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1 Introduction to rock physics

Make your theory as simple as possible, but no simpler.

Albert Einstein

1.1 Introduction

The sensitivity of seismic velocities to critical reservoir parameters, such as porosity, lithofacies, pore fluid type, saturation, and pore pressure, has been recognized for many years. However, the practical need to quantify seismic-to-rock-property transforms and their uncertainties has become most critical over the past decade, with the enormous improvement in seismic acquisition and processing and the need to interpret amplitudes for hydrocarbon detection, reservoir characterization, and reservoir monitoring. Discovering and understanding the seismic-to-reservoir relations has been the focus of rock physics research.

One of our favorite examples of the need for rock physics is shown in Plate 1.1. It is a seismic P–P reflectivity map over a submarine fan, or turbidite system. We can begin to interpret the image without using much rock physics, because of the striking and recognizable shape of the feature. A sedimentologist would tell us that the main feeder channel (indicated by the high amplitude) on the left third of the image is likely to be massive, clean, well-sorted sand – good reservoir rock. It is likely to be cutting through shale, shown by the low amplitudes. So we might propose that high amplitudes correspond to good sands, while the low amplitudes are shales.

Downflow in the lobe environment, however, the story changes. Well control tells us that on the right side of the image, the low amplitudes correspond to *both* shale and clean sand – the sands are transparent. In this part of the image the bright spots are the poor, shale-rich sands. So, what is going on?

We now understand many of these results in terms of the interplay of sedimentologic and diagenetic influences. The clean sands on the left (Plate 1.1) are very slightly cemented, causing them to have higher acoustic impedance than the shales. The clean sands on the right are uncemented, and therefore have virtually the same impedance as the shales. However, on the right, there are more facies associated with lower energy

2 Introduction to rock physics

deposition, and these tend to be more poorly sorted and clay-rich. We know from laboratory work and theory that poor sorting can also influence impedance. In the turbidite system in Plate 1.1 both the clean, slightly cemented sand and the clean uncemented sand are oil-saturated. These sands have essentially the same porosity and composition, yet they have very different seismic signatures.

This example illustrates the need to incorporate rock physics principles into seismic interpretation, and reservoir geophysics in general. Despite the excellent seismic quality and well control, the correct interpretation required quantifying the connection between geology and seismic data. A purely correlational approach, for instance using neural networks or geostatistics, would not have been so successful.

Our goal in this first chapter is to review some of the basic rock physics concepts that are critical for reservoir geophysics. Although the discussion is not exhaustive, we assess the strengths, weaknesses, and common pitfalls of some currently used methods, and we make specific recommendations for seismic-to-rock-property transforms for mapping of lithology, porosity, and fluids. Several of these rock physics methods are further discussed and applied in Chapters 2, 3, and 5.

1.2 Velocity–porosity relations for mapping porosity and facies

Rock physics models that relate velocity and impedance to porosity and mineralogy (e.g. shale content) form a critical part of seismic analysis for porosity and lithofacies. In this section we illustrate how to recognize the appropriate velocity–porosity relation when approaching a new reservoir geophysics problem.

Pitfall

One of the most serious and common mistakes that we have observed in industry practice is the use of inappropriate velocity–porosity relations for seismic mapping of porosity and lithofacies. The most common error is to use overly stiff velocity–porosity relations, such as the classical empirical trends of Wyllie *et al.* (1956), Raymer, Hunt, and Gardner (Raymer *et al.*, 1980), Han (1986), or Raiga-Clemenceau *et al.* (1988), the critical porosity model (Nur, 1992), or penny-shaped crack models. "Sonic porosity," derived from sonic logs using the Wyllie time average, is perhaps the worst example. Implicit in these relations is that porosity is controlled by variations in diagenesis, which is not always the case. Hence, critical sedimentologic variations are ignored.

3 1.2 Velocity–porosity relations for mapping porosity and facies

Solution

Rock physics diagnostic analysis of well logs and cores, coupled to the geologic model, usually leads to more rational velocity–porosity relations. Certain aspects are highlighted in this section.

The importance of velocity–porosity relations applies to other rock physics problems, as well. Even seismic pore fluid analysis, which we discuss in the next section, depends on the velocity–porosity relation. We can start to see this by looking at the Gassmann (1951) relation, which can be represented in the form (Zimmerman, 1991; Mavko and Mukerji, 1995; Mavko *et al.*, 1998):

$$\frac{1}{K_{\rm rock}} = \frac{1}{K_{\rm mineral}} + \frac{\phi}{\tilde{K}_{\phi}}$$

where K_{rock} , K_{mineral} , and \tilde{K}_{ϕ} are the bulk moduli of the saturated rock, the mineral, and the saturated pore space, respectively, and ϕ is the porosity. The pore space modulus is approximately the sum of the dry pore modulus and the fluid modulus: $\tilde{K}_{\phi} \approx K_{\phi} + K_{\text{fluid}}$. (We will define these more carefully later.) Hence, we can see that the sensitivity of rock modulus (and velocity) to pore fluid changes depends directly on the ratio of pore space stiffness to porosity, K_{ϕ}/ϕ . Rocks that are relatively stiff have a small seismic sensitivity to pore fluids, and rocks that are soft have a large sensitivity to pore fluids.

