Complete In situ Hydro-mechanical Characterization of a Confined Coastal Aquifer Bounded by an Active Impervious Fault

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Abstract. Earth tides, oceanic load and barometric effects affect pore pressure variations that are monitored in the 1000 m deep Aig10 well, located some 500 m off the southern shore of the Corinth Gulf, in western Greece. The open-hole section of the well intersects two confined aquifers separated by the active, 15 km long, Aigion normal fault. The high quality of the pressure signal has been taken to advantage for separating effects of these three loading functions. Amplitudes and phases of the respective response coefficients provide all information required for a complete hydro-mechanical characterization of the system.

The lower aquifer is karstic and its tidal response is analysed according to the theory of homogeneous porous media. The results are consistent. The elastic mechanical behavior is described by the two undrained coefficients: the Poisson ratio $\nu_u \sim 0.2$ and the bulk incompressibility $K_u \sim 20$ GPa. The coupling between fluid pressure and rock deformation is expressed through the Skempton coefficient $B \sim 0.8$, and the storativity $S = (7.7 \pm 2) \times 10^{-7}$ m$^{-1}$. With a coefficient $\frac{d\rho}{d\epsilon} = B K_u \sim 17$, GPa and a pressure sensor able to resolve pressure steps of 10 Pa, strain steps as small as $5 \times 10^{-10}$ can be detected.

Full analytical solutions, for complex realistic loading or simple Boussinesq strip loading have been developed for modeling the aquifer pressure response to oceanic load. The latter method leads to a closed-form solution for estimating the hydraulic conductivity of the karst $C = (2 \pm 1) \times 10^{-5}$ m/s. The hydraulic diffusivity $D$ is approximately equal to 20 m$^2$/s. This analy-
sis also indicates that the fault intersected by the well is impervious much
below the interval where clay smearing occurs.
1. Introduction

In order to study in situ the interaction between fluid pressure and fault movements, a 1000 m deep borehole has been drilled through the active Aigio Fault, in the Corinth Rift (western Greece). This fault is impervious and separates two distinct artesian aquifers. This hydrogeological system is dominated by the lower aquifer, which is karstic. The borehole is closed by a packer set near the surface and the well head pressure is monitored with high-resolution sensors. Our objective is to relate the recorded pressure to the mechanical deformation supported by the medium and to apply tidal analysis for evaluating relevant characteristic parameters [Bredehoeft, 1967].

After a short presentation of the hydraulic configuration and of the data acquisition system, we present a tidal analysis method for separating the respective effects of Earth tides from those of oceanic loading. From the amplitude of these coefficients, we estimate poroelastic parameters for the main aquifer. To interpret the phase lag of the signal, a general analytic 1-D solution is developed for evaluating the hydraulic response of a confined bounded aquifer to the oceanic load. Applied to the Aigio results, this development helps constraining the location of the aquifer boundaries and highlights the imperviousness of the active Aigio fault.

2. Presentation of the data

2.1. Hydrogeological setting

The 120 km long Corinth Gulf that separates Peloponnesus from mainland Greece is one of the most active continental rift in the world. The 1.5 cm/yr opening rate generates an intense microseismic activity [Lyon-Caen et al., 2004] with some occasional larger
magnitude earthquakes such as the $M_s = 6.2$ Aigio earthquake, which occurred on June 1995 [Bernard et al., 1997]. Although the main shock mobilized a shallow dipping off-shore fault, it induced a 3 cm vertical offset on the $60^\circ$ dipping, 15 km long Aigio fault [Koukouvelas, 1998].

The Aigio fault has been chosen to be the site of the Deep Geophysical LABoratory (DGLab), installed within the framework of the European-Union funded Corinth Rift Laboratory (CRL). DGLab mainly consists of a 1000 m-deep instrumented borehole [Cornet et al., 2004b] whilst CRL involves a multidisciplinary set of surface sensors for monitoring the regional deformation process [Cornet et al., 2004a]. The geological information collected during the drilling of the borehole is sketched on figure 1.

The syn-rift sediments are encountered down to 490 m. Below the 130 m thick upper recent clay deposit (less than 35 kyr), these sediments involve roughly 250 m of permeable conglomerates deposited above about 110 m of marine and lacustrine impervious clayish material. Below 490 m, the borehole enters the pre-rift Pindos nappe that involves formations deposited in a benthic environment and composed by an alternation of clay, radiolarite and limestone. As the borehole is cased down to 708 m, the borehole pressure is controlled only by the two deepest aquifers intersected by the well below 700 m. The first 50 m of the open-hole section corresponds to a fractured limestone aquifer. Hydraulic testings performed while the borehole had not yet crossed the Aigio fault enabled to constrain its hydraulic conductivity to $10^{-7}$ m/s, corresponding to a permeability of $10^{-14}$ m$^2$ [Giurgea et al., 2004]. It is artesian with an initial overpressure estimated to $P_u = 0.5 \pm 0.1$ MPa.
The second aquifer is a karstic limestone aquifer. From 770 m to 1000 m, borehole logging shows cavities some of them reaching 1 m. The aquifer is also artesian with an overpressure \( P_L = 0.9 \pm 0.1 \text{ MPa} \). A hydraulic test has been performed in September 2003 to estimate its hydraulic conductivity. The production rate was estimated at \( 225 \pm 25 \text{ m}^3/\text{hr} \), so that flow was turbulent. The corresponding head loss (close to 0.9 MPa), as evaluated from Colebrook and White equation [Lencastre, 1987], is found to be the factor limiting the flow rate. Hence, the value derived from Thiem-Dupuit equation [Cornet et al., 2004b] is an underestimate of the effective hydraulic conductivity \( C \geq 5 \cdot 10^{-6} \text{ m/s} \), which corresponds to a \( 5 \cdot 10^{-13} \text{ m}^2 \) permeability.

The storativity of the bottom aquifer could not be determined from this test since no significant variation in outflow was observed. The difference between the overpressures in the two aquifers indicates that the Aigio fault is impervious. The aim of the borehole is to detect eventual pressure variations in the aquifers because of movements along the Aigio fault. To interpret them, we must answer two questions. What is the significance of the measured pressure? To which aquifer are related the pressure data?

2.2. Permanent instrumentation

On September 2003, after the borehole was let to produce water, it was sealed with an inflatable packer set close to the well head. This provided means to monitor downhole water pressure variations. But the sealing of the well disturbed the downhole hydraulic system since the two previously independent aquifers got connected through the open section of the well [Doan and Cornet, 2005b]. Pressure has been monitored by a Paroscientific Digiquartz sensor, with a 10 Pa resolution and a 1/8 Hz sampling rate.
An example of the recorded signal is presented in figure 2. It shows very clear semi-diurnal variations with a 2.5 kPa amplitude, acquired with a precision better than 1%. The average value around which the signal oscillates varies from 0.870 to 0.854 MPa during the first three months recording [Doan and Cornet, 2005b]. The slow drop in pressure is caused by a fluid flow between the two aquifers. However, the pressure recorded remains very close to the initial pressure of the karstic aquifer.

