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P. Cosenza, Mehdi Ghoreychi, G. de Marsily, Guy Vasseur, S. Violette

Category: Subsurface Hydrology

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Ph. Cosenza[1,2], M. Ghoreychi[3], G. de Marsily [1], G. Vasseur [1], S. Violette [1]

[1] UMR 7619 Sisyphe, Paris VI University, France
Case courrier 105,
4 Place Jussieu
75252 Paris cedex 05
tél : 33/0 1 44 27 43 81 - fax : 33/0 1 44 27 45 88
email : cosenza@ccr.jussieu.fr

[2] Institut de Sciences et Technologie, Paris VI University, France

[3] Institut National de l'Environnement Industriel et des Risques (INERIS), France

Abstract

An integrated and theoretical approach is proposed to determine poroelastic properties of isotropic or transversely isotropic argillaceous rocks from conventional mechanical and hydraulic field tests. It uses values of specific storage coefficient, "undrained" elastic parameters and porosity. No special hydromechanical test is required. In the isotropic case, a simple binomial equation is derived, whereby the poroelastic properties can be rapidly obtained. An application to deep argillaceous formations has shown that calculated poroelastic property values are in satisfactory agreement with direct experimental measurements. In the transversely isotropic case, which is more representative of sedimentary rocks, the approach is generalized from a theoretical relationship between the specific storage coefficient and anisotropic poroelastic properties. Nevertheless, in both cases, sensitivity analyses have shown that the method is very sensitive to input data, especially the specific storage coefficient and the "undrained" mechanical parameters which have to be measured with accuracy.

Index terms:

Physical properties of rocks: 5199 General or Miscellaneous

Hydrology: 18199 General or Miscellaneous

Key words: Specific storage coefficient, Poroelastic properties, Biot's coefficient, Claystone.

1. Introduction

The coupling of stress-strain and pore fluid pressure in deep clayey rocks is relevant to many academic and practical problems in Earth Sciences: pore pressure build-up resulting from compaction of a sedimentary basin during its formation [e.g., *Bredehoeft and Hanshaw*, 1968]; the hydromechanical behaviour of clayey rocks due to excavation of tunnels or deep repositories [e.g., *Benamar et al,* 1998], the hydromechanical response of nuclear waste disposal due to heating [e.g., *Palciauskas and Domenico*, 1982] and the interpretation of hydromechanical tests in such a geological medium [e.g., *Djeran-Maigre and Gasc-Barbier*, 2000]. All these problems involve strong coupling between pressurization, fluid motion and deformation of the porous rock. The pioneering theoretical work on hydromechanical behaviour of isothermal porous media was done by *Biot* [1941, 1956]. *Rice and Cleary* [1976] have developed general solutions to classical initial/boundary value problems and recast Biot's theory in terms of new material parameters, more directly open to physical interpretation.

The poroelastic parameters are measured by "drained" and "undrained" triaxial compression tests [e.g., *Charlez*, 1997]. The terms "drained" and "undrained" refer to the boundary conditions where there are no changes respectively in pore fluid pressure and in pore fluid mass. However, considering tight argillaceous rocks, it is not practical to measure all the poroelastic parameters, in particular the "drained" elastic moduli. Indeed, considering low permeability media with permeabilities between 10^{-19} and 10^{-21} m^2, it is difficult to measure and *a fortiori* to control pore pressure in the sample. These measurements require special experimental techniques, which demand strictly controlled tightness and may be unacceptably long. Moreover, during long transient flow, viscosity of the porous skeleton may also play a significant role leading to effective stress changes and time-dependent mechanical behavior. Hence the interpretation of mechanical tests may become more complex.

Consequently, it is desirable to have some method of correlating the "drained" elastic moduli to physical parameters, which would be easier to obtain experimentally. One idea, initially used by *Domenico and Mifflin* [1965] is to relate "drained" mechanical parameters to the specific storage coefficient. In ground-water hydraulics or hydrogeology, their expression is generally used to compute an order of magnitude of the specific storage for a confined aquifer for typical lithology [e.g., *Domenico and Schwartz,* 1997). Later, *Green and Wang* [1990] derived a new relationship between the specific storage coefficient and "drained" poroelastic moduli written using the formalism of *Rice and Cleary* [1976]:

$$S_s = \rho^{fl} g\left[\left(\frac{1}{K}-\frac{1}{K_s}\right)\left(1-\frac{4G(1-K\ /K_s)}{K\ +4/3G}\right)+\phi\left(\frac{1}{K_f}-\frac{1}{K_s}\right)\right] \quad (1)$$

where S_s is the specific storage coefficient, ρ^{fl} is the pore fluid density, K is the "drained" bulk modulus, K_s is the "unjacketed" bulk modulus which is approximately equal to the modulus of the solid part of the rock, G is the shear modulus, ϕ is the open porosity, K_f is the bulk modulus of the fluid, g is gravitational acceleration. This expression is often applied to determine the value of the specific storage coefficient prior to model permeability field tests [e.g., *Beauheim et al.*, 1991] and laboratory tests on transient hydraulic flow [e.g., *Neuzil,* 1986]

In this paper, we investigate this relationship in a way opposite to that usually used in hydrogeology: how can we use the specific storage measurements of *in situ* argillaceous rocks to calculate their poroelastic properties?

In the first part of the paper, the proposed method is shown to simply and rapidly provide an estimate of the poroelastic properties of isotropic argillaceous rocks from the specific storage coefficient with a satisfactory range of uncertainty. This method is based on an equation similar to that of *Green and Wang* [1990], which is reformulated in order to introduce the "undrained" bulk modulus, easier to measure in argillaceous rocks. It requires values of the specific storage coefficient obtained from "modified" slug tests as discussed by *Bredehoeft and Papadopoulos* [1980] for low-permeability formations. The assumptions and underlying uncertainties of such an approach are discussed and an attempt is made to validate it. This approach is applied to three deep argillaceous formations from the Paris sedimentary basin (taken at a depth of 375 m to 855 m).

In the second part, our aim is to establish an expression relating the specific storage coefficient to poroelastic constants from *transversely* isotropic clayey rocks. In this particular condition of anisotropy, a non-linear system of equations is derived and solved numerically for the Toarcian shale from the Tournemire site in southern France.

2. Isotropic case

2.1 Governing equations

Let us consider the following assumptions :

- An infinitely long and vertical borehole is drilled in a homogenous and isotropic porous medium. The borehole radius is designated as r_0.
- The vector of displacement $\underline{u}$ is purely radial : $\underline{u} = u(r,t)\underline{e}_r$ where $\underline{e}_r$ is the radial unit vector in a cylindrical coordinate system.
- The boundary conditions far from the borehole wall, are written as follows:

  $P(r \to \infty, t) = P_0$ or $\partial P / \partial r = 0$ (hydraulic condition)

  $tr\underline{\underline{\varepsilon}} = 0$ (no volumetric strain) or $\sigma_r = \sigma_0$ (mechanical condition),

  tr : trace operator; $\underline{\underline{\varepsilon}}$: strain tensor; P : pore fluid pressure ; σ_r : radial component of the stress tensor $\underline{\underline{\sigma}}$, σ_0 : initial lithostatic pressure.

From these assumptions, the combination of Darcy's law, of the continuity equation for the fluid mass and of a constitutive poroelastic law yields [e.g., *Rice and Cleary*, 1976, or *Coussy*, 1995):

$$\frac{\partial P}{\partial t} = c_m \nabla^2 P \qquad (2)$$

where c_m is the hydraulic diffusivity coefficient (m^2/s) defined by :

$$c_m = \frac{k}{\eta} M \frac{3K + 4G}{3K^u + 4G} \tag{3}$$

where k is the intrinsic permeability (m^2), η is the fluid dynamic viscosity (Pa.s), M is Biot's modulus (Pa), K is the « drained » bulk modulus (Pa), K^u is the « undrained » bulk modulus (Pa), G is the shear modulus (Pa).

It should be emphasized that equation (2) is independent of the boundary condition defined at the borehole wall ($p(r=r_0,t)$). In our problem, the parameters k, η, K^u, G are given or can be measured, but constants M and K are unknown.

In hydrogeology, the specific storage coefficient (expressed in m^{-1}) can be defined from the following governing equation for transient flow [e.g., *Marsily*, 1986]:

$$S_s \frac{\partial P}{\partial t} = \nabla . \left[\frac{k}{\eta} \rho^{fl} g \left(\underline{\nabla} P + \rho^{fl} g \underline{\nabla} z \right) \right] \tag{4}$$

where g is the gravitational acceleration, ρ^{fl} is the pore fluid density, z is the elevation. If the flow is purely radial ($\underline{\nabla} z = 0$) and if parameters k, η and ρ^{fl} are constants (i.e. independent of the radius r), equation (4) can be rewritten as follows:

$$S_s \frac{\partial P}{\partial t} = \frac{k}{\eta} \rho^{fl} g \nabla^2 P \tag{5}$$

By comparing equations (2) and (5) one can express the specific storage coefficient S_s as a function of poroelastic properties:

$$\frac{S_s}{\rho^{fl} g} = \frac{1}{M} \frac{3K^u + 4G}{3K + 4G} \tag{6}$$

This expression is consistent with that of *Green and Wang* [1990] as shown in Appendix A.

