$$

$$\mathbf{n} \cdot \nabla p_{f1} = 0 \quad \text{on} \quad \partial E_i,$$
where $\Omega_i$ is the region that each phase occupies within the averaging volume, $\partial E_i$ is that portion of the external surface of the averaging volume that is in contact with phase $i$, and the brackets in equation (62) again denote jumps across the interface. One also needs to insert equations (3) and (4) into (1) to obtain a second-order partial differential equation for the displacements $u_i$. In general, the local confining pressures $p_{ci}$ are determined using

$$p_{ci} = -K \nabla \cdot u_i + \alpha p_{f1},$$  

(64)

once the displacements $u_i$ are known.

### 3.2.3. Low-frequency limit of $\gamma(\omega)$

As $\omega \to 0$, we can represent the local fields as perturbation expansions in the small parameter $-i\omega$

$$p_{f1} = p_{f1}^{(0)} - i\omega p_{f1}^{(1)} + O(\omega^2),$$  

(65)

$$p_{ci} = p_{ci}^{(0)} - i\omega p_{ci}^{(1)} + O(\omega^2),$$  

(66)

and equivalently for $u_i$. The zeroth-order response corresponds to uniform fluid pressure in the pores and is therefore given by $p_{ci}^{(0)} = p_{f1}^{(0)} = \Delta P$ and

$$\frac{p_{f1}^{(0)}}{\Delta P} = B_0 = -\frac{a_{12} + a_{13}}{a_{22} + 2a_{23} + a_{33}} = \frac{1}{v_1/B_1 + v_2/B_2},$$  

(67)

where the patchy-saturation $a_{ij}$ have been employed. The fact that the quasi-static Smekton's coefficient in the patchy-saturation model is exactly the harmonic average of the constituents $B_i$ is equivalent to saying that at low frequencies, the fluid bulk modulus is given by $1/K_f = v_1/K_{f1} + v_2/K_{f2}$. The quasi-static response is thus completely independent of the spatial geometry of the fluid patches; it depends only on the volume fractions occupied by the patches.

The leading order correction to uniform fluid pressure is then controlled by the boundary-value problem

$$\frac{Kk}{\alpha \eta_1} \nabla^2 p_{f1}^{(1)} = \frac{\eta_2}{\eta_1} \left( 1 - \frac{B_a}{B_2} \right) \Delta P \quad \text{in} \quad \Omega_1,$$  

(68)

$$\frac{Kk}{\alpha \eta_1} \nabla^2 p_{f1}^{(1)} = \frac{1}{B_2} \Delta P \quad \text{on} \quad \partial \Omega_1,$$  

(69)

$$\mathbf{n} \cdot \nabla p_{f1}^{(1)} = \frac{\eta_2}{\eta_1} \mathbf{n} \cdot \nabla p_{f1}^{(1)} \quad \text{on} \quad \partial \Omega_1,$$  

(70)

$$\mathbf{n} \cdot \nabla p_{f1}^{(1)} = 0 \quad \text{on} \quad \partial E_i.$$

(71)

It is now assumed that for patchy-saturation cases of interest (air/water or water/oil), the ratio $\eta_2/\eta_1$ can be considered small. To leading order in $\eta_2/\eta_1$, equations (68), (71), and (72) require that $p_{f1}^{(1)}(r) = \overline{p}_{f1}^{(1)}$ (a spatial constant). The fluid pressure in phase 1 is now rewritten as

$$p_{f1}^{(1)}(r) = \overline{p}_{f1}^{(1)} - \frac{\eta_1 \alpha}{KK} \left( 1 - \frac{B_a}{B_1} \right) \Delta P \Phi_1(r),$$  

(73)

where, from equations (69), (70) and (72) and to leading order in $\eta_2/\eta_1$, the potential $\Phi_1$ is the solution of the same elliptic boundary-value problem (28)–(30) given earlier.

Upon averaging (73) over all of $\Omega_1$, the leading order in $-i\omega$ difference in the average fluid pressures can be written

$$\overline{p}_{f1} - \overline{p}_{f2} = -i\omega \left( \frac{p_{f1}^{(1)} - p_{f2}^{(1)}}{\Delta P} \right) = i\omega \frac{\eta_1 \alpha K}{KK} \left( 1 - \frac{B_a}{B_1} \right) L_1,$$  

(74)

where $L_1$ is again the length defined by equation (27).

To connect this fluid-pressure difference to the increment $\zeta_{int}$, we use the divergence theorem and the no-flow boundary condition on $\partial E_i$ to write equation (58) as

$$-i\omega \zeta_{int} = \frac{i\omega}{V \eta_1} \int_{\partial \Omega_1} \mathbf{n} \cdot \nabla p_{f1}^{(1)} dS = i\omega \frac{\alpha_1 K}{K} \left( 1 - \frac{B_a}{B_1} \right) \Delta P.$$  

(75)

Replacing $\Delta P$ with $\overline{p}_{f1} - \overline{p}_{f2}$ using equation (74) then gives the desired law $-i\omega \zeta_{int} = \gamma(p_{f1} - p_{f2})$ with

$$\gamma_p = \frac{v_1 k}{\eta_1 L_1} \left[ 1 + O \left( \frac{\eta_2}{\eta_1} \right) \right],$$  

(76)

being the low-frequency limit of interest.

### 3.2.4. High-frequency limit of $\gamma(\omega)$

It has already been commented that in the extreme high-frequency limit where each patch behaves as if it were sealed to flow ($\zeta_{int} = 0$), we assume here that, to a very good approximation, the theory of Hill [1963] applies. Hill’s work demonstrates, among other things, that when each isotropic patch has the same shear modulus, the volumetric deformation within each patch is a spatial constant. The fluid pressure response in this limit $p_{f1}^\infty$ is thus a uniform spatial constant throughout each phase except in a vanishingly small neighborhood of the interface $\partial \Omega_{12}$ where equilibration is attempting to take place. The small amount of fluid-pressure penetration that is occurring across $\partial \Omega_{12}$ can be locally modeled as a one-dimensional process normal to the interface.