We encounter the link between fluid substitution and velocity-porosity relations in several common ways:

- When first analyzing well logs to derive a velocity–porosity relation, it is essential first to map the data to a common fluid. Otherwise, the effects of the rock frame and pore fluid become mixed.
- When interpreting 3D seismic data for hydrocarbon detection, the Gassmann analysis requires a good estimate of porosity, which also must be mapped from the seismic data.
- When populating reservoir models with acoustic properties (V_P and V_S) for 4D feasibility studies, we often need to map from porosity to velocity. Beginning the exercise with the incorrect mapping quickly makes the fluid substitution analysis wrong.

1.2.1 Background on elastic bounds

We begin with a discussion of upper and lower bounds on the elastic moduli of rocks. The bounds provide a useful and elegant framework for velocity–porosity relations.

Many "effective-medium" models have been published, attempting to describe theoretically the effective elastic moduli of rocks and sediments. (For a review, see





Figure 1.2 Conceptual illustration of bounds for the effective elastic bulk modulus of a mixture of two materials.

Mavko *et al.*, 1998.) Some models approximate the rock as an elastic block of mineral perturbed by holes. These are often referred to as "inclusion models." Others try to describe the behavior of the separate elastic grains in contact. These are sometimes called "granular-medium models" or "contact models." Regardless of the approach, the models generally need to specify three types of information: (1) the volume fractions of the various constituents, (2) the elastic moduli of the various phases, and (3) the geometric details of how the phases are arranged relative to each other.

In practice, the geometric details of the rock and sediment have never been adequately incorporated into a theoretical model. Attempts always lead to approximations and simplifications, some better than others.

When we specify only the volume fractions of the constituents and their elastic moduli, without geometric details of their arrangement, then we can predict only the upper and lower *bounds* on the moduli and velocities of the composite rock. However, the elastic bounds are extremely reliable and robust, and they suffer little from the approximations that haunt most of the geometry-specific effective-medium models. Furthermore, since well logs yield information on constituents and their volume fractions, but relatively little about grain and pore microstructure, the bounds turn out to be extremely valuable rock physics tools.

Figure 1.2 illustrates the concept for a simple mixture of two constituents. These might be two different minerals or a mineral plus fluid (water, oil, or gas). At any given volume fraction of constituents the effective modulus of the mixture will fall between the bounds (somewhere along the vertical dashed line in the figure), but its precise value depends on the geometric details. We use, for example, terms like "stiff pore shapes" and "soft pore shapes" to describe the geometric variations. Stiffer grain or pore shapes cause the value to be higher within the allowable range; softer grain or pore shapes cause the value to be lower.

5 1.2 Velocity–porosity relations for mapping porosity and facies

The Voigt and Reuss bounds

The simplest, but not necessarily the best, bounds are the Voigt (1910) and Reuss (1929) bounds. The *Voigt upper bound* on the effective elastic modulus, M_V , of a mixture of N material phases is

$$M_{\rm V} = \sum_{i=1}^{N} f_i M_i \tag{1.1}$$

with

 f_i the volume fraction of the *i*th constituent M_i the elastic modulus of the *i*th constituent

There is no way that nature can put together a mixture of constituents (i.e., a rock) that is elastically *stiffer* than the simple arithmetic average of the constituent moduli given by the Voigt bound. The Voigt bound is sometimes called the *isostrain* average, because it gives the ratio of average stress to average strain when all constituents are assumed to have the same strain.

The Reuss lower bound of the effective elastic modulus, $M_{\rm R}$, is

$$\frac{1}{M_{\rm R}} = \sum_{i=1}^{N} \frac{f_i}{M_i}$$
(1.2)

There is no way that nature can put together a mixture of constituents that is elastically *softer* than this harmonic average of moduli given by the Reuss bound. The Reuss bound is sometimes called the *isostress* average, because it gives the ratio of average stress to average strain when all constituents are assumed to have the same stress.

Mathematically the *M* in the Voigt and Reuss formulas can represent any modulus: the bulk modulus *K*, the shear modulus μ , Young's modulus *E*, etc. However, it makes most sense to compute the Voigt and Reuss averages of only the shear modulus, $M = \mu$, and the bulk modulus, M = K, and then compute the other moduli from these, using the rules of isotropic linear elasticity.