A simple quasi-stationary regime can explain this fact. As the new pressure equilibrium requires more than three months, a permanent regime approximates satisfactorily the daily pressure variations. Following the figure 1 notation, the measured pressure can be approximated as:

\[ P = \frac{\left( \frac{1}{K_L} + \frac{1}{K_B} \right) P_U + \frac{1}{K_U} P_L}{\frac{1}{K_L} + \frac{1}{K_B} + \frac{1}{K_U}} \]  

(1)

where \( K_i \) denotes a hydraulic conductance, and \( P_i \) the pressure far from the borehole in each aquifer. The equivalent hydraulic conductance for the two aquifers are given by the Dupuit equation. For instance, the conductance of the lower aquifer

\[ K_L = \frac{2 \pi c_L H_L}{\rho_f g \log \left( \frac{R_\infty}{r_b} \right)} \]  

(2)

depends on its hydraulic conductivity \( c_L \), its height \( H_L \) and on its influence radius \( R_\infty \) that delimits the region disturbed by the flow in the well.

The borehole radius \( r_b = 8.25 \) cm is the same for both aquifers. A potential problem lies in the choice of the influence radius. Its value is expected to vary between 100 m and 10 km. As it intervenes in a logarithm function, it only changes the value of the conductance by a factor of 2. This is not enough to compensate the disparity of hydraulic conductivity and height between the two aquifers. Indeed, if we take \( R_\infty = 100 \) m for both aquifers, we get...
a conductance of $3 \times 10^{-10}$ m$^3$/s/Pa for the upper aquifer and of $3 \times 10^{-7}$ m$^3$/s/Pa only for the lower aquifer. The conductance of the borehole can be estimated from Poiseuille equation. Its value of $3 \times 10^{-2}$ m$^3$/s/Pa is very high, so that it does not intervene in equation 1, and the pressure we measure is equal to $P \sim 0.001 P_U + 0.999 P_L$. Thus, the recorded pressure is dominated by the response of the bottom aquifer. Consequently, the tidal variations observed in figure 2 reflect the response of the karstic aquifer to tidal excitations.

Yet, this aquifer remains very poorly known since the hydraulic test could not provide its storativity. Moreover, the metric size of its pore space prevents the extrapolation of laboratory data to field conditions so that the hydraulic and mechanical parameters that characterize the karst must be retrieved from in-situ observations.

3. Tidal variation analysis

3.1. Quantifying the origin of the tidal variations

The recorded pressure variations have three main sources: the solid Earth tides, the loading induced by the oceanic tides, the barometric pressure variations. To interpret these variations, we need to quantify all these three effects. Let respectively $p_{\text{ter}}$, $p_{\text{oc}}$ and $p_{\text{atm}}$ be the corresponding pressure variations.

We use the PREDICT module of the ETERNA v3.3 tidal analysis package\cite{Wenzel1996} for evaluating earth tides effects. ETERNA is the software distributed by the International Center of Earth Tides (ICET). The strain predicted by ETERNA follows a dilatational convention, as a compression is depicted by a negative strain. The time series computed with ETERNA at a depth of 700 m on the location of Aigio borehole are shown on figure 4. It does not take into account the effect of oceanic loading.
The proximity of the borehole from the seashore requires a good knowledge of the oceanic water level variations. As the Gulf is only 120 km long, we rely on local sea temporal variations as recorded by three permanent tide gages in Trizonia Island, Galaxidi and Egira (figure 3). [Boudin, 2004] has shown that the three permanent tide gauges have identical tidal variations.

Barometric pressure is monitored in Temeni, about 5 km east to the borehole. Despite the distance, we assume that its pressure data describes barometric variation at the borehole location. Let us note that barometric loading is also taken into account in the data of the tide gages of Trizonia and Aigio, since these sensors are barometers measuring the total load induced by the atmospheric pressure and the load of the water column.

### 3.2. Linear regression in time domain

The high sampling rate of the pressure data, together with the high sensor resolution makes the frequency analysis relatively inefficient. Further, working directly in the time domain solves difficulties raised by the multiple short interruptions that resulted in irregular sampling. Assuming a linear relationship between excitation sources and the respective pressure variations, we simply try to determine the coefficient $c_{oc}$, $c_{\epsilon}$ and $c_{atm}$:

$$p(t) = c_{oc} p_{oc}(t) + c_{\epsilon} p_{ter}(t) + c_{atm} p_{atm}(t)$$ (3)

Let us note that tide gage data $p_{oc}$ is total pressure, so that the coefficient we derive from the ”atmospheric” $c_{atm}$ only partially accounts for the sensitivity to barometric pressure. With a 1/8 Hz sampling rate, more than 10,000 data are generated a day. Resampling 1 point per minute still leaves 1440 data per day for constraining the inversion.
For the time period spanning the end of November to the beginning of December 2003, the best fit is obtained for $c_\epsilon = -17.6$ GPa/str, $c_{oc} = 27.4$ Pa/cm and $c_{atm} = 0.23$. The relative effect of Earth tides, of atmospheric loading and of oceanic tides are respectively 49%, 4% and 47%. Visual observation of figure 5 shows that the fitting is satisfactory. The quality of the fit is quantified by the determination coefficient $R^2 = 1 - \|p_{obs} - p_{fit}\|^2/\|p_{obs} - \bar{p}_{obs}\|^2$ that reaches a value slightly greater than 98.7% for figure 5.

Another test performed during May 2004 yields similar results: $c_\epsilon = -16$ GPa and $c_{oc} = 28$ Pa/cm, suggesting that these coefficients are stable. This good fit implies that the eventual phase lag is small for both semi-diurnal and diurnal frequencies. This conclusion is certainly valid for Earth tides but less evident for oceanic load whose semi-diurnal component is predominant. As in the previous paragraph, we would like to refine the inversion by separating both types of variations. This can be performed with the PRE-DICT program that was applied for computing the theoretical semi-diurnal and diurnal Earth tides. The prediction is performed by using again the coefficients obtained by tidal analysis, i.e. those given in table 1.

Inverting equation 3, we used directly the recorded signals, thus including both diurnal and semi-diurnal variations.