Using the coordinate $x$ to measure linear distance normal to the interface (and into phase 1), one has that equation (61) is satisfied by [Johnson, 2001]

$$p_{f1} = p_{f1}^\infty + C_1 e^{-i\omega D_1 x},$$  

(77)

$$p_{f2} = p_{f2}^\infty + C_2 e^{-i\omega D_2 x},$$  

(78)

where the diffusivities are defined $D_i = kK B_i/\eta_1 0$. The constants $C_i$ are found from the continuity conditions (62) to be

$$C_1 = \frac{-1}{1 + \sqrt{\eta_2 B_2/(\eta_1 B_1)}} (p_{f1}^\infty - p_{f2}^\infty),$$  

(79)

$$C_2 = \sqrt{\eta_2 B_2/(\eta_1 B_1)} \frac{p_{f1}^\infty - p_{f2}^\infty}{1 + \sqrt{\eta_2 B_2/(\eta_1 B_1)}},$$  

(80)

Although not actually needed here, we have that $p_{f1}^\infty = B_1 p_{ci}$, where the uniform confining pressure of each patch is given by equations (45) and (46), so that the fluid pressure difference between the phases goes as

$$\frac{p_{f1}^\infty - p_{f2}^\infty}{\Delta P} = \frac{B_1 - B_2}{1 - \beta (B_1/v_1 + B_2/v_2)},$$  

(81)

This equation is exactly the difference between equations (55) and (56). Because the penetration distance $\sqrt{D_i/\omega}$ vanishes at high-frequencies, we may state that to leading order in the high-frequency limit, $p_{f1} - p_{f2} = p_{f1}^\infty - p_{f2}^\infty$.

To obtain the high-frequency limit of the transport coefficient $\gamma(\omega)$, we use the definition (58) of the internal transport (note that $-\mathbf{n} \cdot \nabla p_{f1} = \partial p_{f1}/\partial x$)

$$-i\omega \zeta_{int} = \frac{1}{V \eta_1} \int_{\partial \Omega_1} \frac{\partial p_{f1}}{\partial x} dS$$  

(82)

along with equations (77) and (79). The result is

$$\gamma(\omega) \sim \sqrt{|\omega| S \left( \frac{\sqrt{\kappa \eta / (\eta_1 B_1 K)}}{1 + \sqrt{\eta_2 B_2/(\eta_1 B_1)}} \right)}.$$  

(83)
as $\omega \to \infty$. Here, $S$ is again the area of $\partial \Omega_{12}$ contained within a volume $V$ of the patchy composite.

3.2.5. Full-model for $\gamma(\omega)$.

The high- and low-frequency limits of $\gamma$ are then connected by a simple frequency function to obtain the final model

$$\gamma(\omega) = \gamma_p \sqrt{1 - i\omega/\omega_p}, \quad (84)$$

where the transition frequency $\omega_p$ is defined

$$\omega_p = \frac{B_1 K k(v_1 V/S)^2}{\eta_1 \alpha} \left(1 + \frac{\eta_2 B_2}{\eta_1 B_1}\right)^2, \quad (85)$$

and where $\gamma_p = v_1 k/(\eta_1 L_1^2)$. Equation (84) has just one singularity (a branch point) at $\omega = -i\omega_p$. Causality requires that with an $e^{-i\omega t}$ time dependence, all singularities and zeroes of a transport coefficient like $\gamma(\omega)$ must reside in the lower-half complex $\omega$ plane. Equation (84) satisfies this physically important constraint.

3.3. Patchy-Saturation Modeling Choices

To use the patchy-saturation model, appropriate values for the two geometric terms $L_1$ and $V/S$ must be specified. Immiscible fluid distributions in the earth have very complicated geometries since they arise from slow flow that often produces fractal patch distributions. In particular, analytical solutions of the boundary-value problem (28)-(30) that defines $L_1$ for such real-earth situations are impossible. Recall that $L_1$ is a characteristic length of phase 1 (the phase having the smaller fluid mobility $k/\eta_1$) that defines the distance over which the fluid-pressure gradient is defined during the final stages of equilibration. For complicated geometries it may either be numerically determined, treated as a target parameter for a full-waveform inversion of seismic data, or simply estimated qualitatively. In the numerical examples that follow, we will assume (for convenience) that the individual patches correspond to disconnected spheres for which simple analytical results are available for $L_1$ and $V/S$.

If we consider phase 2 (porous continuum saturated by the less viscous fluid) to be in the form of spheres of radius $a$ immersed within each radius $R$ sphere of the two-phase composite, then $v_2 = (a/R)^3$, $V/S = a v_2 /3$, and $L_{1}^2 = 9 v_2 /4 a^2/4[1 - 7v_2 /3]$ (where $v_2$ is taken to be immobile relative to the framework of grains in the wavelength-scale Biot equilibration, so that the inertial properties of equations (34) and (35) are identified as $\rho_f = \rho_{f1}$, $\rho = (1 - \phi)\rho_s + \phi(v_1 \rho_{f1} + v_2 \rho_{f2})$, and $\bar{\rho} = -\eta_1/(i\omega k)$.

In situations where it is more appropriate to treat fluid 1 (the more viscous fluid) as occupying disconnected patches (e.g., when $v_1 \ll v_2$), the effective poroelastic moduli are defined by interchanging 2 and 3 in the subscripts of equations (11)-(13). Again assuming the phase-1 patches to be spheres of radius $a$ embedded within radius $R$ sphere of the two-phase composite, we have that $v_1 = (a/R)^3$ and $V/S = a v_1 /3$. The elliptic boundary-value problem (28)-(30) can be solved in this case to give $L_{1}^2 = a^2/15$. Furthermore, the effective inertial coefficients in the Biot theory are defined $\rho_f = \rho_{f2}$, $\rho = (1 - \phi)\rho_s + \phi(v_1 \rho_{f1} + v_2 \rho_{f2})$, and $\bar{\rho} = -\eta_2/(i\omega k)$.