2.2 A method for determining poroelastic properties from S_s

Equation (6) is now used to constrain the determination of poroelastic properties of low-permeability media. But this determination will be achieved if complementary "micromechanical" equations are considered. The term "micromechanical" refers to fluid and matrix properties, which are included in these equations. These equations are widely used in rock mechanics and may reduce the number and hence the cost of the measurements [e.g., *Charlez*, 1991]. They are based on the two following assumptions [*Nur and Byerlee*, 1971):

- The matrix is homogeneous, isotropic and elastic. Here, the matrix is considered, not only as a solid (grains), but as the association of the grains and the unconnected porosity.
- The connected porosity is supposed to be constant when the rock sample is subjected to the particular hydromechanical loading: $\Delta\sigma_m = \Delta P$ with σ_m : total mean stress (positive in compression) and P : pore pressure.

These "micromechanical" equations are written as follows:

$$\alpha = 1 - \frac{K}{K_s} \tag{7}$$

$$\frac{1}{M} = \frac{\alpha - \phi}{K_s} + \frac{\phi}{K_{fl}} \tag{8}$$

where K_s and K_{fl} are the bulk modulus of the matrix and of the liquid respectively, α is Biot's coefficient which can also be defined by the concept of effective stress :

$$\underline{\underline{\sigma}}^{eff} = \underline{\underline{\sigma}} - \alpha P \underline{\underline{1}} \tag{9}$$

where $\underline{\underline{\sigma}}^{eff}$ is the effective stress tensor, $\underline{\underline{\sigma}}$: total stress tensor, $\underline{\underline{1}}$:is the unit tensor. Equation (9) can be viewed as a generalization of the so-called Terzaghi's equation in soil mechanics (with α=1 considering $K_s \to \infty$ in equation 7).

As previously mentioned, the "drained" bulk modulus K is difficult to measure experimentally in low-permeability media. It is thus convenient to introduce the "undrained" bulk modulus K^u which can be related to K as follows:

$$K^u = K + \alpha^2 M \tag{10}$$

Equation (10) is obtained by comparing the poroelastic constitutive laws written for "drained" and "undrained" conditions [*Coussy*, 1995]. By combining of equations (7), (8) and (10) one can rewrite the specific storage coefficient (equation 6) according to the form given by *Green and Wang* [1990] (equation 1) (see appendix A).

The set of equations (6), (7), (8) and (10) forms a system of 4 equations with 4 unknowns: α, M, K, K_s. In order to solve this system, the following data are required:

- The "undrained" bulk modulus K^u and the shear modulus G, which are obtained from conventional "undrained" compression tests in rock mechanics [e.g., *Charlez*, 1997]. These "undrained" tests are easily performed since the pore pressure in the sample is not measured.
- The specific storage S_s, which can be obtained from "modified" slug tests [*Bredehoeft and Papadopoulos*, 1980] for low-permeability formations.
- The connected porosity ϕ measured by a standard mercury porosimetry technique.
- The bulk modulus of the liquid (mostly water) K_{fl}.

No hydromechanical coupling parameter is required.

By manipulating the system of equations (6), (7), (8) and (10), a simple binomial equation is obtained:

$$C_1\alpha^2 + C_2\alpha + C_3 = 0 \qquad (11)$$

with :

$$C_1 = \frac{K^u + 4/3G}{\Sigma}\left(\frac{\phi}{K_{fl}} - \frac{K^u\Sigma}{K^u + 4/3G}\right)^2 + (1+\phi)^2 \qquad (12a)$$

$$C_2 = -(1+\phi)\left[\frac{\phi}{K_{fl}}(4/3G - K^u) + 2\phi + K^u\Sigma\right] \qquad (12b)$$

$$C_3 = -\phi\left(\frac{4}{3}\frac{G}{K_{fl}} + 1\right)\left(\frac{K^u}{K_{fl}}\phi - \phi - K^u\Sigma\right) \qquad (12c)$$

where $\Sigma = \dfrac{S_s}{\rho^{fl} g}$ (in Pa^{-1}).

Equation (11) can be easily solved and has real solution(s) if the following obvious inequality is satisfied $C_2^2-4C_1C_3 \geq 0$. The others unknowns K_s, K and M are determined by combining equations (7), (8) or (10). For instance, Biot's modulus M is given by the following relationship:

$$M = \frac{\dfrac{\phi(K^u + 4/3G)}{\Sigma K_{fl}} - K^u}{\dfrac{4}{3}\phi\dfrac{G}{K_{fl}} + \phi - \alpha(1+\phi)} \qquad (13)$$

It should be emphasized that all the values of parameters α, K_s, K and M have to satisfy the following inequalities [*Coussy*, 1995]:

$$K \leq K^u \leq K_s \qquad 0 \leq \alpha \leq 1 \qquad \beta_{skempton} = \frac{\alpha M}{K^u} \leq 1 \qquad (14)$$

Parameter $\beta_{skempton}$ is called Skempton's coefficient or the pore pressure build-up coefficient,

$$\beta_{skempton} = \left(\frac{\partial P}{\partial \sigma_m}\right)_{dm=0} \qquad (15)$$

defined as the change in pore pressure P per unit change in total mean stress σ_m applied in undrained condition (dm=0, no mass supply of liquid).

2.3 Validation

As mentioned above, this method requires a value of the specific storage coefficient S_s which can be obtained from "modified" slug tests for low-permeability formations [*Bredehoeft and Papadopoulos*, 1980]. Such a value obtained from a field experiment is preferable to laboratory data since it is the more representative of *in situ* rock behaviour.

The "modified" slug test or "Pulse" test is based on an equation that describes the decay of an abrupt head change caused by pressurizing the volume stored in a closed well. This equation is an analytical solution of the following boundary value problem, which can be expressed as

$$\frac{\partial^2 h}{\partial r^2}+\frac{1}{r}\frac{\partial h}{\partial r}=\frac{S_s}{K}\frac{\partial h}{\partial t} \qquad (16)$$

with the initial and boundary conditions :

$h(r,0)=0$

$h(\infty,t)=0 \quad ; \quad h(r_s,t)=H(t) \quad ; \quad H(0)=H_0$

$$2\pi Ke\frac{\partial h}{\partial r}(r_s,t)=V_w C^* \rho^{fl} g\frac{\partial H}{\partial t}(t)$$

where h : head change in the tested interval of the formation due to pressurization (m), r : radial distance from the center of the well (m), S_s : the specific storage coefficient of the tested interval (m^{-1}), K : hydraulic conductivity (m/s) , r_s : radius of the well in the tested interval (m), e: thickness of the tested interval (m); V_w : volume of water within the pressurized section of the system (m^3), C* : effective compressibility of water below the packers which includes the compressibility of both water and test equipment (Pa^{-1}).

The analytical solution of (16) has the following form :

$$\Delta H / \Delta H_0 = F(a,b) \qquad (17a)$$

$$\text{with } a = S_s\, e\frac{\pi r_s^{\,2}}{V_w C^* \rho^{fl} g} \text{ et } b = Ket\frac{\pi}{V_w C^* \rho^{fl} g} \qquad (17b)$$

Bredehoeft and Papadopoulos [1980] proposed a procedure to determine the dimensionless parameters *a* and *b*, and hence the physical parameters K and S_s. As mentioned by these authors, we are aware that this method may not be able to measure small S_s, which is found in stiff rocks with low porosity (typically crystalline rocks). Very small S_s may lead to the same transient response curve for different pressurizations. In particular, the authors stated that a determination of S_s has questionable reliability when the dimensionless number *a* is smaller than 10^{-5} [*Papadopoulos et al.*, 1973]. Consequently, in the following, the inequality $a > 10^{-5}$ will be considered as a criterion for applying the approach presented in this paper.

The proposed method is difficult to validate for at least two reasons. On the one hand, natural variability may affect numerous parameters, which have to be quantified (see for instance Table 1). On the other hand, this validation requires a set of data for a given clayey formation from different scientific fields: hydrogelogical, mechanical and hydromechanical. These data have to be determined independently of each other.

To our knowledge, the only clayey formation for which a rather extensive set of data is available, is the claystone belonging to the Callovo-Oxfordian (sampled at about 400-500 m deep in the Paris basin). This claystone is under study by Agence Nationale de Gestion des Déchets Radiocatifs (ANDRA) for a potential deep radioactive waste repository. Table 1 shows mean experimental values of the data we used to solve equation (11). From Table 1., the mean calculated value of Biot's coefficient α, is 0.4 and the dimensionless number a is equal to $5\ 10^{-2}$. This value of Biot's coefficient is physically acceptable and can be compared with that obtained directly from hydromechanical tests. By using a special consolidation test-oedometer, *Vincké et al.* [1997] measured Biot's coefficient on the same claystone in the range [0.4-0.8] at an applied uniaxial stress in the range [9-35 MPa]. By using a triaxial cell and an experimental procedure similar to that of *Neuzil et al.* [1981], *Coste et al.* [1999] obtained a value of Biot's coefficient of 0.36 for the same claystone.