To detect phase lag, we shifted the dates of the source signals by steps of 5 minutes before performing the linear regression. A positive time offset $\Delta t$ applied on the time labels of the records of a source signal is equivalent to an equivalent time lag in response of the output signal regarding this source. The best phase shifts are determined by maximizing the determination coefficient $R^2$. 
The phase lag for both diurnal and semi-diurnal Earth tides are negligible. The result for the tidal load is displayed in figure 6 for the period covering November 2003 and figure 7 for May 2004. The oceanic load is mainly semi-diurnal, so that its pattern is mainly vertical, with a maximum correlation for $\Delta t_{1/2\text{diurne}} \sim 20$ minutes. The diurnal component slightly modulates the shape of the iso-$R^2$ lines, where $\Delta t_{\text{diurne}} \sim -100$ minutes. The maximum of the determination coefficient is obtained in November 2003 for an offset in time label $\Delta t_{1/2\text{diurne}} \sim 10$ minutes and $\Delta t_{\text{diurne}} \sim -60$ minutes. For May 2004, the coefficients become $\Delta t_{1/2\text{diurne}} \sim -10$ minutes and $\Delta t_{\text{diurne}} \sim -70$ minutes.

The semi-diurnal component is only slightly time-shifted, as the best regressions were obtained for a time shift comprised between -10 minutes and 15 minutes. This involves a phase shift in pressure response with a maximum absolute amplitude $d\psi_2 = 15/(12 \times 60) \times 360^\circ = 7.5^\circ$.

For the diurnal tides, the scatter is greater, which is logical since their amplitude is smaller. We have a time correction of $-70 \pm 30$ minutes, so that the complex response coefficient of the borehole pressure to diurnal oceanic tides is $d\psi_2 = +(70 \pm 30)/(24 \times 60) \times 360^\circ = 17.5^\circ \pm 7.5^\circ$. One notices that this component is in advance relative to the source excitation, a point that is discussed in section 5.3.

This linear regression can be performed on short time spans, so that long term changes in the behavior of the hydraulic system associated with the Aigio fault may be detected. Results are consistent with those derived from a spectral analysis [Doan, 2005].

4. Poroelastic coefficients

The faint phase lags determined from tidal calibration hints that the response of the aquifer to this load is elastic. The elastic response of porous media is characterized by four
coefficients: the two coefficients which appear in classical elasticity, a coefficient coupling deformation to fluid pressure and a storage coefficient for characterizing the increment of fluid volume in the pore space due to an increase in fluid pressure. In addition, the permeability must be evaluated for quantifying fluid flow through the medium.

Tidal wavelengths are a fraction of the Earth perimeter, and reach several thousands of kilometers, much larger than the aquifer expansion. The description of the karst as a continuous medium is therefore justified for tidal excitation. Moreover, the large wavelength also generates a uniform deformation inside the aquifer. As no pressure gradient is induced, the rock responds to Earth tides in an undrained manner and the response to Earth tides can be quantified as the tidal efficiency

$$\frac{\partial P}{\partial \epsilon} \bigg|_{\text{undrained}} = -BK_u$$

where $B$ is the Skempton coefficient describing the undrained increase in pressure to an applied confining stress and $K_u$ is the undrained bulk modulus of the poroelastic material. From the tidal analysis, we get $BK_u = 17 \pm 1 \text{ GPa}$.

The coefficient relative to the oceanic load is not as simple to interpret since the loading is not uniform inside the aquifer. This will be discussed in the following section.

The response to barometric loading could not be retrieved directly from tidal analysis. Indeed, the amplitude of the diurnal atmospheric pressure variations are small, equal to about 200 Pa. As the barometric efficiency is inferior to 1, the resulting influence is less than 1% of the recorded tide. The barometric effect is more clearly felt during the large depressions of the winter. Figure 8 compares the variations of barometric pressure to the detided data recorded in the well. The two oscillations exhibit similar pattern and evolve jointly. However, the fit is not as clean as in figure 5, and a large uncertainty
in the determined barometric efficiency remains. Therefore, we assign to the barometric efficiency a value \( \gamma = 0.3 \pm 0.1 \). The coefficient \( c_{atm} \) found during the tidal analysis in time domain falls within this interval. Within the framework of poroelastic theory, it is interpreted as

\[
\frac{\partial P}{\partial P_{atm}}\Bigg|_{\epsilon_{hor}=0, undrained} = \frac{B}{3} \frac{(1 + \nu_u)}{(1 - \nu_u)}
\]

where \( \nu_u \) is the undrained Poisson ratio.

The Vertical Seismic Profile (VSP) performed by [Naville et al., 2004] provides some additional information, namely the velocities of the P waves \( (V_P \sim 5000 \text{m/s}) \) and S waves \( (V_P \sim 2000 \text{m/s}) \). However, these parameters involve another unknown, the rock density, but they help to ascertain values derived from the tidal analysis as discussed hereafter.

4.1. Consistency test without any simplifying assumption

The seismic waves generated during VSP tests are supposed to occur so rapidly (frequency domain larger than 1 Hz) that no fluid moves during their transmission. The product \( \rho V_P^2 \) is related to the undrained uniaxial modulus \( K_u^{(u)} \). From [Wang, 2000] (eq.3.56), this quantity equals:

\[
\rho V_P^2 = \frac{\partial \sigma_{kk}}{\partial \epsilon_{33}}\bigg|_{\epsilon_{hor}=0} = \frac{3 K_u (1 - \nu_u)}{1 + \nu_u} = \frac{B K_u}{\gamma}
\]

We use here the same notation as [Wang, 2000]. The volumetric mass of the bottom porous limestone is \( \rho = 2600 \pm 1000 \text{kg/m}^3 \). The resolution is not very good but the value is within the proper order of magnitude.
The ratio $V_P/V_S$ involves the undrained Poisson ratio $\nu_u$. Indeed, the product $\rho V_S^2$ is equal to the shear modulus $G$, so that $V_P^2/V_S^2 = \frac{2(1-\nu_u)}{1-2\nu_u}$. One therefore obtains the value

$$\nu_u = \frac{(V_P^2 - 2V_S^2)}{2(V_P^2 - V_S^2)} \sim 0.4 \quad (7)$$

As the seismic waves velocity were extrapolated from borehole sonic logs [Naville et al., 2004], this large value denotes a problem in the scaling of the seismic properties from the centimetric to the decametric size. We will show in a latter section that the response to oceanic tides provides means to get large-scale Poisson ratio, with a more satisfactory value. Inserting the solution 7 into equation 5 enables to retrieve the value of the Skempton coefficient $B = 0.38 \pm 0.13$. Using equation 4, we estimate also the undrained bulk modulus $K_u \approx 46 \pm 17$ GPa.

One parameter is still missing, namely the storativity of the porous medium. This parameter is important since it enables to derive the hydraulic diffusivity, an essential factor in the modeling of the pressure anomalies that might arise from seismic events [Doan, 2005]. To obtain this information, we have to formulate additional assumptions.