In situations where both phases form continuous paths across each averaging volume, it is best to determine the attenuation and phase velocity by seeking the plane longitudinal-wave solution of non-reduced “double-porosity” governing equations of the form (6)-(10). However, this approach is not pursued here. We conclude by noting that, if the embedded fluid is fractally distributed, the lengths $L_1$ will remain finite while $(V/S)/L_1 \to 0$ as the fractal surface area $S$ becomes large (however, $V/S$ never reaches zero because the fractality has a small-scale cutoff fixed by the grain size of the material).

3.4. Numerical Examples

In figure 4, we compare the Johnson [2001] prediction of $K_U$ to our own for a consolidated sandstone (frame properties as determined in the appendix with $k = 100$ mD, $c = 10$, $\phi = 0.20$) in which phase 1 is saturated with water and phase 2 is taken to be spherical regions saturated with air. The two estimates have identical asymptotic dependence in both the limits of high and low frequencies. In the cross-over range, the physics is not precisely modeled in either approach. However, even in the cross-over range, the differences in the two models is slight.

Figure 5 gives the P-wave velocity and attenuation for a model in which the frame properties correspond to $k = 10$ mD, $c = 15$, and $\phi = 0.15$. Phase 2 is saturated by air and is taken to be isolated spheres of radius $a = 1$ cm. Phase 1 is saturated with water. The volume fraction $v_2$ occupied by these 1 cm spheres of gas is as shown in the figure. Even tiny amounts of gas saturation yield rather large amounts of attenuation and dispersion; yet these predictions are consistent with the magnitudes of observed attenuation and dispersion in rocks.

4. Squirt-Flow Model

Laboratory samples of consolidated rock often have broken grain contacts and/or microcracks in the grains. Much of this damage presumably occurs as the rock is brought.
from depth to the surface. Since diagenetic processes in a sedimentary basin tend to cement microcracks and grain contacts, it is uncertain whether in situ rocks have significant numbers of open microcracks. Nonetheless, when such grain-scale damage is present, it always is in laboratory rock samples at ambient pressures, the fluid-pressure response in the microcracks will be greater than in the principal porespace when the rock is compressed by a P-wave. The resulting flow from crack to pore is called “squirt flow” [e.g., Matko and Nur, 1973].

In the squirt model of Dvorkin et al. [1995], the grains of a porous material are themselves allowed to have porosity in the form of microcracks. The effect of each broken grain contact is taken as equivalent to a microcrack in a grain. The number of such microcracks per grain is thus limited by the coordination number of the packing and so the total porosity contribution coming from the grains is always negligible compared to the porosity of the main porespace.

The grain space in the Dvorkin et al. [1995] model is taken to be a spatially uniform porous continuum. These authors provide an approximate analysis of their model in which the terms that are left out of the bulk modulus dispersion are as large as the dispersion itself. In the present section, we use the double-porosity framework to analyze the Dvorkin et al. [1995] squirt model with the goal of obtaining exact results at both low and high frequencies. As in the previous two sections, our exact limits are approximately connected by a causal frequency function containing a relaxation frequency appropriate for a grain space of arbitrary geometry.

Phase 1 is now defined to be the pure fluid within the main porespace of a sample and is characterized elastically by the single modulus $K_f$ (fluid bulk modulus). Phase 2 is taken to be the porous (i.e., cracked) grains and characterized by the poroelastic constants $K_g$ (the drained modulus of an isolated porous grain), $\alpha_2$ (the Biot-Willis constant of an isolated grain), and $B_2$ (Skempton’s coefficient of an isolated grain) as well as by a permeability $k_2$. The overall composite of porous grains (phase 2) packed together

within the fluid (phase 1) has two distinct properties of its own that must be specified: an overall drained modulus $K$, and an overall permeability $k$ associated with flow through the main porespace. The volume fractions occupied by each phase are again denoted $v_i$, where $v_i = \phi$ is the porosity associated with the main porespace.

The theoretical approach is to obtain again the average fluid response in each of these two phases and then to make an effective Biot theory by saying that the fluid within the grains cannot communicate directly with the outside world, i.e., the fluid in the grains can only communicate with the main pores. Equations (11)-(13) again define the effective poroelastic moduli in the squirt model and we need only determine the $a_{ij}$ constants and internal transport coefficient $\gamma(\omega)$ that are appropriate to squirt.

### 4.1. Squirt $a_{ij}$ Coefficients

To obtain the $a_{ij}$ coefficients in the squirt model, we first note that these coefficients are defined under conditions where $Q_{int} = 0$ (no fluid passing between the porous grains and the principal pore space). Under these conditions, the rate of fluid depletion $\nabla \cdot q$, of a sample (rate of fluid volume being extruded from the principal porespace via the exterior sample surface as normalized by the sample volume) is due to the difference between the rate of dilatation of the principal porespace (denoted here as $\dot{e}_1$) and the rate at which fluid in the pores is diluting $-\frac{1}{K_f}$. If we also perform a volume average of equation (3) over the porous grain space and use the notation that $v_2\dot{e}_2 = \nabla \cdot (v_2\mathbf{u}_2)$ we obtain the following three equations

$$-\nabla \cdot q_1 = v_1 \dot{e}_1 + \frac{v_1}{K_f} \dot{p}_f$$

$$-\nabla \cdot q_2 = -\frac{v_2 \alpha_2}{K_g} \dot{p}_f + \frac{v_2 \alpha_2}{B_2 K_g} \dot{p}_f$$

$$-v_2 \dot{e}_2 = \frac{v_2}{K_g} \dot{p}_2 - \frac{v_2 \alpha_2}{K_g} \dot{p}_f.$$  \hfill (86)

$$-\nabla \cdot v = v_1 \dot{e}_1 + v_2 \dot{e}_2$$  \hfill (87)

$$-\nabla \cdot v = \nabla \cdot (v_1 \dot{p}_f + v_2 \dot{p}_2)$$  \hfill (88)