However, the previous comparison cannot be considered as a complete validation of the method since:

(a) In *Vincké*'s experiments, the values of Biot's coefficient are measured on samples for which the stress state and the induced porosity changes are unknown. The extrapolation to the hydromechanical conditions of the *in situ* rock is difficult. In Coste's experiment, the value of 0.36 was obtained on a single sample.

(b) This preliminary calculation does not take into account the variation range of the different parameters. As will be shown later, the variation range may be large for particular parameters and hence may significantly influence our results. This point is discussed in the following section.

This comparison with all of these experimental values leads to apply the method to other argillaceous rocks.

2.3 Applications and Parameter Sensitivity Analyses

In order to identify a suitable site for disposal of French high radioactive waste, ANDRA has conducted geophysical and hydrogeological surveys in the main low-permeability argillaceous formations of the Paris basin. In borehole number 901 drilled during the "Aisne" campaign (1987-1990), "Pulse-tests" have been performed in two clayey rocks belonging to the Callovo-Oxfordian [375-478 m deep] and Toarcian-Domerian [692-855 m deep].

Laboratory measurements of mechanical parameters were carried out by Groupement pour l'étude des Structures Souterraines de Stockage (G.3S). The compressional and shear wave velocities were measured by ultrasonic pulse in two perpendicular directions (parallel and perpendicular to the stratification). No clear anisotropy was observed. The total porosity was measured with a mercury porosimeter at the Bureau de Recherche Géologique et Minières (BRGM). All these data are available in *Cellier* [1998].

The dimensionless parameter a was calculated from equation (17b) and from data given in tables 2 and 3. The effective compressibility C*, which is not available from Cellier [1988], was taken as equal to six times the compressibility of water on the basis of Neuzil's work [*Neuzil*, 1982]. Table 4 shows calculated values in the range of $2\ 10^{-2}$ to $6\ 10^{-2}$, which are larger than 10^{-5}.

On this site, a lithostratigraphic study has shown that the Toarcian-Domerian formation could be divided into three structural units: a Toarcian claystone [692-772 m deep], a "liassic" sandstone [772-782 m deep] and a Domerian claystone [782-855 m deep]. All the calculations performed by the present approach concerned both claystones. The results are given in Tables 2 and 3. On the basis of the results, the following comments can be made:

- Biot's coefficient α and hence the other poroelastic properties are extremely sensitive to the "undrained" mechanical parameters, E^u and ν^u, and the specific storage coefficient S_S. It should be mentioned that the calculated variation ranges in tables 1 to 3 have to be compared with the strong effect of natural variability associated with mechanical parameters. In rock mechanics, natural variability induces currently variations of up to 100 %.
- The effect of porosity ϕ is small.
- Values of the calculated bulk modulus of the matrix, K_s are small compared to that of sandstone and limestone [*Rice and Cleary*, 1976; *Charlez*, 1997]. This may be due to two features that may contribute to diminish the matrix rigidity: (1) a significant isolated porosity; (2) presence of dispersed clay that may adhere and coat the quartz and carbonate grains and diminish the matrix rigidity.

Figure 1 shows the calculated Biot's coefficients of the Toarcian-Domerian formations and experimental values from tests on other rocks (especially on sandstone and limestone). In Figure 1, error bars are calculated from results in Table 2 and 3 by considering the widest variation ranges. Results of the Callovo-Oxfordian formations are not plotted since the calculated variation range is too wide.

The comparison of results with published values is difficult due to lack of experimental data for deep argillaceous rocks. As mentioned, this is mainly due to the difficulty in measuring such properties in the laboratory. Moreover, the rocks considered in Figure 1 have a liquid phase organization (free water and bound water), which is very different from that of argillaceous rocks and we feel that this organization and the corresponding physical properties may influence significantly the macroscopic poroelastic properties. In particular, bound water related to clay particles leads to increase the compressibility of the solid skeleton since bound water is associated with the solid phase in our approach.

Nevertheless, it is more relevant to plot the result as a function of the permeability of each geomaterial instead of its porosity. Indeed, Figure 2 shows that a representation using permeability seems to be a better tool to compare the different values of Biot's coefficient. In spite of the relatively small number of data, it is interesting to note that results of the "Toarcian - Domerian" formations confirm the general trend observed in Figure 2.

3. Transversely isotropic case

3.1 Governing equations

Ultrasonic measurements and triaxial compression tests with different loading orientations performed on the Callovo-Oxfordian and Toarcian-Domerian formations did not show significant any anisotropic effects. But, it is well known that most argillaceous geomaterials are typically anisotropic. Although they exhibit many forms of anisotropy, this part focuses on the transversely isotropic case. Transverse isotropy is often used to describe the symmetry of rocks with one dominant system of layers, such as foliated and sedimentary rocks. In this case, there exists a rotational symmetry around the axis perpendicular to the bedding planes. The physical properties in all directions parallel to the bedding planes are the same and differ from those perpendicular to the bedding planes.

In this section, the same initial/boundary value problem as in the isotropic case is considered. Moreover, the vertical axis of the borehole is assumed to be perpendicular to the bedding planes (Figure 3). Cheng's formalism is used to describe the mechanical behavior of anisotropic rocks [*Cheng*, 1997]. Considering the previous assumptions, the following diffusion equation is obtained (see Appendix B):

$$\frac{\partial P}{\partial t} = c_m^a \nabla^2 P \tag{18}$$

where c^a_m is the hydraulic diffusivity coefficient (m^2/s) of transversely isotropic rock defined by :

$$c_m^a = \frac{k}{\eta} M \frac{M_{11}^u - \alpha^2 M}{M_{11}^u} = \frac{k}{\eta} M \frac{M_{11}}{M_{11} + \alpha^2 M} \tag{19}$$

where k is the intrinsic permeability in the bedding planes (x-y), α is Biot's coefficient in the x-y plane, M is Biot's modulus, M_{11} is the "drained" poroelastic modulus which is given by [*Cheng*, 1997] :

$$M_{11} = \frac{E(E' - E\nu'^2)}{(1+\nu)(E' - E'\nu - 2E\nu'^2)} \tag{20}$$

where E is the "drained" Young modulus in the x-y plane, E' is the "drained" Young's modulus in the 0z direction (*i.e.* perpendicular to the x-y plane), ν is the "drained" Poisson's coefficient in the x-y plane (defining the lengthening deformation in the x-y plane due to loading in the x-y plane), ν' is the "drained" Poisson's coefficient associated with the 0z direction (defining the lengthening deformation in the x-y plane due to loading normal to the x-y plane). The moduli M_{11} and M_{11}^u are related to each other by the following relationship:

$$M_{11} = M_{11}^u - \alpha^2 M \tag{21}$$

By comparing equations (5) and (18) one obtains the following relationship, which defines the specific storage coefficient of a transversely isotropic rock :

$$\frac{S_s}{\rho^{fl} g} = \frac{1}{M} \frac{M_{11}^u}{M_{11}^u - \alpha^2 M} \tag{22}$$

This relationship is a generalization of the *Green and Wang* [1990] equation to the transversely isotropic case.

3.2 A method for determining poroelastic properties from S_s

Considering a transversely isotropic rock, 8 poroelastic constants are required in order to study its hydromechanical behavior : 5 "drained" moduli M_{ij}, 2 Biot's coefficients α (parallel to the bedding planes) and α' (perpendicular to the bedding planes), and a Biot's modulus M [*Cheng*, 1997]. Moreover, the 5 "drained" moduli can be replaced by 5 "undrained" moduli M^u_{ij}, which are easier to measure in the laboratory:

$$\begin{cases} M_{11} = M^u_{11} - \alpha^2 M \\ M_{12} = M^u_{12} - \alpha^2 M \\ M_{13} = M^u_{13} - \alpha\alpha' M \\ M_{33} = M^u_{33} - \alpha'^2 M \end{cases} \qquad \text{(23a,b,c,d)}$$

As for the isotropic case, "micro-mechanical" assumptions are required to obtain all the poroelastic properties in a simple way: "micro-isotropy" and "micro-homogeneity" assumptions. The latter is described in the first section of this paper.