### 4.2. Hypothesis of an incompressible matrix

We suppose that the rock matrix is incompressible. The equation that yields the tidal and barometric efficiency are simplified as [Wang, 2000]:

$$BK_u = \frac{K_f}{\phi} \quad (8)$$

$$\gamma = \frac{K_f}{K_f + \phi K_v} \quad (9)$$

where $K_v$ is the drained uniaxial compressibility.

$$\phi = \frac{K_f}{BK_u} = 0.13 \pm 0.01 \quad (10)$$

$$K_v = BK_u \frac{1 - \nu_u}{\gamma} = 40 \pm 20 \text{ GPa} \quad (11)$$
One verifies that $K_s \gg K_f$, so that the hypothesis of incompressible matrix is at least self-consistent. The modulus values are among the high values listed by [Mavko et al., 1998]. Wang [2000] also gives the uniaxial specific storativity:

$$\frac{S}{\rho_f g} = \frac{\phi}{K_f} + \frac{1}{K_v} \quad (12)$$

which may be computed as $S = (7.7 \pm 2) \cdot 10^{-7} \text{m}^{-1}$. The hydraulic storativity we derived is an hybrid value between two extreme cases. The storativity $S_\epsilon$ at constant strain ($\epsilon_{kk} = 0$) is smaller than the storativity $S_\sigma$ at constant stress ($\sigma_{kk} = 0$). Therefore, $S_\epsilon \leq S \leq S_\sigma$, with the relationship $S_\epsilon = S_\sigma (1 - \alpha B)$, where $\alpha$ is the Biot coefficient, equal to 1 in the case of a stiff matrix. With the value of $B$ previously found, we conclude that the uniaxial storativity approximates the two other storativity values with a precision at least equal to 20%. Given the large uncertainty surrounding the value of $S$, we use the value obtained by equation 12 whatever the mechanical constrain applied on the medium.

So far, we have dealt only with information provided by the amplitude of the response coefficients. We turn now to phase shift analysis for evaluating the permeability.

5. Phase lag induced by oceanic loading

Hsieh et al. [1987] relate the delay of an aquifer response in an open well to the necessity of transferring large volumes of water for converting the overpressure induced by the aquifer deformation into change in water level. But only a negligible flow is involved in closed wells so that no phase lag is associated with Earth tides, as observed in the Aigio well. The volume of the well is $V_b \sim 23 \text{m}^3$, so that the volume of water needed to accommodate the overpressure $dP = 100 \text{Pa}$ is only $\frac{V_b dP}{K_f} \sim 10^{-6} \text{m}^3$, whereas the volume necessary to raise the water level to an equivalent amount of $dh = 10 \text{cm}$ is
\[ \pi r_b^2 dh \sim 2.3 \times 10^{-4} \text{ m}^3, \] as the borehole inner diameter \( 2r_b \) is 6.3/4. Therefore we concentrate on phase shifts associated with oceanic tidal loads.

### 5.1. van der Kamp’s model

The effect of cotidal oceanic load has been analytically studied for various configurations of the aquifer\[ Li \text{ and Jiao}, 2003 \]. However, the only previous publication we found relevant to the case of a deep confined aquifer is that by \textit{van der Kamp} [1972]. He supposes the aquifer is infinitely long and penetrates below the ocean. It is a unidimensional model, for which the pressure diffusion is simplified into:

\[
\frac{\partial p}{\partial t} - B \frac{\partial \sigma}{\partial t} = D \frac{\partial^2 p}{\partial x^2}
\]  

(13)

where \( D \) is the hydraulic diffusivity. The load function \( \sigma = \sigma_{kk}/3 \) is an Heaviside function passing suddenly from \( \rho_f g h_0 \cos(\omega t) \) to 0 when crossing the shoreline. The expected response of the aquifer can be computed in the complex variables formalism to give:

\[
P(x, t) = \frac{B h_0}{2} e^{-x/\delta} \sin(\omega t - x/\delta)
\]  

(14)

where \( \delta = \sqrt{\frac{2D}{\omega}} \) is the skin depth beyond which the oceanic load becomes negligible. In particular, the permeability deduced from the semi-diurnal phase shift \( d\psi_2 = 10^\circ \) is \( D = \left( \frac{x}{d\psi_2} \right)^2 \frac{\omega}{2} \). The distance \( x \) separating the borehole to the shoreline is taken as \( x = 500 \text{ m} \), so that we get \( D \sim 600 \text{ m}^2 \text{s}^{-1} \), from which the hydraulic conductivity is found to be \( 5 \cdot 10^{-4} \text{ m/s} \) (equivalent to a permeability of \( 5 \cdot 10^{-11} \text{ m}^2 \)). This is very large a value, that is likely to result from too simple a modeling of the oceanic load.

### 5.2. Boussinesq’s load

Solution for a general load function approximated by its Taylor expansion is discussed in appendix B. We discuss here the Boussinesq loading function. With this loading function
[Craig, 1987], we observe that the Heaviside function does not fit well the effective loading profile. This model supposes a uniform load at the surface along the Gulf, supposed to be infinitely long along the east-west direction, and with a uniform width $W = 10 \text{ km}$, as described in figure 9. With these figure conventions, we get the quantity:

$$\sigma_{xx} = \frac{P_0}{\pi} \left( \alpha + \sin \alpha \times \cos (\alpha + 2\beta) \right)$$  \hspace{1cm} (15)

$$\sigma_{zz} = \frac{P_0}{\pi} \left( \alpha - \sin \alpha \times \cos (\alpha + 2\beta) \right)$$  \hspace{1cm} (16)

As the Gulf is supposed to be infinitely long, plane strain condition prevails ($\epsilon_{yy} = 0$). Hence, the west-east stress component induced by oceanic load is $\sigma_{yy} = -\nu_u(\sigma_{xx} + \sigma_{zz})$.

The load to be used in equation 13 is:

$$\frac{\sigma_{kk}(x, z)}{3} = \frac{1 - \nu_u \rho_f g h}{3} \times \frac{\pi}{2} \left( \tan \left( \frac{z}{x} \right) - \tan \left( \frac{z}{x + W} \right) \right)$$  \hspace{1cm} (17)

The resulting load for a constant depth $z_0 = 1000 \text{ m}$ is presented in figure 10. We find a factor of about 0.25, similar to the coefficient found from the time signals inversion. The borehole, located 500 m away from the seashore is within the transition zone between the loaded and the unloaded region. The sharp transition of the van der Kamp [1972] model is therefore invalid for the case of the Aigio borehole. Figure 10 indicates that the smallness of phase lag to oceanic load is paradoxical. Contrary to the van der Kamp model, the large gradient of the loading function is expected to be compensated by a fluid flow governed by Darcy law. As a diffusion equation controls the response of the aquifer, we intuitively expect a significant phase lag.