Figure 1.3 shows schematically the bounds for elastic bulk and shear moduli, when one of the constituents is a liquid or gas. In this case, the lower bound corresponds to a suspension of the particles in the fluid, which is an excellent model for very soft sediments at low effective stress. Note that the lower bound on shear modulus is zero, as long as the volume fraction of fluid is nonzero.

The Reuss average describes exactly the effective moduli of a suspension of solid grains in a fluid. This will turn out to be the basis for describing certain types of clastic sediments. It also describes the moduli of "shattered" materials where solid fragments are completely surrounded by the pore fluid.



Figure 1.3 Conceptual illustration of upper and lower bounds to bulk and shear moduli for a mixture of two materials, one of which is a fluid.

When all constituents are gases or liquids with zero shear modulus, then the Reuss average gives the effective moduli of the mixture, exactly.

In contrast to the Reuss average which describes a number of real physical systems, real isotropic mixtures can never be as stiff as the Voigt bound (except for the single-phase end members).

Hashin-Shtrikman bounds

The best bounds for an *isotropic* elastic mixture, defined as giving the narrowest possible range of elastic moduli without specifying anything about the geometries of the constituents, are the *Hashin–Shtrikman bounds* (Hashin and Shtrikman, 1963). For a mixture of two constituents, the Hashin–Shtrikman bounds are given by

$$K^{\text{HS}\pm} = K_1 + \frac{f_2}{(K_2 - K_1)^{-1} + f_1(K_1 + 4\mu_1/3)^{-1}}$$

$$\mu^{\text{HS}\pm} = \mu_1 + \frac{f_2}{(\mu_2 - \mu_1)^{-1} + 2f_1(K_1 + 2\mu_1)/[5\mu_1(K_1 + 4\mu_1/3)]}$$
(1.3)

with

 K_1 , K_2 bulk moduli of individual phases μ_1 , μ_2 shear moduli of individual phases f_1 , f_2 volume fractions of individual phases

Upper and lower bounds are computed by interchanging which material is subscripted 1 and which is subscripted 2. Generally, the expressions give the upper bound when the *stiffest* material is subscripted 1 in the expressions above, and the lower bound when the *softest* material is subscripted 1.

The physical interpretation of a material whose bulk modulus would fall on one of the Hashin–Shtrikman bounds is shown schematically in Figure 1.4. The space is filled by an assembly of spheres of material 2, each surrounded by a spherical shell of material 1. Each sphere and its shell have precisely the volume fractions f_1 and f_2 .

2

7 1.2 Velocity–porosity relations for mapping porosity and facies



Figure 1.4 Physical interpretation of the Hashin–Shtrikman bounds for bulk modulus of a two-phase material.

The upper bound is realized when the stiffer material forms the shell; the lower bound, when it is in the core.

A more general form of the Hashin–Shtrikman bounds, which can be applied to more than two phases (Berryman, 1995), can be written as

$$K^{\text{HS}+} = \Lambda(\mu_{\text{max}}), K^{\text{HS}-} = \Lambda(\mu_{\text{min}})$$

$$\mu^{\text{HS}+} = \Gamma(\zeta(K_{\text{max}}, \mu_{\text{max}})), \mu^{\text{HS}-} = \Gamma(\zeta(K_{\text{min}}, \mu_{\text{min}}))$$
(1.4)

where

$$\Lambda(z) = \left\langle \frac{1}{K(r) + 4z/3} \right\rangle^{-1} - \frac{4}{3}z$$
$$\Gamma(z) = \left\langle \frac{1}{\mu(r) + z} \right\rangle^{-1} - z$$
$$\zeta(K, \mu) = \frac{\mu}{6} \left(\frac{9K + 8\mu}{K + 2\mu} \right)$$

The brackets $\langle \cdot \rangle$ indicate an average over the medium, which is the same as an average over the constituents, weighted by their volume fractions.

The separation between the upper and lower bounds (Voigt–Reuss or Hashin–Shtrikman) depends on how elastically different the constituents are. As shown in Figure 1.5, the bounds are often fairly similar when mixing solids, since the elastic moduli of common minerals are usually within a factor of two of each other. Since many effective-medium models (e.g., Biot, 1956; Gassmann, 1951; Kuster and Toksöz, 1974) assume a homogeneous mineral modulus, it is often useful (and adequate) to represent a mixed mineralogy with an "average mineral" modulus, equal either to one of the bounds computed for the mix of minerals or to their average $(M^{HS+} + M^{HS-})/2$.