The macroscopic dilatation of interest is $\nabla \cdot v = v_1 \dot{e}_1 + v_2 \dot{e}_2$. In order to obtain the macroscopic compressibility laws for the porous-grain/principal-porespace composite, we introduce linear response laws of the form

$$\dot{p}_{c2} = a_1 \dot{p}_c + a_2 \dot{p}_f + a_3 \dot{p}_{f2}$$  \hfill (89)

$$\dot{e}_1 = b_1 \dot{p}_c + b_2 \dot{p}_f + b_3 \dot{p}_{f2}$$  \hfill (90)

where the $a_i$ and $b_i$ must be found. We note immediately that from the definition $\overline{\dot{p}_c} = v_1 \overline{\dot{p}_f} + v_2 \overline{\dot{p}_{f2}}$ one has

$$0 = (1 - v_2 a_1) \dot{p}_c - (v_1 + v_2 a_2) \overline{\dot{p}_f} - v_2 a_3 \overline{\dot{p}_{f2}},$$  \hfill (91)

which must hold true for any variation of the independent pressure variables so that $a_1 = 1/v_1$, $a_2 = -v_1/v_2$, $a_3 = 0$.

To obtain the $b_i$ coefficients, we now combine the results above into macroscopic laws

$$-\nabla \cdot v = \left[ -v_1 b_1 + \frac{1}{K_f} \right] \dot{p}_c,$$

$$-v_1 b_2 + \frac{v_1}{K_f} \dot{p}_f + \left[ v_1 b_3 + \frac{v_2 \alpha_2}{K_g} \right] \overline{\dot{p}_f}.$$  \hfill (92)

$$-\nabla \cdot q_1 = v_1 \dot{p}_c + \frac{v_1}{K_f} \overline{\dot{p}_f} + v_1 b_2 \dot{p}_f$$  \hfill (93)

$$-\nabla \cdot q_2 = -\frac{v_2 \alpha_2}{K_g} \dot{p}_c + \frac{v_2 \alpha_2}{K_g} \overline{\dot{p}_f} + \frac{v_2 \alpha_2}{B_2 K_g} \dot{p}_f,$$  \hfill (94)

and use the fact that the coefficients of the matrix must be symmetric ($a_{ij} = a_{ji}$). With $a_{11} = 1/K$ corresponding to
the overall drained frame modulus of the composite (to be independently specified), we obtain \( v_1 b_1 = -(1/ K - 1/K_2^d) \), \( v_1 b_2 = 1/K - (1 + v_1)/K_2^d \), and \( b_3 = \alpha_2/K_2^d \). The final \( a_{ij} \) coefficients are exactly
\[
\begin{align*}
  a_{11} &= 1/K \quad (95) \\
  a_{22} &= 1/K - (1 + v_1)/K_2^d + v_1/K_f \quad (96) \\
  a_{33} &= v_2 \alpha_2 / B_2 K_2^d \quad (97) \\
  a_{12} &= -1/K + 1/K_2^d \quad (98) \\
  a_{13} &= -\alpha_2/K_2^d \quad (99) \\
  a_{23} &= v_1 \alpha_2 / K_2^d. \quad (100)
\end{align*}
\]
Reasonable models for \( K \) and \( K_2^d \) will be discussed shortly.

### 4.2. Squirt Transport

We next must obtain the coefficient \( \gamma(\omega) \) in the mesoscopic transport law \(-\omega \zeta_{int} = \gamma(\omega) (\mathbf{p}_{f1} - \mathbf{p}_{f2}) \). Again, the approach is to first obtain the limiting behaviour at low and high frequencies and then to connect the two limits by a simple function.

The fluid response in phase 1 (the principal pore space) is governed by the Navier-Stokes equation \(-\nabla p_{f1} + \eta \nabla^2 v_1 = -i \omega \rho_{f1} \) and the compressibility law \( K_f \nabla \cdot v_1 = i \omega p_{f1} \) where \( v_1 \) is the local fluid velocity in the pores. Since for all frequencies of interest we have that \( \omega \ll K_f/\eta \) (note that \( K_f/\eta \approx 10^{12} \) s\(^{-1} \) for liquids and \( 10^{10} \) s\(^{-1} \) for gases), the fluid pressure in phase 1 is governed by the wave equation
\[
\nabla^2 p_{f1} + \omega^2 \frac{\rho_f}{K_f} p_{f1} = 0, \quad (101)
\]
and, since the acoustic wavelength in the fluid is always much greater than the grain size, the fluid pressure in the principal pore space satisfies \( p_{f1}(r) = \overline{p}_{f1} \) (a spatial constant) at all frequencies.

The focus, then, is on determining how the fluid pressure within the cracked grains (phase 2) is governed by the local porous-continuum laws \( Q_2 = -(k_2/\gamma) \nabla p_{f2} \) and
\[
\frac{k_2}{\eta} \nabla^2 p_{f2} + i \omega \frac{\alpha_2}{K_2^d K_2^d} p_{f2} = -i \omega \frac{\alpha_2}{K_2^d} p_{f2}, \quad (102)
\]
where \( p_{f2} = -K_2^d \nabla \cdot v_2 + \alpha_2 p_{f2} \). This deformation and pressure change is excited by applying a uniform normal stress \(-\Delta P n\) to the surface of the averaging volume with the fluid pressure satisfying the boundary conditions \( \mathbf{n} \cdot \nabla p_{f2}(r) = 0 \) on \( \partial E_2 \) and \( p_{f2}(r) = \overline{p}_{f2} \) on \( \partial \Omega_{12} \).