The "micro-isotropy" assumption is directly associated with the anisotropic geomaterials: the solid constituents that compose the porous material are isotropic at the microscopic (pore and grains) level. The macroscopic anisotropy is of structural origin, i.e. a consequence of directional pore or microcrack arrangements. Both assumptions, "micro-isotropy" and "micro-homogeneity" allow us to relate the bulk modulus of the matrix K_s to other poroelastic properties according to the relationship [*Cheng*, 1997]:

$$M = \frac{K_s}{(1 - \frac{K^*}{K_s}) - \phi(1 - \frac{K_s}{K_f})} \qquad (24)$$

where

$$K^* = \frac{1}{9}(M_{11} + M_{33} + 2M_{12} + 2M_{13}) = \frac{1}{9}(M^u_{11} + M^u_{33} + 2M^u_{12} + 2M^u_{13} - 3\alpha^2 - \alpha'^2 + 2\alpha\alpha')$$

$$\begin{cases} \alpha = 1 - \dfrac{M_{11} + M_{12} + M_{13}}{3K_s} = 1 - \dfrac{M_{11}^u + M_{12}^u + M_{13}^u - 2\alpha^2 M - \alpha\alpha' M}{3K_s} \\ \alpha' = 1 - \dfrac{2M_{13} + M_{33}}{3K_s} = 1 - \dfrac{2M_{13}^u + M_{33}^u - \alpha' M - 2\alpha\alpha' M}{3K_s} \end{cases} \qquad (25a,b,c)$$

In the expression of K* and in equations (25b,c), "drained" moduli M_{ij} are replaced by "undrained" moduli M^u_{ij} from relations (23,a,b,c,d).

Finally, the 4 equations (22), (25b), (25c) and (24) where parameter K* is replaced by (25a) constitute the following system (I) of nonlinear equations with 4 unknowns, α, α', M and K_s :

$$(\mathrm{I}) \quad \left| \begin{array}{l} \alpha^2 M^2 \dfrac{S_s}{\rho^{fl} g} - M M_{11}^u \dfrac{S_s}{\rho^{fl} g} + M_{11}^u = 0 \\ -3\alpha K_s + 3K_s + 2\alpha^2 M + \alpha\alpha' M - (M_{11}^u + M_{12}^u + M_{13}^u) = 0 \\ -3\alpha' K_s + 3K_s + 2\alpha\alpha' M + \alpha'^2 M - (M_{33}^u + 2M_{13}^u) = 0 \\ -\dfrac{M}{9}(M_{11}^u + M_{33}^u + 2M_{12}^u + 2M_{13}^u) + \dfrac{M^2}{9}(3\alpha^2 + \alpha'^2 + 2\alpha\alpha') + \\ \qquad + (1-\phi)MK_s + \dfrac{\phi}{K_{fl}} MK_s^2 - K_s^2 = 0 \end{array} \right.$$

In order to solve system (I), the following data are required: connected porosity ϕ, specific storage coefficient S_s and "undrained" moduli M_{ij} which can be calculated from elastic moduli E^u, E'^u, ν^u and ν'^u (see Appendix B).

3.3 Application

In its research program of safety studies of geological waste disposal, the French Institute for Nuclear Protection and Safety (IPSN) is studying, at the Tournemire site near Roquefort, Aveyron, France, a geological formation constituted by claystones of Toarcian. This site has been selected by IPSN because of its geological simplicity and also because a former railway tunnel gives access to the center of the Toarcian formation. The tunnel crosses a 200 m-thick formation of claystones (Toarcian) covered by more than 250 m of limestone (Dogger).

System (I) has been solved numerically with petrophysical [*Barbeau and Boisson*, 1993], hydrogeological [*Boisson et al.,* 1998] and mechanical [*Niandou et al.*, 1997] data from the Tournemire site. The results of *in-situ* "Pulse-tests" performed at this site are given in Table 5. The porosity, which was measured with a mercury porosimeter is in the range of [2.2-4.2 %] [*Barbeau and Boisson*, 1993].

The inequality $a > 10^{-5}$ was checked by using the same compressibility value $C^{*}=3\ 10^{-9}\ Pa^{-1}$ as for the previous calculations in isotropic claystones. The "undrained" poroelastic parameters E^u, E'^u, ν^u and ν'^u were calculated from the following empirical equations, which were established by *Niandou et al.* [1997] from their mechanical data:

$$E'^u = E'_s - (E'_s - E'_0)\ e^{-\alpha(p/p^*)} \tag{26}$$

with E'_s= 17 GPa; E'_0=4 GPa ; α=0.032

$$E^u = E_s - (E_s - E_0)\ e^{-\beta(p/p^*)} \tag{27}$$

with E_s= 45 GPa; E_0=22 GPa ; β=0.016

$$\nu'^u = \nu'_s - (\nu'_s - \nu'_0)\ e^{-\gamma(p/p^*)} \tag{28}$$

with $\nu'_s = 0.75$; $\nu'_0 = 0.2$; $\gamma = 0.0068$

$$\nu^u = \nu_s - (\nu_s - \nu_0)\ e^{-\lambda(p/p^*)} \tag{29}$$

with $\nu_s = 0.2$; $\nu_0 = 0.12$; $\lambda = 0.013$

In these equations, parameters p and p* are the mean stress and a reference value (p*=1 GPa) respectively. The parameters with subscript "0" are the initial values of elastic parameters for $p \leq 0$ and those with subscript "s" the asymptotic values for very high stress levels (i.e. at great depth).

In Table 6, the elastic parameters E^u, E'^u, ν^u and ν'^u were calculated considering the depth of the hydraulic tests. Note in Table 6 that a strong difference exists between the moduli E^u and E'^u (ratio greater than 3) due to the significant anisotropic behavior of this claystone.

A numerical globally convergent method was used to solve the nonlinear system (I) of the equations [*Press et al.*, 1992] From the results obtained with different porosity values given in Table 6, the following remarks can be made:

(1) Values of Biot's coefficient α and bulk modulus of the matrix K_s are consistent with the values for other sedimentary rocks [*Rice and Cleary*, 1976; *Charlez*, 1997].

(2) Considering a constant specific storage coefficient and constant porosity (tests 4 and 5), the changes in properties α and M induced by "undrained" moduli changes with depth are small, below 6%.

(3) Values of Biot's coefficient perpendicular to the bedding planes α' are large and greater than 1. These values may be explained by the strong difference between Young's moduli parallel to the bedding planes E^u and those perpendicular to the bedding planes E'^u. Indeed, the ratio between E^u and E'^u is 3.5 on average (see Table 5), which is very high for an argillaceous rock. By comparison, for schists, which are highly anisotropic, typical ratios given by *Talobre* [1967] are in the range [1.2-2]. Schists habitually have macroscopic joints, which can evolve into planes of weakness and on the contrary, Tournemire argillaceous rocks are macroscopically homogeneous [*Barbeau and Boisson*, 1993]. Consequently, the high E^u/E'^u ratio obtained by Niandou *et al.* [1997], suggests that the samples were initially damaged and have open fractures/bedding planes due to the decompression of the rock. When mechanical loading was applied in order to measure the Young modulus E'^u, the strain measurement was certainly associated with the closure of such open fractures/bedding planes and not representative of the solid skeleton deformation. Consequently, the values of parameter E'^u are questionable and this leads to an inconsistent and unrealistic set of data for the porous, transversely isotropic argillaceous rock at the Tournemire site. This inconsistency may explain the large values of Biot's coefficient α' compared to Biot's coefficient for the in-bedding plane α.

4. Conclusion

In our opinion, this paper is primarily of methodological interest. Considering an isotropic clayey rock, this method allows us to determine its poroelastic properties in a simple way: it is based on the resolution of a binomial equation (equation 11), which requires data from conventional mechanical and hydrological tests. No special hydromechanical tests, which are difficult to perform on low-permeability materials, are required. Applications to deep argillaceous formations, which are being studied for deep disposal in France, have shown that the calculated poroelastic properties are in satisfactory agreement with direct experimental measurements. Nevertheless, sensitivity analyses have shown that the method is very sensitive to input data, especially the specific storage coefficient and the "undrained" Young's modulus. In this respect, this method can be used as a first approach to rapidly estimate poroelastic properties of low-permeability media by integrating existing data from different scientific fields.

This method was generalized to the transversely isotropic case, which is more representative of sedimentary rocks. In particular, a theoretical relationship was established between the specific storage coefficient and anisotropic poroelastic properties. An application to an anisotropic argillaceous rocks from the Tournemire site in southern France confirmed that the approach is very sensitive to the input parameters, both mechanical and hydrogeological. The more accurate these parameters are, the more efficient the method. This fact underlines the need to continue improving *in situ* experimental techniques in order to obtain data that are not only more accurate but also more representative of the field conditions.