Figure 1.5 On the left, a mixture of two minerals. The upper and lower bounds are close when the constituents are elastically similar. On the right, a mixture of mineral and water. The upper and lower bounds are far apart when the constituents are elastically different.

On the other hand, when the constituents are quite different – such as minerals and pore fluids – then the bounds become quite separated, and we lose some of the predictive value.

Note that when $\mu_{\min} = 0$, then $K^{\text{HS}-}$ is the same as the Reuss bound. In this case, the Reuss or Hashin–Shtrikman lower bounds describe exactly the moduli of a suspension of grains in a pore fluid. These also describe the moduli of a mixture of fluids and/or gases.

1.2.2 Generalized velocity–porosity models for clastics

Brief "life story" of a clastic sediment

The bounds provide a framework for understanding the acoustic properties of sediments. Figure 1.6 shows P-wave velocity versus porosity for a variety of water-saturated sediments, ranging from ocean-bottom suspensions to consolidated sandstones. The Voigt and Reuss bounds, computed for mixtures of quartz and water, are shown for comparison. (Strictly speaking, the bounds describe the allowable range for elastic moduli. When the corresponding P- and S-wave velocities are derived from these moduli, it is common to refer to them as the "upper and lower bounds on velocity.")

Before deposition, sediments exist as particles suspended in water (or air). As such, their acoustic properties must fall on the Reuss average of mineral and fluid. When the sediments are first deposited on the water bottom, we expect their properties still to lie on (or near) the Reuss average, as long as they are weak and unconsolidated. Their porosity position along the Reuss average is determined by the geometry of the particle packing. Clean, well-sorted sands will be deposited with porosities near 40%. Poorly sorted sands will be deposited at lower porosities. Chalks will be deposited at



9 1.2 Velocity–porosity relations for mapping porosity and facies

Figure 1.6 P-wave velocity versus porosity for a variety of water-saturated sediments, compared with the Voigt–Reuss bounds. Data are from Yin (1992), Han (1986) and Hamilton (1956).

high initial porosities, 55–65%. We sometimes call this porosity of the newly deposited sediment the *critical porosity* (Nur, 1992). Upon burial, the various processes that give the sediment strength – effective stress, compaction, and cementing – must move the sediments off the Reuss bound. We observe that with increasing diagenesis, the rock properties fall along steep trajectories that extend upward from the Reuss bound at critical porosity, toward the mineral end point at zero porosity. We will see below that these diagenetic trends can be described once again using the bounds.

Han's empirical relations

Figure 1.7 shows typical plots of seismic V_P and V_S vs. porosity for a large set of laboratory ultrasonic data for water-saturated sandstones (Han, 1986). All of the data points shown are at 40 MPa effective pressure. In both plots, we see the usual general trend of decreasing velocity with increasing porosity. There is a great deal of scatter around the trend, which we know from Han's work is well correlated with the clay content. Han described this velocity–porosity–clay behavior with the empirical relations:

 $V_{\rm P} = 5.59 - 6.93\phi - 2.13C$ $V_{\rm S} = 3.52 - 4.91\phi - 1.89C$

(1.5)



Figure 1.7 Velocity versus porosity for water-saturated sandstones at 40 MPa. Data are ultrasonic measurements from Han (1986).

where the velocities are in km/s, ϕ is the porosity, and *C* is the clay volume fraction. These relations can be rewritten slightly in the form

$$V_{\rm P} = (5.59 - 2.13C) - 6.93\phi$$

$$V_{\rm S} = (3.52 - 1.89C) - 4.91\phi$$
(1.6)

which can be thought of as a series of parallel velocity–porosity trends, whose zeroporosity intercepts depend on the clay content. These contours of constant clay content are illustrated in Figure 1.8, and are essentially the steep diagenetic trends mentioned in Figure 1.6. Han's clean (clay-free) line mimics the diagenetic trend for clean sands, while Han's more clay-rich contours mimic the diagenetic trends for dirtier sands. Vernik and Nur (1992) and Vernik (1997) found similar velocity–porosity relations, and were able to interpret the Han-type contours in terms of petrophysical classifications of siliciclastics. Klimentos (1991) also obtained similar empirical relations between velocity, porosity, clay content and permeability for sandstones.

As with any empirical relations, equations (1.5) and (1.6) are most meaningful for the data from which they were derived. It is dangerous to extrapolate them to other situations, although the concepts that porosity and clay have large impacts on P- and S-wave velocities are quite general for clastic rocks.

When using relations like these, it is very important to consider the coupled effects of porosity and clay. If two rocks have the same porosity, but different amounts of clay, then chances are good that the high clay rock has lower velocity. But if porosity decreases as clay volume increases, then the high clay rock might have a higher velocity. (See also Section 2.2.3.)