#### 4.2.1. Low-frequency limit of \( \gamma(\omega) \)

The fluid pressure and confining pressure in the grains can again be developed as a power series in \(-\omega I\) as in equations (65)-(66). The zero-order response corresponds to the static limit in which the fluid pressure is everywhere the same and given by \( p_{f2}(0) = \overline{p}_{f2} = B_2 \Delta P \) with \( B_2 = (\alpha_1 + \alpha_2)/(2 \alpha_2 + 2 \alpha_3 + 3 \alpha_3) \) and with the \( \alpha_i \) given by equations (95)-(100). The detailed result for \( B_2 \) can be expressed
\[
\frac{1/K - (1 - \alpha_2)/K_2^d}{B_2} = \frac{1}{K} - \frac{1 - \alpha_2}{K_2^d} + v_1 \left[ \frac{1}{K_f} - \frac{(1 - \alpha_2)}{K_2^d} \right] + \frac{v_2}{B_2} \frac{\alpha_2}{K_2^d} \left[ \frac{1}{B_2} - 1 \right], \quad (103)
\]
which reduces to the standard Gassmann expression given in the appendix (with a total porosity given by \( v_1 + \alpha_2 v_2 \)), when \( B_2 \) and \( \alpha_2 \) are themselves given by the Gassmann expressions. In this same zero-order limit, the undrained bulk modulus is defined as \( 1/K_0 = a_{11} + (a_{12} + a_{13}) B_0 \), which also reduces to the standard Gassmann expression, when \( B_2 \) and \( \alpha_2 \) are themselves given by Gassmann expressions.

The correction to uniform flow pressure that is leading order in \(-\omega I\) is thus governed by the problem
\[
\nabla^2 p_{f2}^{(1)} = \frac{\eta \alpha_2}{k_2 K_2^d} \overline{p}_{f2}^{(0)}, \quad (104)
\]
\[
\nabla \cdot \nabla p_{f2}^{(1)} = 0 \quad \text{on} \quad \partial E_2, \quad (105)
\]
\[
\overline{p}_{f2}^{(1)} = 0 \quad \text{on} \quad \partial \Omega_{12}. \quad (106)
\]
Here, \( p_{f2}^{(0)} \) is the local confining pressure in the grain space in the static limit that can be written \( p_{f2}^{(0)}(r) = \overline{p}_{f2}^{(0)} + \delta P(r) \). The average static confining pressure throughout the grains is determined from equation (84) with \( P = \Delta P \) and \( p_{f2} = p_{f2}^{(1)} = B_0 \Delta P \) to yield
\[
\overline{p}_{f2}^{(0)} = \frac{(1 - v_1 B_0)}{v_2 B_2 K_2^d} \Delta P. \quad (107)
\]
The deviations \( \delta P(r) \) thus integrate by volume to zero \( \Delta P = 0 \) and are formally defined
\[
\delta P(r) = -\left( \frac{1 - (v_1 + v_2 \alpha_2) B_0}{v_2} \right) \Delta P - \frac{K_2^d}{\alpha_2} \nabla \cdot u^{(0)}(r). \quad (108)
\]
The local perturbations \( \delta P(r) \) are thus highly sensitive to the detailed nature of the grain packing and grain geometry. Fortunately, the details of these perturbations do not play an important role in the theory.

The fluid pressure in the grains is now written in the scaled form
\[
p_{f2}^{(1)}(r) = -\frac{\eta \alpha_2 (1 - v_1 B_0)}{v_2 B_2 K_2^d} \Delta P \Phi(r), \quad (109)
\]
where the potential \( \Phi(r) \) is independent of \( \Delta P \) and is a solution of the elliptic problem
\[
\nabla^2 \Phi = -1 - \frac{v_2}{1 - v_1 B_0} \frac{\partial \Phi}{\partial E_2}, \quad (110)
\]
\[
\nabla \cdot \Phi = 0 \quad \text{on} \quad \partial E_2, \quad (111)
\]
\[
\Phi = 0 \quad \text{on} \quad \partial \Omega_{12}. \quad (112)
\]
To leading order in \(-\omega I\), an average of equation (109) gives
\[
\overline{p}_{f2} - \overline{p}_{f2}^{(1)} = i \omega \left( \frac{\partial P_0}{\partial E_2} + O(\omega^2) \right) = -i \omega \frac{\eta \alpha_2 (1 - v_1 B_0)}{v_2 K_2^d B_2} L_2^d \Delta P + O(\omega^2), \quad (113)
\]
where the squared length \( L_2^d \) is defined
\[
L_2^d = \frac{D}{\overline{P}_0} \left[ 1 + \frac{v_2}{1 - v_1 B_0} \frac{\overline{P}_0 \Delta P}{\Phi \Delta P} \right], \quad (115)
\]
with overbars denoting volume averages over the grain space and with the potential \( \Phi \) defined as the solution of
\[
\nabla^2 \Phi = -1, \quad (116)
\]
\[
\nabla \cdot \Phi = 0 \quad \text{on} \quad \partial E_2, \quad (117)
\]
\[
\Phi = 0 \quad \text{on} \quad \partial \Omega_{12}. \quad (118)
\]
Although it is not generally true that \( \Phi \Delta P = 0 \) for all grain geometries, we nevertheless expect this integral to be small in general because \( \Phi \) is a smooth function and \( \Delta P = 0 \). The local perturbations in the static confining pressure \( \delta P(r) \) require a solution of the static displacements throughout the
entire grain space—a daunting numerical task. Whenever the length $L_2$ needs to be estimated, such as in the numerical results that follow, our approach is simply to use the reasonable approximation that $L_2 = \bar{D}_n$.