Appendix A. Consistency of the specific storage coefficient formulation (equation 6) with the *Green and Wang* [1990] equation

In this paper, the specific storage coefficient S_s is expressed as:

$$\frac{S_s}{\rho^{fl} g} = \frac{1}{M} \frac{3K^u + 4G}{3K + 4G} \tag{A1}$$

where S_s is the specific storage coefficient, ρ^{fl} is the pore fluid density, g is gravity, M is Biot's modulus, K is the "drained" bulk modulus, K^u is the "undrained" bulk modulus, G is the shear modulus. In order to obtain the same formulation as that of *Green and Wang* [1990], consider the following "micromechanical" and compatibility equations:

$$\alpha = 1 - \frac{K}{K_s} \tag{A2}$$

$$\frac{1}{M} = \frac{\alpha - \phi}{K_s} + \frac{\phi}{K_{fl}} \tag{A3}$$

$$K^u = K + \alpha^2 M \tag{A4}$$

Using equation (A4) to eliminate K^u we obtain:

$$\frac{S_s}{\rho^{fl} g} = \frac{1}{K + \frac{4}{3}G}\left(\frac{K}{M} + \frac{4}{3}\frac{G}{M} + \alpha^2\right) \tag{A5}$$

From equation (A5), consider the following form:

$$\frac{S_s}{\rho^{fl} g} = \frac{1}{K + \frac{4}{3}G}\left[\frac{K}{M} + \frac{4}{3}\frac{G}{M} + \alpha^2 - \alpha + \alpha + \alpha(1-\alpha)\frac{4G}{3K} - \alpha(1-\alpha)\frac{4G}{3K}\right] \tag{A6}$$

or

$$\frac{S_s}{\rho^{fl} g} = \frac{1}{K + \frac{4}{3}G}\left\{\frac{\alpha}{K}\left[K + (1-\alpha)\frac{4G}{3}\right] + (K + \frac{4}{3}G)\left[\frac{1}{M} - \frac{\alpha(1-\alpha)}{K}\right]\right\} \tag{A7}$$

From equation (A3), we have

$$\frac{1}{M}-\frac{\alpha}{K_s}=\phi\left(\frac{1}{K_{fl}}-\frac{1}{K_s}\right) \tag{A8}$$

And from equation (A2):

$$K_s=\frac{K}{1-\alpha} \tag{A9}$$

Introducing equation (A9) into equation (A8), we obtain:

$$\frac{1}{M}-\frac{\alpha(1-\alpha)}{K}=\phi\left(\frac{1}{K_{fl}}-\frac{1}{K_s}\right) \tag{A10}$$

Introducing equation (A10) into equation (A7):

$$\frac{S_s}{\rho^{fl} g}=\frac{1}{K+\frac{4}{3}G}\left\{\frac{\alpha}{K}\left[K+(1-\alpha)\frac{4G}{3}\right]+(K+\frac{4}{3}G)\left[\phi\left(\frac{1}{K_{fl}}-\frac{1}{K_s}\right)\right]\right\} \tag{A11}$$

Equation (A2) can be written as follows:

$$\frac{\alpha}{K}=\frac{1}{K}-\frac{1}{K_s} \tag{A12}$$

Introducing equation (A12) into equation (A11), developing the term $\dfrac{1}{K+\frac{4}{3}G}$ and considering the expression (A2) of Biot's coefficient, we obtain the *Green and Wang* [1990] equation:

$$S_s=\rho^{fl} g\left[\left(\frac{1}{K}-\frac{1}{K_s}\right)\left(1-\frac{4G(1-K/K_s)}{K+4/3G}\right)+\phi\left(\frac{1}{K_f}-\frac{1}{K_s}\right)\right] \tag{A13}$$

Consequently, our formulation of the specific storage coefficient (*i.e.* equation 6) is consistent with that of *Green and Wang* [1990].

Appendix B. Expression of the specific storage coefficient for a transversely isotropic rock

B.1 Assumptions

As for the isotropic case, an infinitely long and vertical borehole is drilled. A cylindrical coordinate system is considered. Axis $0\underline{z}$, which coincides with the axis of the borehole, is perpendicular to the bedding planes and constitutes an axis of symmetry (Figure 3). Cheng's formalism is used [*Cheng*, 1997]. As for the isotropic case, consider the following assumptions :

- the vector of displacement $\underline{u}$ is purely radial : $\underline{u} = u(r,t)\underline{e}_r$ where $\underline{e}_r$ is the radial unit vector of a cylindrical coordinate system.
- the boundary conditions far from the borehole wall, are written as follows:

  $P(r \rightarrow \infty, t) = P_0$ or $\partial P / \partial r = 0$ (hydraulic condition)

  $tr\underline{\underline{\varepsilon}} = 0$ (no volumetric strain) or $\sigma_r = \sigma_0$ (mechanical condition),

  tr : trace operator; $\underline{\underline{\varepsilon}}$: strain tensor; P : pore fluid pressure ; σ_r : radial component of the total stress tensor $\underline{\underline{\sigma}}$; σ_0 : initial lithostatic pressure.

B2. Constitutive equations

Two constitutive laws are introduced. The first is mechanical and is defined by:

$$\begin{Bmatrix} \sigma_{rr} \\ \sigma_{\theta\theta} \\ \sigma_{zz} \\ \tau_{r\theta} \\ \tau_{\theta z} \\ \tau_{rz} \end{Bmatrix} = \begin{bmatrix} M_{11} & M_{12} & M_{13} & 0 & 0 & 0 \\ M_{12} & M_{11} & M_{13} & 0 & 0 & 0 \\ M_{13} & M_{13} & M_{33} & 0 & 0 & 0 \\ 0 & 0 & 0 & M_{44} & 0 & 0 \\ 0 & 0 & 0 & 0 & M_{55} & 0 \\ 0 & 0 & 0 & 0 & 0 & M_{55} \end{bmatrix} \begin{Bmatrix} \varepsilon_{rr} \\ \varepsilon_{\theta\theta} \\ \varepsilon_{zz} \\ \gamma_{r\theta} \\ \gamma_{\theta z} \\ \gamma_{rz} \end{Bmatrix} - \begin{Bmatrix} \alpha \\ \alpha \\ \alpha' \\ 0 \\ 0 \\ 0 \end{Bmatrix} (P - P_0) \qquad \text{(B1)}$$

where

$$M_{11} = \frac{E(E' - Ev'^2)}{(1+v)(E' - E'v - 2Ev'^2)} \; ; \; M_{12} = \frac{E(E'v + Ev'^2)}{(1+v)(E' - E'v - 2Ev'^2)}$$

$$M_{13} = \frac{E'Ev'}{E'-E'v-2Ev'^2} \; ; \; M_{33} = \frac{E'^2(1-v)}{E'-E'v-2Ev'^2}$$

$$M_{44} = G = \frac{M_{11}-M_{12}}{2} = \frac{E}{2(1+v)} \; ; M_{55} = G'$$

σ_{rr}, $\sigma_{\theta\theta}$,σ_{zz}, $\tau_{r\theta}$, $\tau_{\theta z}$, τ_{rz}. are the components of the total stress tensor in the cylindrical coordinate system.

Moduli E, ν and G are, respectively the "drained" Young's modulus, "drained" Poisson's coefficient and shear modulus. Poroelastic properties α and M are Biot's coefficient in the x-y plane and Biot's modulus, respectively.

Symbol ' indicates poroelastic properties in the axis 0<u>z</u>, perpendicular to the bedding planes. Properties with no symbol ' are associated to the bedding planes. "Undrained" moduli are related to "drained" moduli by:

$$\begin{cases} M_{11} = M_{11}^u - \alpha^2 M \\ M_{12} = M_{12}^u - \alpha^2 M \\ M_{13} = M_{13}^u - \alpha\alpha' M \\ M_{33} = M_{33}^u - \alpha'^2 M \end{cases} \tag{B2}$$

The second hydromechanical constitutive law is [e.g., *Coussy*, 1995]:

$$\frac{m}{\rho_0^{fl}} = \frac{1}{M}(P - P_0) + \alpha\varepsilon_{rr} + \alpha\varepsilon_{\theta\theta} + \alpha'\varepsilon_{zz} \tag{B3}$$

where m is the liquid mass supply (the variation of fluid volume per unit reference volume), ε_{rr}; $\varepsilon_{\theta\theta}$ and ε_{zz} are the components of the strain tensor in the cylindrical coordinate system.