5.3. Case of a semi-infinite ocean

We compute now from the Boussinesq loading function the expected phase delay, depending on the position of the borehole, as described in appendix A.
We use a graphical representation of the formula, to determine the effect of the boundary positions onto the phase shift of the pressure response to ocean load. Rather than parameterizing the map with the individual location of the northern \( x_N \) and southern \( x_S \) boundaries, we will use an equivalent set of parameters: the north-south extension of the aquifer \( L \) and the relative distance of the borehole to the northern aquifer boundary \( \lambda = (x - x_N)/L \). As the Skempton coefficient retrieved from seismic data is not satisfactory, we assume \( B = 1 \), which does not change the phase map as it is only a scaling factor (figure 13).

We observe in this figure several qualitative observations consistent with the Taylor expansion discussion (appendix B). For instance, we find a null-phase shift for a small aquifer extension \( L \). Also, when the borehole is close to an aquifer boundary (\( \lambda = 0 \) or \( \lambda = 1 \)), phase lags become important as suggested by the exponential function in equations B1 and B2. Finally, the heterogeneous loading of figure 10 implies a non-null phase shift in the case of the infinite medium, here approximated by the case of a large aquifer (large \( L \)), when the borehole is far from the boundaries (\( \lambda \sim 0.5 \)).

As experimental evidence shows that the Aigio fault acts as a hydraulic barrier, at least in the vicinity of the AIG10 well, this fault constitutes the northern boundary of the karstic aquifer, and the actual configuration corresponds to a small \( \lambda \). We therefore expect that the aquifer responds in advance to the oceanic load excitation. This is indeed the case for both diurnal and semi-diurnal observations.

Also, we find an amplitude of \( dp/d\sigma \sim 0.3 \) in accordance with the amplitude obtained by inversion of the tidal signal in time domain. This model thus provides satisfactory results
in spite of the simplicity of the unidimensional assumption. It is used for approximating the permeability.

5.4. Estimation of the aquifer permeability

Equations B1 to B5 directly involve the hydraulic diffusivity $D$ though the skin length $\delta$ of equation 14. From the estimate of the response of the aquifer pressure to oceanic tides, it is possible to derive constraints on the hydraulic diffusivity, by a method similar to that adopted for drawing figure 13. As the diffusivity is unknown, the extension of the system must be precised. Figure 13 indicates that beyond 3,000 m, the aquifer extension does not play a crucial factor onto the value of the phase lag. The response is rather dominated by the boundary closer to the borehole. The north-south extension of the aquifer (here 5000 m) is controlled by the distance that separates the two most probable hydraulic barriers in the north-south direction, namely the Aigio fault and the Helike faults (figure 3).

The concept of distance between the borehole and the Aigio fault is to be clarified. As the fault dip is $\theta = 60^\circ$, we take as the distance from the borehole to the no flow boundary the average of the distance of the vertical borehole to the fault plane along the thickness of the borehole aquifer $H$: $X = x - x_N = \cotan(\theta) \frac{H}{2}$. We took here into account the Gulf width, by considering the full form of equation 17 with a Gulf width $W = 10$ km.

We then obtain the maps of figure 11. The phase shift map shows that this parameter exhibits a high sensitivity to the hydraulic diffusivity of the aquifer. When this diffusivity is large, the fluid transfer is very rapid, so that large phase shift cannot occur. When the diffusivity is small, the length skin $\delta$ is small, so that the effective distance $X$, separating the borehole to the northern aquifer boundary, is so important that the boundary is screened out and does not affect any more the borehole. The study on Earth tides
performed in section 3.2 indicates that the diurnal phase shift is $17.5^\circ \pm 7.5^\circ$ while the semi-diurnal phase shift is comprised between $-7.5^\circ$ and $7.5^\circ$. The two constraints agree on a region where $X \geq 100 \text{ m}$ and $10^{1.2} \text{ m}^2/\text{s} \leq D \leq 10^{1.5} \text{ m}^2/\text{s}$. From the first inequality and the expression of $X$, we deduce the minimum height of the aquifer to be $170 \text{ m}$, consistent with the 230-m length of the karst intersected by the borehole. This is consistent with the presence of internal heat convection in this aquifer [Doan and Cornet, 2005a].

As $S = (7.7 \pm 2) \times 10^{-7} \text{ m}^{-1}$, we constrain the hydraulic conductivity $C = D \times S$ to be between $9 \times 10^{-6} \text{ m}/\text{s}$ and $3 \times 10^{-5} \text{ m}/\text{s}$ (corresponding to a hydraulic permeability ranging between $9 \times 10^{-13} \text{ m}^2$ and $3 \times 10^{-12} \text{ m}^2$). We also computed the amplitude of the response of the aquifer to oceanic tides $dP/B$ for a unit $d\sigma = \rho_f g h$. It increases when the northern boundary is closer to the seashore, and induces larger $X$ value. It also diminishes as the hydraulic diffusivity is stronger. Intuitively, in such a case, the pressure field is equilibrated more easily, and the contribution of southern parts of the aquifer, where the loading is smaller, is more important, to the detriment to the high load near the borehole and the northern boundary. The region compatible with the phase shift corresponds to a response $\frac{\partial P}{\partial \sigma} = 0.27 \pm 0.02$. To recover the experimental value $\frac{\partial P}{\partial \sigma} = 0.2 \pm 0.1$, this suggests a minimum for $B = 0.25/0.3 \sim 0.8$. This value is larger than values retrieved from seismic data. It is however more reliable, as the method implies a larger scale than the seismic celerities determined through borehole logging on a metric scale. From equation 5, we derive also the undrained Poisson coefficient $\nu_u = \frac{3\gamma - B}{3\gamma + B} \sim 0.2$, a value more reasonable than that extracted directly from the borehole logging. From equation 5, we can also recover an estimate of the undrained bulk modulus $K_u \sim 20 \text{ GPa}$, which is also a reasonable order of magnitude.
6. Discussion

6.1. Pertinence of the aquifer model

In this paper, we applied Darcy equation and poroelastic theory of homogeneous medium to derive the permeability. The Darcy law underlies equation 13. But, in a medium with large pores, as the karstic aquifer, it may be more appropriate to use the Forchheimer equation [Barenblatt et al., 1990]:

\[-\nabla p = \frac{\eta}{K} \vec{v} + b \rho_f v \vec{v}\]

which introduces a non-linear term in addition to the Darcy relation that involves the permeability $K$ of the medium and the dynamic viscosity $\eta$ of the fluid. Ward [1964] related the $b$ factor to the permeability $K$ through the relationship $b = C/\sqrt{K}$, where $C \sim 0.55$. Taking a permeability of the karst $K = 1.5 \times 10^{-12} \text{m}^2$, we get $b = 5 \times 10^5 \text{m}^{-1}$. The non-linear term becomes prominent only if we respect the condition on fluid velocity $v \gg \frac{\eta}{K b \rho_f} = 1.8 \text{m/s}$. This threshold is very large, so that Darcy law is valid for the present analysis.