Finally, from the definition $\zeta_{\text{int}}$ of the internal transfer we have that to leading order in $-i\omega$

$$-i\omega\zeta_{\text{int}} = \frac{i\omega k_2}{V\eta} \int_{\partial\Omega_{12}} \mathbf{n} \cdot \nabla \tau^{(1)}_{f2} \nabla^{(2)}_{f2}$$

(119)

$$= -\frac{i\omega k_2}{V\eta} \int_{\partial\Omega_{12}} \nabla \tau^{(1)}_{f2} \nabla^{(2)}_{f2} = -i\omega \frac{\alpha_2}{K_2} v_2 \tau^{(1)}_{f2}$$

(120)

$$= \frac{v_2 k_2}{\eta L_2^2} (\tau_{f1} - \tau_{f2}).$$

(121)

The normal $\eta$ in equation (119) is outward to phase 1 which accounts for the sign change in equation (120). Note as well that equation (120) is a volume average of equation (104) while equation (121) follows from equations (107) and (114). The desired result is thus $\lim_{\omega \to 0} \gamma(\omega) = \gamma_{\text{eq}} = v_2 k_2 / (\eta L_2^2)$.

4.2.2. High-frequency limit of $\gamma(\omega)$.

In the extreme high-frequency limit, the fluid has no time to escape in significant amounts from the porous grains (phase 2) and enter the main pore space (phase 1). As such, the fluid pressure distribution in each phase is reasonably modeled as

$$p_{f1}(r) = B_{f1}^{\infty} \Delta P,$$

(122)

$$p_{f2}(r) = B_{f2}^{\infty} \Delta P + C_2 \Delta P e^{-\gamma \eta \sqrt{\omega\eta / \eta_2}}$$

(123)

where $x$ is again a local coordinate measuring distance normal to the interface $\partial\Omega_{12}$, and where $D_2$ is the fluid-pressure diffusivity within the porous grains that is given by $D_2 = k_2 K_2^d B_2 / (\eta_2)$. In reality, the local confining pressure $p_{c2}(r)$ throughout the grains has spatial fluctuations about the average value and we have made the approximation that the average fluid pressure throughout the grain space is $B_2 p_{c2}(r) \approx B_2^{\infty} \Delta P$. It is easy to demonstrate that under undrained and unrelaxed conditions,

$$B_{f1}^{\infty} = \frac{a_{12} \eta_{12}}{a_{22} \eta_{22}} - \frac{a_{31} \eta_{12}}{a_{33} \eta_{22}} - \frac{a_{31} \eta_{12}}{a_{33} \eta_{22}},$$

(124)

$$B_{f2}^{\infty} = \frac{a_{12} \eta_{12}}{a_{22} \eta_{22}} - \frac{a_{21} \eta_{12}}{a_{22} \eta_{22}} - \frac{a_{31} \eta_{12}}{a_{33} \eta_{22}}.$$  (125)

However, since these $B_{f1}^{\infty}$ do not appear in the final result, they will not be algebraically developed.

The continuity of fluid pressure $p_{f2} = p_{f1}$ along $\partial\Omega_{12}$ ($x = 0$) requires that $C_2 = B_{f1}^{\infty} - B_{f2}^{\infty}$. The definition of $\zeta_{\text{int}}$ may now be used to write

$$-i\omega\zeta_{\text{int}} = \frac{1}{V} \int_{\partial\Omega_{12}} k_2 \frac{\partial p_{f2}}{\partial x}$$

(126)

$$= \frac{k_2}{\eta} \sqrt{\frac{\omega}{D_2 V}} (B_{f1}^{\infty} - B_{f2}^{\infty}) \Delta P$$

(127)

$$= i^{3/2} \sqrt{\omega} \frac{k_2 \alpha_2}{\eta B_2 K_2^d} \frac{S}{V} (\tau_{f1} - \tau_{f2}).$$

(128)

where we have used, to leading order in the high-frequency limit, that $\tau_{f1} - \tau_{f2} = (B_{f1}^{\infty} - B_{f2}^{\infty}) \Delta P$. The desired result is then

$$\gamma(\omega) \sim \frac{S}{V} \sqrt{-i\omega k_2 \alpha_2}{\eta B_2 K_2^d}$$

(129)

as $\omega \to \infty$.

4.2.3. Full model for $\gamma(\omega)$.

The high- and low-frequency limits are again causally connected via the simple function

$$\gamma(\omega) = \gamma_{\text{eq}} \sqrt{1 - \frac{i\omega}{\omega_{\text{eq}}}}$$

(130)

but now the parameters are defined as

$$\gamma_{\text{eq}} = \frac{v_2 k_2}{\eta L_2^2}$$

(131)

$$\omega_{\text{eq}} = \frac{B_2 K_2^d}{\eta_2} \frac{k_2}{L_2^2} (v_2 S / L_2)^2.$$  (132)

4.3. Squirt-Flow Modeling Choices

To make numerical predictions of attenuation and dispersion, models must be proposed for the phase 2 (porous grain) parameters.

If the grains are modeled as spheres of radius $R$, the fluid-pressure gradient length within the grains can be estimated as $L_2 = R / \sqrt{7} \pi$ and the volume to surface ratio as $V / S = R / (3 \sqrt{2} \pi)$. The grain porosity is assumed to be in the form of microcracks and so it is natural to define an effective aperture $h$ for these cracks. If the cracks have an average effective radius of $R / N_R$ (where $N_R$ is roughly 2 or 3), and if there are on average $N_c$ cracks per grain (where $N_c$ is also roughly 2 or 3), then the permeability and porosity of the grains are reasonably modeled as

$$\phi_2 = \frac{3N_c h}{4N_R R}$$

(133)

and $k_2 = \phi_2 h^2 / 12$.

where $\phi_2$ is the fracture porosity within the porous grains. The dimensionless parameters $k_2 / L_2^d$ and $(v_2 S / L_2)$ required in the expressions for $\gamma_{\text{eq}}$ and $\omega_{\text{eq}}$ are now given by

$$k_2 = \frac{15N_c}{16N_R R} \left( \frac{h}{R} \right)^3$$

(134)

and $(v_2 S / L_2)^2 = \frac{5}{3}$. 