B.3 Field equations

In a cylindrical coordinate system, equilibrium equations are expressed by:

$$\begin{cases} \sigma_{rr,r} + \frac{1}{r}\tau_{\theta r,\theta} + \tau_{rz,z} + \frac{\sigma_{rr} - \sigma_{\theta\theta}}{r} = 0 \\ \tau_{\theta r,r} + \frac{1}{r}\sigma_{\theta\theta,\theta} + 2\frac{\tau_{r\theta}}{r} + \tau_{\theta z,z} = 0 \\ \tau_{rz,r} + \frac{1}{r}\tau_{\theta z,\theta} + \sigma_{zz,z} + \frac{\tau_{rz}}{r} = 0 \end{cases} \tag{B4}$$

Introducing equation (B1) into the equilibrium equations (B4), we obtain:

$$\begin{aligned} M_{11}\,\varepsilon_{rr,r} + M_{12}\,\varepsilon_{\theta\theta,r} + M_{13}\,\varepsilon_{zz,r} - \alpha\delta P_{,r} + \frac{1}{r}M_{44}\,\gamma_{r\theta} + \\ + M_{55}\,\gamma_{rz},_{z} + \frac{1}{r}(M_{11} - M_{12})(\varepsilon_{rr} - \varepsilon_{\theta\theta}) = 0 \end{aligned} \tag{B5a}$$

$$\begin{aligned} M_{44}\gamma_{\theta r,r} + \frac{1}{r}(M_{12}\varepsilon_{rr,\theta} + M_{11}\varepsilon_{\theta\theta,\theta} + M_{13}\varepsilon_{zz,\theta} - \alpha\delta P_{,\theta}) \\ + \frac{2}{r}M_{44}\gamma_{r\theta} + M_{55}\gamma_{\theta z,z} = 0 \end{aligned} \tag{B5b}$$

$$\begin{aligned} M_{55}\,\gamma_{rz,z} + \frac{1}{r}M_{55}\gamma_{\theta z,\theta} + M_{13}\varepsilon_{rr,z} + M_{13}\varepsilon_{\theta\theta,z} + M_{33}\varepsilon_{zz,z} \\ - \alpha\delta P_{,z} + \frac{1}{r}M_{55}\gamma_{rz} = 0 \end{aligned} \tag{B5c}$$

Under the condition of purely radial displacement i.e. $\underline{u} = u(r,t)\underline{e}_r$, the strain tensor may be simplified in the following way:

$$\underline{\underline{\varepsilon}} = \begin{pmatrix} \frac{\partial u}{\partial r} & 0 & 0 \\ 0 & u/r & 0 \\ 0 & 0 & 0 \end{pmatrix} \tag{B6}$$

where $tr\underline{\underline{\varepsilon}} = div\underline{u} = \frac{\partial u}{\partial r} + \frac{u}{r}$

By introducing components of the strain tensor in equations (A5a,b,c) and considering the pore fluid pressure P, one obtains the following differential equation :

$$M_{11}\frac{\partial^2 u}{\partial r^2} + M_{12}\frac{\partial}{\partial r}\left(\frac{u}{r}\right) - \alpha\frac{\partial P}{\partial r} + \frac{1}{r}(M_{11} - M_{12})\underbrace{\left(\frac{\partial u}{\partial r} - \frac{u}{r}\right)}_{r\frac{\partial}{\partial r}\left(\frac{u}{r}\right)} = 0 \tag{B7}$$

or

$$M_{11}\frac{\partial}{\partial r}\underbrace{\left(\frac{\partial u}{\partial r}+\frac{u}{r}\right)}_{tr\underline{\underline{\varepsilon}}}=\alpha\frac{\partial P}{\partial r} \tag{B8}$$

Integration of equation (B8) in one dimension r gives:

$$M_{11}tr\underline{\underline{\varepsilon}}=\alpha\,\delta P+C(t) \tag{B9}$$

where C(t) is an arbitrary function of time. The following boundary conditions

$\delta P(r\to\infty)=0\ (i.e. P=P_0)$ and $tr\underline{\underline{\varepsilon}}(r\to\infty)=0$ (no displacement and hence no strain) yields:

$C(t)=0,$

hence $tr\underline{\underline{\varepsilon}}=\dfrac{\alpha}{M_{11}}\delta P$ (B10)

The plane strain condition ($\underline{u}=u(r,t)\underline{e}_r$) gives $\varepsilon_{zz}=0$ and allows us to simplify the constitutive law

(B3):

$$\frac{m}{\rho_0^{fl}}=\frac{1}{M}\delta P+\alpha(\varepsilon_{rr}+\varepsilon_{\theta\theta})=\frac{1}{M}\delta P+\alpha tr\underline{\underline{\varepsilon}} \tag{B11}$$

By substituting equation (B10) into equation (B11):

$$\frac{m}{\rho_0^{fl}}=\frac{1}{M}\delta P+\alpha^2\frac{1}{M_{11}}\delta P=(\frac{1}{M}+\frac{\alpha^2}{M_{11}})\delta P \tag{B12}$$

Considering a Darcian flow, the fluid mass balance can be written in the form:

$$\frac{\partial}{\partial t}\left(\frac{m}{\rho_0^{fl}}\right)=\frac{k}{\eta}\nabla^2 P \tag{B13}$$

Combination of equations (B12) and (B13) yields:

$$\left[\frac{M_{11}+\alpha^2 M}{MM_{11}}\right]\delta\dot{P}=\frac{k}{\eta}\nabla^2 P \tag{B14}$$

Note that from equation (B2) $M_{11}^u=M_{11}+\alpha^2 M$, equation (B14) can be written in the form:

$$\rho_0^{fl} g \left[\frac{M_{11}^u}{MM_{11}} \right] \delta \dot{P} = \rho_0^{fl} g \frac{k}{\eta} \nabla^2 P \tag{B15}$$

Comparison of (B15) with equation (5) commonly used in hydrogeology gives the following expression:

$$\frac{S_s}{\rho_0^{fl}\, g} = \frac{1}{M} \frac{M_{11}^u}{M_{11}} \tag{B16}$$

Acknowledgments. This research was funded by the Centre National de la Recherche Scientifique (PNRH 99/35 Project; paper n° 307) and the Groupement pour l'étude des Structures Souterraines de Stockage (G.3S). Their support is gratefully acknowledged.

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Table and figure captions

Table 1. Sensitivity analysis for Callovo-Oxfordian Claystone (375-478 m deep) from the Paris basin

Table 2. Sensitivity analysis for Toarcian Claystone (692-772 m deep) from Paris basin.

Table 3. Sensitivity analysis for Domerian Claystone (782-855 m deep) from Paris basin

Table 4. Hydrogeological properties of Toarcian-Domerian formations [*Cellier*, 1998]

Table 5. Hydrogeological and mechanical parameters of the Toarcian formation of the Tournemire site.[a] *Boisson et al.* [1998]; [b] *Niandou et al.* [1997].

Figure 1. Biot's coefficient α as a function of porosity for Toarcian-Domerian claystones. Also shown for comparison are published values of α for other geological formations [*Rice and Cleary*; 1976, *Charlez*, 1997; *Fabre and Gustkiewicz*, 1997]. The dashed line represents the values given by Boutéca's model [*Boutéca and Sarda*, 1991].

Figure 2. Biot's coefficient α as a function of permeability for Toarcian-Domerian claystones. Also shown for comparison are published values of α for other geological formations [*Rice and Cleary*; 1976, *Charlez*, 1997; *Fabre and Gustkiewicz*, 1997]. The dashed line represents the values given by a logarithmic regression.

Figure 3. Definition sketch for a vertical borehole in a transversely isotropic rock.

Figure 4. Horizontal Biot's coefficient α as a function of permeability for the Tournemire claystone. Also shown for comparison are published values of α for other geological formations [*Rice and Cleary*; 1976, *Charlez*, 1997; *Fabre and Gustkiewicz*, 1997]. The dashed line represents the values given by a logarithmic regression.

Table 1. Sensitivity analysis for

Callovo-Oxfordian Claystone (375-478 m deep) from the Paris basin

Measured specific storage coefficient (m^{-1})[a]	Measured "undrained" Young's modulus E^u (GPa)[b]	Measured "undrained" Poisson's Ratio ν^{u} [b]	Measured Porosity ϕ (%)[b]	Calculated Biot's coefficient α		Calculated Biot's modulus M (GPa)		Calculated "drained" bulk compressibility K (GPa)		Calculated bulk modulus of matrix K_s (GPa)	
				Δ_a	Δ_r (%)	Δ_a (GPa)	Δ_r (%)	Δ_a (GPa)	Δ_r (%)	Δ_a (GPa)	Δ_r (%)
(10^{-7}-$3\ 10^{-6}$)	4.9	0.3	14	0-0.69	200	5.46-26.32	124	1.51-4.08	92	4.12-4.82	15.6
$1.6\ 10^{-6}$	(2.3-11)	0.3	14	0.25-0.7	95	7.69-9.36	20	1.42-4.56	105	1.9-15.27	156
$1.6\ 10^{-6}$	4.9	(0.17-0.4)	14	0.31-0.59	62	7.43-9.31	22	1.78-4.94	94	2.56-12.01	130
$1.6\ 10^{-6}$	4.9	0.3	(9-18)	0.37-0.42	13	7.69-8.36	8	2.58-3.02	16	4.48-4.8	7

Abbreviations Δ_a and Δ_r indicate absolute variation and relative variation (in %) respectively.
Measured values are mean values. Variation ranges of experimental values are given in brackets.
Considering the variation range (A-B), relative variation Δ_r is calculated by:

$$\Delta_r = \frac{2(B-A)}{(B+A)}$$

[a] *Cellier* [1998]
[b] *Ghoreychi* [1998]