The assumption of a unidimensional flow is also questionable. The ratio of the horizontal pressure gradient to the vertical pressure gradient predicted from equation 17 is only 1.5, for $x=500 \text{m}$, $z_0=700 \text{m}$ and $W=10 \text{km}$. The effect of a vertical diffusion may be quantified by examining results provided by the Taylor expansion or by looking at the exact response to idealized loading function as in figure 13. With a distance between the two boundaries of the aquifer below 700 m, the phase shift induced by the vertical diffusion is negligible.

The choice of the distance $X$ of the borehole to the aquifer boundary is also arbitrary, as Aigio fault plane is tilted relatively to the vertical and to the borehole axis. However,
the vertical pattern of figure 11 limits the sensitivity of our results to the arbitrary choice
of the distance $X$.

A last strong hypothesis is the assumption of a uniform aquifer. We are unfortunately
unable to tackle a more realistic description, due to our lack of pertinent data. One can
only notice that cavities of metric size were visible all along the 230 m-long imaging log of
the AIG10 borehole. The homogeneity then holds vertically for the upper three hundred
meters of the karst, and is assumed to be valid for its lower part.

6.2. Consequence for the permeability of the Aigio fault

The original difference in overpressure between the two aquifers on both sides of the
Aigio fault indicates that the fault is impervious. Other arguments confirm this fact. The
two aquifers display a large difference in crystallization process. Whereas the hanging
wall is rich in neo-formed calcite crystals, the footwall exhibits dissolution features. Its
fractures are empty or filled by automorphous calcite \cite{Frima et al., 2005}. This is consis-
tent with Giurgia et al.’s finding \cite{Giurgea et al., 2004} that there is a sharp difference in
chemical composition for the fluids collected on both sides of the fault.

Also laboratory tests conducted on samples of the smeared clay layer yield a permeabil-
ity value equal to $0.8 - 2 \cdot 10^{-18}$ m$^2$ \cite{Song et al., 2004}. However, it is not clear whether the
fault is impervious only where the clay has been smeared or whether it remains impervi-
ous at greater depth. Indeed, the karstified carbonates intersected in the footwall belong
to the Tripolitza nappe, made of limestone free of thick clay layers. This is demonstrated
by the thermal convection observed inside the karst, which requires that this formation
be homogeneous over a thickness greater than 400 m \cite{Doan and Cornet, 2005a}, a feature
not compatible with the Olonos-Pindos formation \cite{Rettenmaier et al., 2004}. But the
response of the karstic aquifer to oceanic load is consistent with an impervious fault at all depths, down to the bottom of the karst. Although the fault offset is of the order of 150 m. This suggests that the fault constitutes an impervious boundary extending down to the bottom of the karst, so that the fault gouge is impervious even in the absence of smeared radiolarite.

7. Conclusions

Tide induced pressure variations have been recorded in the deep karstic aquifer encountered by the AIG10 well, below the active Aigion fault. Their analysis has provided means to obtain a complete hydro-mechanical characterization of this formation. Located near the sea, the borehole is greatly affected by the oceanic load. The high quality of the pressure signal has been taken to advantage for separating effects of Earth tides, oceanic loading and barometric effects. Amplitudes and phases of the respective response coefficients have been shown to provide all required information for a complete hydro-mechanical characterisation of the system.

The elastic mechanical behavior is described by the two undrained coefficients: the Poisson ratio $\nu_u \sim 0.2$ and the bulk incompressibility $K_u \sim 20$ GPa. The coupling between fluid pressure and rock deformation is expressed through the Skempton coefficient $B \sim 0.8$, while the storativity is $S = (7.7 \pm 2) \cdot 10^{-7}$ m$^{-1}$. Further we have shown that the borehole pressure gauge is equivalent to a very efficient strainmeter: with a coefficient $\frac{dp}{d\epsilon} = BK_u \sim 17$, GPa and a pressure sensor able to resolve pressure steps of 10 Pa, we can identify strain steps as small as $5 \cdot 10^{-10}$.

Full analytical solutions, for both complex realistic loading (after Taylor expansion) or simple Boussinesq strip loading have been developed for modeling the aquifer pressure
response to oceanic load. These solutions show that the rapid drop in amplitude of the pressure response with diffusivity, together with the absence of strong phase shift, delimits a narrow domain of values for the aquifer conductivity \((20\pm10)\cdot 10^{-5} \text{ m/s}\) (corresponding to a permeability \(K = (2\pm1)\cdot 10^{-12} \text{ m}^2\)). Further, the model suggests that the fault remains impervious at greater depth, below the domain where clay smearing has occurred, so that the active Aigio fault constitutes an impervious northern boundary to the karstic aquifer encountered in the footwall of the fault.

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**Appendix A: Analytical solution of the aquifer response to a semi-infinite
ocean**

Equation 13 may be analytically solved for the loading function given by equation 17.
If we suppose that the ocean is infinite ($W = \infty$), it may further be simplified as the
last term in brackets vanishes. Starting from the general solution to any load function

$$\sigma(x) = \sigma_{kk}(x)/3,$$

we integrate by part the integrals. As the derivative of the arc-tangent function $\text{atan}(z)$ is

$$1/(z^2 + 1),$$

we then decompose this polynomial fraction into simple parts in the complex
space. The solution thus involves the exponential integral function $E_i(z) = - \int_{-\infty}^{\infty} \frac{e^{-t}}{t} dt = \frac{(1 - i)}{2 \delta} \left( C_1 + \int_{-\infty}^{x} e^{(1-i) \frac{u}{\delta}} \sigma(u) du \right)$

$$P(x) = \frac{(1 - i) B e^{(1-i) \frac{x}{\delta}}}{2 \delta} \left( C_1 + \int_{-\infty}^{x} e^{(1-i) \frac{u}{\delta}} \sigma(u) du \right) - \frac{(1 - i) B e^{(1-i) \frac{x}{\delta}}}{2 \delta} \left( C_2 + \int_{-\infty}^{x} e^{-(1-i) \frac{u}{\delta}} \sigma(u) du \right)$$
\[ f^x_\infty \frac{e^t}{t} dt \] \cite{Abramovitz and Stegun, 1964}, already tabulated in standard scientific software as Matlab or Mathematica. This function is multivalued and a discontinuity semi-interval has to be arbitrarily set. The standard choice is to put the branch cut on the negative real axis, but another branch cut can be chosen if necessary. The general solution of equation 13 is thus given by:

\[
P(x) = \frac{B}{\pi} \text{atan} \left( \frac{z_0}{x} \right) \\
\quad - \frac{i}{4} B e^{\frac{(1+i)(x-i\delta z_0)}{\delta}} E_i \left( \frac{-(1+i)(x-i\delta z_0)}{\delta} \right) \\
\quad - \frac{i}{4} B e^{\frac{-1-(1+i)(x+i\delta z_0)}{\delta}} E_i \left( \frac{(1+i)(x+i\delta z_0)}{\delta} \right) \\
\quad + \frac{i}{4} B e^{\frac{(1+i)(x+i\delta z_0)}{\delta}} E_i \left( \frac{-1-(1+i)(x+i\delta z_0)}{\delta} \right) \\
\quad + \frac{i}{4} B e^{\frac{-1-(1+i)(x-i\delta z_0)}{\delta}} E_i \left( \frac{(1+i)(x-i\delta z_0)}{\delta} \right) \\
\quad + c_1 e^{(1+i)\frac{x}{\delta}} + c_2 e^{-(1+i)\frac{x}{\delta}}
\]

The two constants \( c_1 \) and \( c_2 \) depend on the no-flow Conditions \( \frac{\partial P}{\partial x} = 0 \) on the northern \((x = x_N)\) and southern \((x = x_S)\) boundaries. This gives a 2-equation system, whose resolution provides:

\[
c_1 = \frac{i}{4 \pi \left( e^{2(1+i)\frac{z_0}{\delta}} - e^{2(1+i)\frac{\delta}{z_0}} \right) \times} \\
\left( e^{2(1+i)\frac{z_0}{\delta}} + (1-i) \frac{\delta}{z_0} \right) E_1 \left( -\frac{1-i}{\delta} x_N - i z_0 \right) \\
\quad - e^{-(1-i)\frac{\delta}{z_0}} E_1 \left( (1+i) \frac{\delta}{z_0} \right) \\
\quad - e^{2(1+i)\frac{z_0}{\delta}} + (1-i) \frac{\delta}{z_0} \right) E_1 \left( -(1+i) \frac{\delta}{z_0} x_S - i z_0 \right) \\
\quad + e^{-(1-i)\frac{\delta}{z_0}} E_1 \left( (1+i) \frac{\delta}{z_0} \right)
\]

\[
\quad - e^{2(1+i)\frac{z_0}{\delta}} - (1-i) \frac{\delta}{z_0} \right) E_1 \left( -(1+i) \frac{\delta}{z_0} x_N + i z_0 \right) \\
\quad + e^{(1-i)\frac{\delta}{z_0}} E_1 \left( (1+i) \frac{\delta}{z_0} x_N + i z_0 \right)
\]
\[ + e^{2(1+i)\frac{2\pi}{T} - (1-i)\frac{2\pi}{T}} E_1\left(-\left(1 + i\right) \frac{x_S + i z_0}{\delta}\right) \]
\[ - e^{(1-i)\frac{2\pi}{T}} E_1\left((1 + i) \frac{x_S + i z_0}{\delta}\right) \]  

and

\[ c_2 = \frac{i}{4 \pi \left( e^{2(1+i)\frac{2\pi}{T}} - e^{2(1+i)\frac{2\pi}{T}} \right)} \times \]
\[ \left(e^{(1+i)(2x_N+2x_S-i z_0)/\delta} E_1\left(-(1 + i) \frac{x_N - i z_0}{\delta}\right) \right) \]
\[ - e^{2(1+i)x_S-(1-i)z_0} E_1\left((1 + i) \frac{x_N - i z_0}{\delta}\right) \]
\[ - e^{(1+i) \frac{2x_N+2x_S-i z_0}{\delta}} E_1\left(-(1 + i) \frac{x_N - i z_0}{\delta}\right) \]
\[ + e^{2(1+i)x_S-(1-i)z_0} E_1\left((1 + i) \frac{x_N + i z_0}{\delta}\right) \]
\[ - e^{(1+i) \frac{2x_N+2x_S+i z_0}{\delta}} E_1\left(-(1 + i) \frac{x_N + i z_0}{\delta}\right) \]
\[ + e^{2(1+i)x_S+(1-i)z_0} E_1\left((1 + i) \frac{x_N + i z_0}{\delta}\right) \]
\[ - e^{(1+i) \frac{2x_N+2x_S+i z_0}{\delta}} E_1\left(-(1 + i) \frac{x_N + i z_0}{\delta}\right) \]  

\[ (A2) \]

\[ (A3) \]

**Appendix B: Effect of borehole boundaries: Taylor expansion**

From borehole data, the Aigio and the Helike faults are known to be impervious. We study now the effect of such boundaries on the borehole pressure. To quantify the effect of aquifer boundaries, we decompose the loading function in a Taylor expansion. As we deal successively with odd and even functions, we will assume that the aquifer boundaries are at \( x = -L/2 \) and \( x = L/2 \). The imperviousness of the Aigio fault and of the nearby Helike fault [Giurgea et al., 2004] leads us to take a no-flow condition at the aquifer boundaries.
B1. Order 0 : constant term

The solution to this loading function is of course uniform, without any phase lag.

B2. Order 1 : linear term

If we apply the function $\sigma(x) = k x$ in equation 13, its solution has a relatively complicated expression:

$$P(x) = B k x + \frac{1 - i}{2(i + e^{(1+i)\frac{L}{2}})} \left( e^{(1+i)\frac{L}{2+x}} - e^{(1+i)\frac{L}{2-x}} \right) B k \delta$$

which may be decomposed into

1. a general solution $B k x$, which implies no phase lag

2. a diffusion term, which is prominent only near the boundaries of the aquifer via the term $\frac{L/2+\delta}{\delta}$. Indeed, far from the impervious boundaries, the two terms in parentheses of the numerator are similar and cancel each other. This phase delay compromises the pressure gradient of the loading function and the null gradient imposed by the no-flow boundary condition.

Hence this component does not result in any phase delay if the borehole is far from the aquifer boundary, i.e. farther than $\delta \sim 600$ m.