The normalized fracture aperture $h / R$ is the key parameter in the squirt model.

The drained grain modulus $K_2^d$ is necessarily a function of the crack porosity $\phi_2$ (and therefore $h / R$). Real crack surfaces have micron (and smaller) scale asperities present upon them. If effective stress is applied in order to make the normalized aperture $h / R$ smaller (so that, for example, the peak in squirt attenuation lies in the seismic band), new contacts are created that make the crack stronger. In the limit as $h / R \to 0$ (large effective stress), the cracks are no longer present and $K_2^d \to K_0$, where $K_0$ is the mineral modulus of the grain.

Many models for such stiffening could be proposed. We intentionally make a conservative estimate here in proposing a simple linear porosity dependence $K_2^d = K_0 (1 - \sigma \phi_2)$ where $\sigma$ is a fixed constant determined from fitting ultrasonic attenuation data. Effective medium theories (see, for example, Berryman et al. [2002]) predict that $\sigma$ should be inversely proportional to the aspect ratios of the cracks present. As a crack closes and asperities are brought into contact, there is naturally a decrease in $\phi_2$ but there should also be a decrease in $\sigma$ due to the fact that the remaining crack porosity becomes more equant as new asperities come into contact. Taking $\sigma$ to be constant as crack porosity decreases is thus a minimalist estimate for how the drained modulus increases.
4.4. Numerical Examples

Thus, the porous-grain elastic properties are taken to be

\[
\begin{align*}
K^d_2 &= K_s (1 - \sigma \phi_2) \quad (135) \\
\alpha_2 &= 1 - K^d_2 / K_s \quad (136) \\
\frac{1}{B_2} &= 1 + \phi_2 \frac{K^d_2}{K_f} \left(1 - K_f / K_s\right) \quad (137)
\end{align*}
\]

where we have used the Gassmann fluid-substitution relations for \( \alpha_2 \) and \( B_2 \). The overall drained modulus \( K \) of the collection of porous (cracked) grains can be modeled for example as

\[
K = \frac{K^d_2 (1 - \phi_1)}{1 + c \phi_2}
\]

which is the same drained-modulus model as given in the appendix but with the solid grain modulus \( K_s \) replaced by the cracked grain modulus \( K^d_2 \).

4.4. Numerical Examples

In figure 6, we plot the P-wave attenuation predicted using the above model when the overall grain packing corresponds to a consolidated sandstone \((\phi_1 = 0.2, c = 5)\) having a permeability of 10 mD. For the grain properties, we take \( \sigma = 0.8/(5 \times 10^{-3}) \), \( 3 N_c/(4 N^2_d) = 1 \), and \( K_s = 38 \) GPa (quartz) as fixed constants. This \( \sigma \) value was chosen so that there would be a significant peak in attenuation at ultrasonic frequencies and is taken to be the same for all values of \( h/R \). The various curves can be thought of as being due to the application of effective stress. The peak in \( Q^{-1} \) near 1 MHz that is invariant to \( h/R \) is the one due to the macroscopic Biot loss (fluid pressure equilibration at the scale of the wavelength). The peak that shifts with \( h/R \) is the one due to the squirt flow.

This figure indicates that, although the squirt mechanism is probably operative and perhaps even dominant at ultrasonic frequencies, it does not seem to be involved in explaining the observed levels of intrinsic attenuation in exploration work. For real cracks inside of real grains, the \( \sigma \) value will diminish with effective stress (i.e., with \( h/R \)), so that the effects of squirt in the seismic band are likely to be even less than shown in figure 6.

![Figure 6](image-url)

**Figure 6.** The squirt-flow model of P-wave attenuation when the grains are modeled as being spherical with radius \( R \) and containing microcracks having effective apertures \( h \). The overall drained modulus of the rock corresponds to a consolidated sandstone.

![Figure 7](image-url)

**Figure 7.** The dispersion in the real parts of the drained bulk modulus \( K_D(\omega) \) [top graph], the undrained bulk modulus \( K_U(\omega) \) [middle graph], and Skempton's coefficient \( B(\omega) \) [bottom graph], as determined both in the present study and by Dvorkin et al. [1995]. The plots were all generated with \( h/R = 5 \times 10^{-3} \). Both theories use identically the same input parameters and are treating identically the same model. The present study may be considered exact in both the low and high frequency limits of the model.

We next introduce the grain parameters \( k_2, \phi_2, \) and \( K^d_2 \) as modeled here along with the same overall drained modulus \( K \) into the equations of Dvorkin et al. [1995] and compare their results to our own when \( h/R = 5 \times 10^{-3} \) (figure 7). Dvorkin et al. [1995] have made a series of approximations in their analysis [starting with equation (3) in their paper] in which the error introduced is often as large as the dispersion being modeled. Figure 7 quantifies this error since our analysis of their model, at least in the limits of both low and high frequencies, is exact.