Table 2. Sensitivity analysis for the

Toarcian Claystone (692-772 m deep) from the Paris basin

Measured specific storage coefficient (m^{-1})	Measured "undrained" Young's modulus E^u (GPa)	Measured "undrained" Poisson's Ratio ν^u	Measured Porosity ϕ (%)	Calculated Biot's coefficient α		Calculated Biot's modulus M (GPa)		Calculated "drained" bulk compressibility K (GPa)		Calculated bulk modulus of matrix K_s (GPa)	
				Δ_a	Δ_r (%)	Δ_a (GPa)	Δ_r (%)	Δ_a (GPa)	Δ_r (%)	Δ_a (GPa)	Δ_r (%)
$7.5\ 10^{-7}$ - $1.3\ 10^{-6}$	3.48	0.29	15	0.14-0.25	56	9.03-13.99	43	2.18-2.49	13	2.89-2.92	1
$1.\ 10^{-6}$	(2.6-7.3)	0.29	15	0.18-0.27	40	11.19-10.87	3	1.7-5.02	99	2.07-6.84	107
$1.\ 10^{-6}$	3.48	(0-0.34)	15	0.17-0.21	200	11.02-10.96	0.5	3.13-8.56	93	1.03-3.97	118
$1.\ 10^{-6}$	3.48	0.29	(13.6-16)	0.19-0.20	5	10.98-11.02	0.4	2.32-2.35	0.3	2.90-2.92	0.7

Abbreviations Δ_a and Δ_r indicate absolute variation and relative variation (in %) respectively.
Measured values are mean values. Variation ranges of experimental values are given in brackets.
Considering the variation range (A-B), relative variation Δ_r is calculated by :

$$\Delta_r = \frac{2(B-A)}{(B+A)}$$

Table 3. Sensitivity analysis for

Domerian Claystone (782-855 m deep) from the Paris basin

Measured specific storage coefficient $(m^{-1})^a$	Measured "undrained" Young's modulus E^u (GPa)	Measured "undrained" Poisson's Ratio ν^u	Measured Porosity ϕ (%)	Calculated Biot's coefficient α		Calculated Biot's modulus M (GPa)		Calculated "drained" bulk compressibility K (GPa)		Calculated bulk modulus of matrix K_s (GPa)	
				Δ_a	Δ_r (%)	Δ_a (GPa)	Δ_r (%)	Δ_a (GPa)	Δ_r (%)	Δ_a (GPa)	Δ_r (%)
$9.1\ 10^{-7}$	(2.6-10.6)	0.21	13.8	0.16-0.27	51	11.53-11.96	4	1.19-5.28	126	1.42-7.19	134
$9.1\ 10^{-7}$	5.44	(0.05-0.42)	13.8	0.17-0.35	69	11.49-12.05	5	1.66-9.85	142	2.02-15.18	153
$9.1\ 10^{-7}$	5.44	0.21	(13.2-14.4)	0.199-0.192	4	11.56-11.62	0.5	2.667-2.698	1	3.33-3.34	0.3

Abbreviations Δ_a and Δ_r indicate absolute variation and relative variation (in %) respectively.
Measured values are mean values. Variation ranges of experimental values are given in brackets.
Considering the variation range (A-B), relative variation Δ_r is calculated by :

$$\Delta_r = \frac{2(B-A)}{(B+A)}$$

[a]Variation range is not given by *Cellier* [1998]

Table 4. Hydrogeological properties of the Toarcian-Domerian formations [*Cellier*, 1998]

Formation	Depth (m)	Permeability (m/s)	Specific storage coefficient (m^{-1})	Calculated dimensionless parameter a
Toarcian	692-772	$3\ 10^{-13}$- 210^{-12}	$7.5\ 10^{-7}$-$1.3\ 10^{-6}$	$2.\ 10^{-2}$-$6\ 10^{-2}$
Domerian	782-855	$6.5\ 10^{-13}$	$9.1\ 10^{-7}$	$3\ 10^{-2}$

Table 5. Hydrogeological and mechanical parameters of the Toarcian formation from the Tournemire site.

[a] *Boisson et al.* [1998]; [b] *Niandou et al.* [1997].

Test location from tunnel [a]	Hydrogeological parameters[a]		Mechanical parameters [b]			
	Horizontal permeability (m/s)	Specific storage coefficient (m^{-1})	Vertical "undrained" Young's modulus E'^u (GPa)	Horizontal "undrained" Young's modulus E^u (GPa)	Vertical "undrained" Poisson's " coefficient ν'^u	Horizontal "undrained" Poisson's coefficient ν^u
Test 2 61.5 - 63 m	$1.3\ 10^{-12}$	$3\ 10^{-7}$	6.58	24.58	0.23	0.13
Test 3 146.3 - 147.8 m	$6.7\ 10^{-14}$	$6.\ 10^{-7}$	7.44	24.27	0.23	0.13
Test 4 103.6 - 105.1 m	10^{-13}	$7.\ 10^{-7}$	7.10	24.92	0.23	0.13
Test 5 41.5-43.0 m	$2.3\ 10^{-11}$ $-1.3\ 10^{-12}$	7.10^{-7}- 1.10^{-6}	6.58	24.40	0.22	0.13

Table 6. Estimates of poroelastic properties of Toarcian claystone

from Tournemire site

Test	Porosity (%)	Calculated poroelastic properties			
		Horizontal Biot's coefficient α	Vertical Biot's coefficient α'	Biot's modulus M (GPa)	Bulk modulus of the matrix K_s (GPa)
2	2.2	0.21	1.66	30.59	36.03
	2.4	0.22	1.56	30.68	36.41
3	2.2	0.49	4.02	18.57	37.09
	2.4	0.55	3.67	19.48	37.94
4	2.2	0.49	4.63	16.49	36.02
	2.4	0.54	4.31	17.14	36.62
	2.2	0.46	4.59	16.29	35.08
5	2.4	0.51	4.31	16.84	35.62
	2.2	0.47	7.96	10.51	34.08
	2.4	0.53	7.70	10.72	34.31

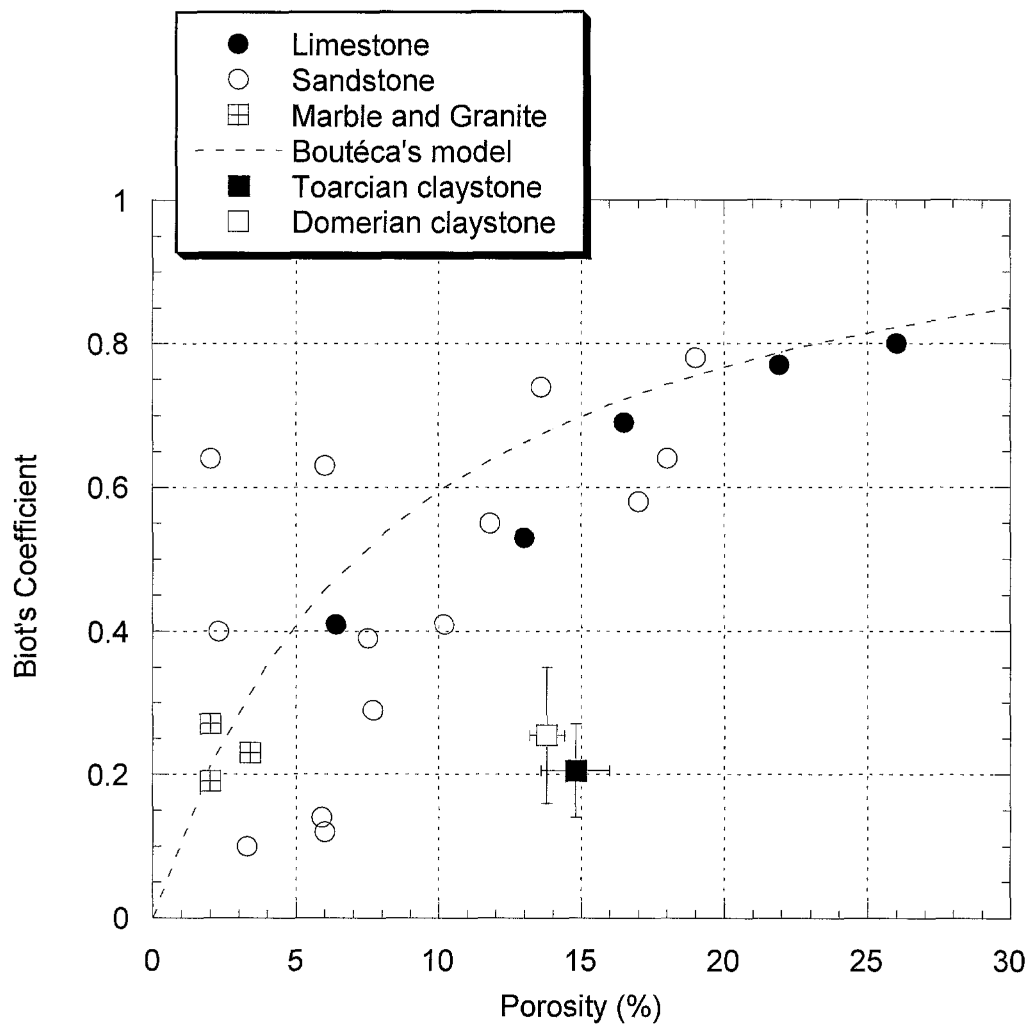
Limestone
Sandstone
Marble and Granite
Boutéca's model
Toarcian claystone
Domerian claystone
1
0.8
0.6
0.4
0.2
0
Biot's Coefficient
0
5
10
15
20
25
30
Porosity (%)