5. Conclusions

Models for three different P-wave attenuation mechanisms were derived using a single theoretical framework. The resulting models differ only in the values of the \( a_1 \) constants and in the values of the parameters contributing to the mesoscopic-transport coefficient \( \gamma(\omega) \). These three models correspond to (1) mesoscopic-scale heterogeneity in the frame moduli or “double porosity”, (2) mesoscopic-scale heterogeneity in the fluid type or “patchy-saturation”, and (3) grain-scale heterogeneity due to microcracks in the grains or “squit”. In all three models, the amount of attenuation is controlled principally by the contrast of elastic compressibility among the constituents along with the assumed mesoscopic geometry. In the double-porosity model, it is necessary that the embedded phase have an elongated or squashed
form and that the contrast between the frame bulk-modulus of the two porous phases is strong in order for the mesoscopic loss to be significant. In the patchy-saturation model, the contrast in the fluid bulk modulus must be strong (immiscible patches of different fluids that have nearly identical bulk moduli would not produce much attenuation), while in the squirt model, it is the contrast between the drained modulus of an isolated cracked grain and that of the entire packing of grains that controls the amount of attenuation.

Putting in thin lenses of unconsolidated sand grains into an otherwise consolidated sandstone can produce attenuation in the seismic band that is comparable to what is measured in the field even when the embedded phase represents only a small amount of the total volume ( \( < 1\% \) volume fractions). Such a model might correspond to a jointed sandstone. Since mesoscopic-scale heterogeneity is rather ubiquitous throughout the earth’s crust, it seems reasonable to suppose that this mechanism may be responsible for most of the attenuation observed in seismograms. The squirt mechanism produces a great deal of attenuation at the ultrasonic frequencies used in laboratory measurements, but has trouble explaining attenuation in the seismic band. This result is important for some applications of the theory because the rate at which the mesoscopic-scale fluid-pressure equilibrates is a strong function of the permeability of the porous material. The rate at which microcracks equilibrate with the main pores in squirt flow is not permeability dependent. This leaves open the possibility of extracting permeability information from the frequency dependence of seismically measured \( Q \).

**Appendix A: Constituent Properties**

In order to use the unified double-porosity framework of the present paper, it is convenient to have models for the various porous-continuum constituent properties.

For unconsolidated sands and soils, the frame moduli (drained bulk modulus \( K^d \) and shear modulus \( G \)) are well modeled using the following variant of the Walton [1987] theory [c.f., Pride, 2003 for details]

\[
K^d = \frac{1}{6} \left[ \frac{4(1 - \phi)^2 n_e^2 P_o}{\pi^2\rho_o^2} \right]^{1/3} \left( \frac{P_o}{P_o} \right)^{1/2} \frac{1}{\{1 + [16P_o/(9P_o)]^{1/4}\}} \right)^{2/3} \tag{A1}
\]

\[
G = 3K^d/5, \tag{A2}
\]

where \( P_o \) is the effective overburden pressure [e.g., \( P_o = (1 - \phi)(\rho_o - \rho_f)gh \), where \( g \) is gravity and \( h \) is overburden thickness] and where \( P_o \) is the effective pressure at which all grain-to-grain contacts are established. For \( P_o < P_o \), the coordination number \( n \) (average number of grain contacts per grain) is increasing as \( (P_o/P_o)^{1/2} \). For \( P_o > P_o \), the coordination number remains constant \( n = n_o \). The parameter \( P_o \) is commonly on the order of 10 MPa. As \( P_o \to 0 \), the Walton [1987] result is obtained (all contacts in place starting from \( P_o = 0 \)). The porosity of the grain pack is \( \phi_o \) and the compliance parameter \( C_s \) is defined

\[
C_s = \frac{1}{4\pi} \left( \frac{1}{G_s} + \frac{1}{K_s + G_s/3} \right). \tag{A3}
\]

where \( K_s \) and \( G_s \) are the mineral moduli of the grains. For unimodal grain-size distributions and random grain packs, one typically has \( 0.32 < \phi_o < 0.36 \) and \( 8 < n_o < 11 \).

For consolidated sandstones, the frame moduli are modelled in the present paper as (c.f., Pride [2003] for details)

\[
K^d = K_s \frac{1 - \phi}{1 + c\phi} \tag{A4}
\]

\[
G = G_s \frac{1 - \phi}{1 + 3c\phi/2} \tag{A5}
\]

The consolidation parameter \( c \) represents the degree of consolidation between the grains and lies in the approximate range \( 2 < c < 20 \) for sandstones. If it is necessary to use a \( c \) greater than say 20 or 30, then it is probably better to use the modified-Walton theory.

The undrained moduli \( K^u \) and \( B \) are conveniently and exactly modeled using the Gassmann [1951] theory whenever the grains are isotropic and composed of a single mineral. The results are

\[
B = \frac{1/K^d - 1/K_s}{1/K^d - 1/K_s + \phi(1/K^d - 1/K_s)} \tag{A6}
\]

\[
K^u = \frac{K^d}{1 - B(1 - K^d/K_s)} \tag{A7}
\]

from which the Brod-Willis constant \( \alpha \) may be determined to be \( \alpha = 1 - K^d/K_s \). These Gassmann results are often called the “fluid-substitution” formulas.

The dynamic permeability \( k(\omega) \) as modeled by Johnson et al. [1987] is

\[
k(\omega) = \left[ \frac{1}{1 - \frac{4}{\pi} \frac{\omega}{\omega_c} - \frac{\omega}{\omega_c}} \right]^{-1}, \tag{A8}
\]

where the relaxation frequency \( \omega_c \), which controls the frequency at which viscous-boundary layers first develop, is given by

\[
\omega_c = \frac{\eta}{\rho_f K_s}. \tag{A9}
\]

Here, \( F \) is exactly the electrical formation factor when grain-surface electrical conduction is not important and is conveniently (though crudely) modeled using Archie’s law \( F = \phi^{-m} \) [Archie, 1942]. The cementation exponent \( m \) is related to the distribution of grain shapes (or pore topology) in the sample and is generally close to 3/2 in clean sands, close to 2 in shaly sands, and close to 1 in rocks having fracture porosity (indeed, a reasonable model is \( m = 3/2 + 1/c \)). The parameter \( n_j \) is, for convenience, taken to be 8 (cylinder model of the porespace).

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