Fig. 1

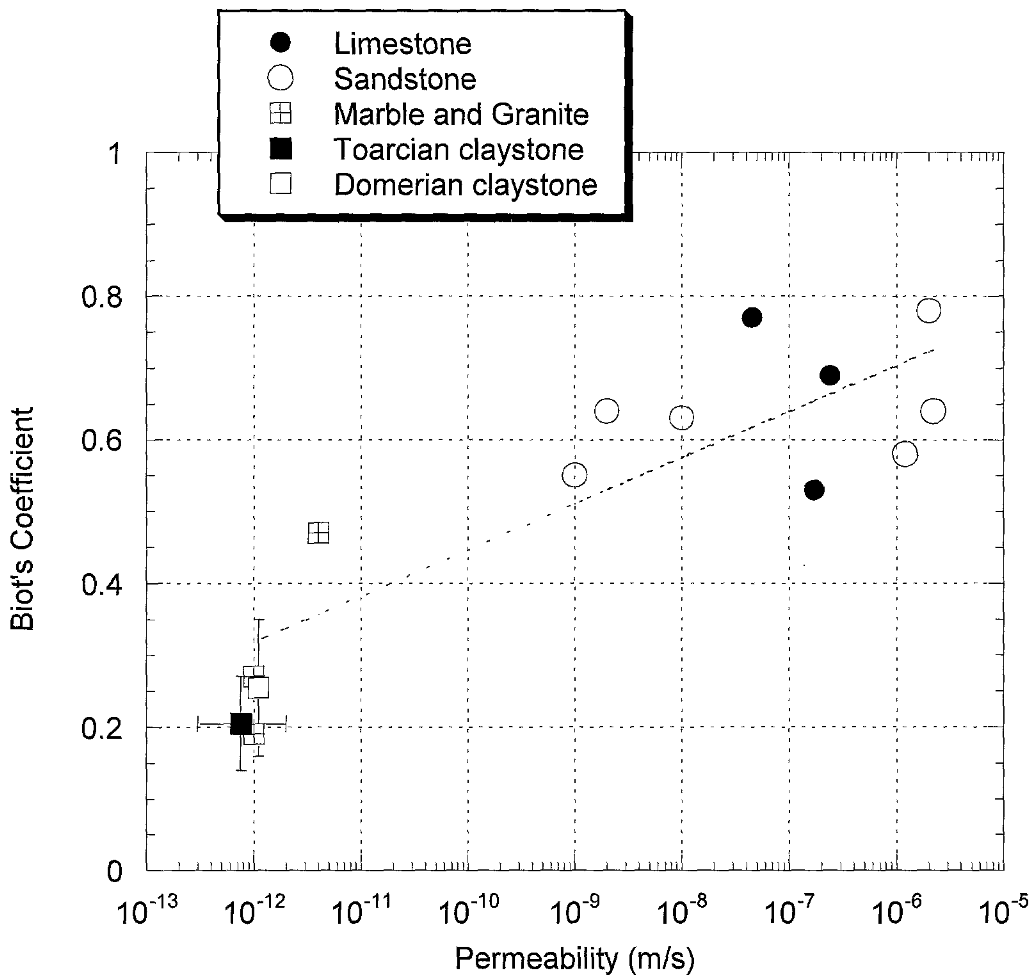
Limestone
Sandstone
Marble and Granite
Toarcian claystone
Domerian claystone
Biot's Coefficient
0
0.2
0.4
0.6
0.8
1
10-13
10-12
10-11
10-10
10-9
10-8
10-7
10-6
10-5
Permeability (m/s)

Fig.2

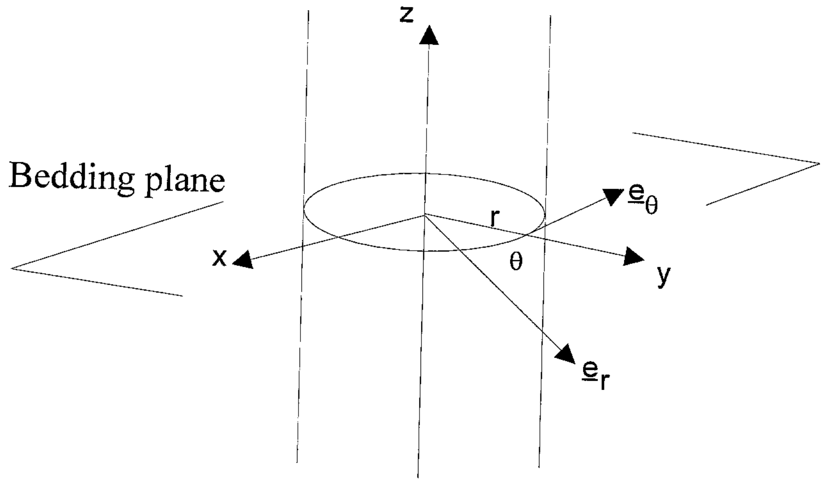

Fig. 3

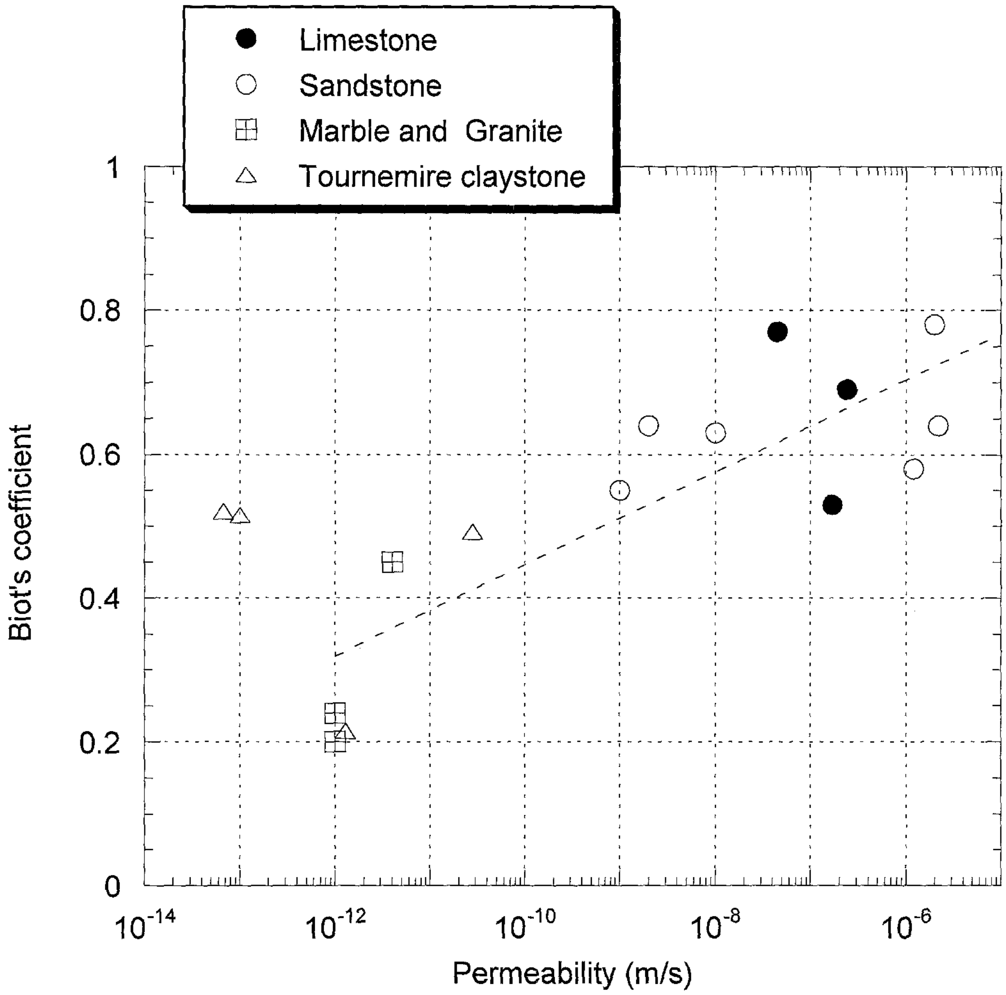
Limestone
Sandstone
Marble and Granite
Tournemire claystone
1
0.8
0.6
0.4
0.2
0
Biot's coefficient
10^{-14}
10^{-12}
10^{-10}
10^{-8}
10^{-6}
Permeability (m/s)

Fig. 4