B3. Order 2 : Quadratic term

The response to a loading function $\sigma = c x^2$ is still more complex:

$$P(x) = B c \left( x^2 - i \delta^2 \right) - \frac{B c L \delta}{(1+i)(e^{(1+i)\frac{L}{2}} - 1)} \left( e^{(1+i)\left(\frac{L}{2}-x\right)} + e^{(1+i)\left(\frac{L}{2}+x\right)} \right)$$
Here, even the general solution $Bc(x^2 - i\delta^2)$ is delayed. This is due to the fact that $\Delta \sigma = \frac{\delta^2}{\delta^4} \neq 0$

**B4. Order 3 : Cubic term**

The response to a loading function $\sigma = cx^3$ is still more complex, but exhibits a few structure:

$$P(x) = Bc x \left(x^2 - 3i\delta^2\right) + \frac{3Bc(1 + i)\delta(iL^2 + 4\delta^2)}{8(e^{(1+i)L/2} + 1)} \left(e^{(1+i)(L/2 + x)} - e^{(1+i)(L/2 - x)}\right)$$

(B3)

As for the quadratic term, both the general solution and the boundary term are phase lagged.

**B5. Other terms**

For the more general loading condition $\sigma = cx^n$, a general solution can be derived.

$$P(x) = C_+ e^{(1+i)x} + C_- e^{-(1+i)x}$$

(B4)

$$+ \frac{Bc}{2} \left(\frac{(1 + i)\delta}{2i}\right)^n \Gamma_{n+1} \left(-\frac{(1 + i)x}{\delta}\right) e^{-\frac{(1+i)x}{\delta}}$$

$$+ \frac{Bc}{2} \left(\frac{(1 + i)\delta}{2i}\right)^n \Gamma_{n+1} \left(\frac{(1 + i)x}{\delta}\right) e^{\frac{(1+i)x}{\delta}}$$

where

$$C_\pm = \frac{Bc \left((\mp L)^n + (\pm L)^n e^{(1+i)L/2}\right)}{e^{2(1+i)L/2} - 1} \times$$

$$\left(\frac{1}{2} \left(\frac{(1 + i)\delta}{2iL}\right)^n \Gamma_{n+1} \left(-\frac{(1 + i)L}{\delta}\right)\right)$$

$$- \left(\frac{1}{2} \left(\frac{(1 + i)\delta}{2iL}\right)^n \Gamma_{n+1} \left(\frac{(1 + i)L}{\delta}\right) e^{-\frac{(1+i)L}{\delta}}\right)$$

(B5)

where $\Gamma_{n+1}(z) = \int_z^\infty t^n e^{-t} dt = n! e^{-z} \sum_{k=0}^n \frac{z^k}{k!}$. This solution also exhibits some phase lag.
B6. Conclusions on the Taylor expansion selection

The Taylor expansion yields some intuitive insight about the influence of boundaries on the expected phase lag. A small aquifer necessitates only a small Taylor expansion. If the aquifer is small or if the borehole is far enough from the shoreline, so that only a linear expansion is necessary, then a null offset occurs. This observation validates in the last case the van der Kamp model, which shows similar results. The advantage of the Taylor expansion is that it is a general method, which can be applied to more realistic loading functions, as would be the case if one takes into account the bathymetry of the Gulf. But the method becomes rapidly tedious if a large expansion is necessary. This is the case for the Aigio borehole, if we assume that the aquifer exhibits an expansion equal to the distance between the Aigio fault and the Helike fault (5 km): an expansion of order 16 is necessary to fit the Boussinesq loading function of figure 10. It is then simpler to take the full simple Boussinesq function, given the simple expression of equation 17.
Figure 1. Schematic structural cross-section through Aigio fault. From 400 m to 700 m, clays and radiolarite isolate the bottom aquifers from surface aquifers. The Aigio fault is impervious and separates two artesian aquifers. The figure is extracted from [Cornet et al., 2004b]
Figure 2. Tidal variations recorded on the pressure sensors. They are sampled at 1/8 Hz, with a resolution better than 1% (Upper graph). The spectrum of these data over the first semester of 2004 shows both diurnal and semi-diurnal variations, corresponding to the major tidal components $O_1$, $K_1$, $M_2$ and $S_2$.

Figure 3. Location of the sensors used for interpreting the tidal variations recorded on the pressure sensor of Aigio well. The large filled circles indicate tide gage and the squares the barometric pressure gages.
Figure 4. Data used to interpret the tidal variations on the Aigio pressure sensor. The signals are first detrended thanks to a four-pole Butterworth high-pass filtering with a corner period of 4 days. We used forward and reverse filtering to minimize phase shifting induced by this operation.
<table>
<thead>
<tr>
<th>Tidal Frequency</th>
<th>Aigio pressure ($\tilde{h}_k$(press))</th>
<th>Trizonia tide gauge ($\tilde{h}_k$(oc))</th>
<th>Aigio Earth tide packet</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(kPa/nstr)</td>
<td>(°)</td>
<td>(cm/nstr)</td>
</tr>
<tr>
<td>Q1</td>
<td>0.706 ± 0.205</td>
<td>0.022 ± 0.001</td>
<td>-175 ± 3</td>
</tr>
<tr>
<td>O1</td>
<td>0.930 ± 0.018</td>
<td>0.0190 ± 0.0002</td>
<td>-176.0 ± 0.7</td>
</tr>
<tr>
<td>M1</td>
<td>0.965 ± 0.017</td>
<td>0.014 ± 0.004</td>
<td>178 ± 17</td>
</tr>
<tr>
<td>P1</td>
<td>0.990 ± 0.008</td>
<td>0.0151 ± 0.0007</td>
<td>-174 ± 2.5</td>
</tr>
<tr>
<td>K1</td>
<td>1.011 ± 0.012</td>
<td>0.0130 ± 0.0002</td>
<td>-164.8 ± 0.8</td>
</tr>
<tr>
<td>J1</td>
<td>1.041 ± 0.017</td>
<td>0.020 ± 0.003</td>
<td>-180 ± 8</td>
</tr>
<tr>
<td>O01</td>
<td>1.264 ± 0.206</td>
<td>0.018 ± 0.004</td>
<td>-166 ± 12</td>
</tr>
<tr>
<td>2N2</td>
<td>1.675 ± 0.205</td>
<td>0.054 ± 0.010</td>
<td>-179 ± 10</td>
</tr>
<tr>
<td>N2</td>
<td>1.897 ± 0.017</td>
<td>0.046 ± 0.002</td>
<td>-157 ± 2.5</td>
</tr>
<tr>
<td>M2</td>
<td>1.932 ± 0.018</td>
<td>0.0513 ± 0.0004</td>
<td>-162.0 ± 0.5</td>
</tr>
<tr>
<td>L2</td>
<td>1.967 ± 0.017</td>
<td>0.08 ± 0.01</td>
<td>-150 ± 7</td>
</tr>
<tr>
<td>S2</td>
<td>1.994 ± 0.009</td>
<td>0.060 ± 0.001</td>
<td>-172 ± 1</td>
</tr>
<tr>
<td>K2</td>
<td>2.227 ± 0.225</td>
<td>0.065 ± 0.003</td>
<td>-166 ± 2</td>
</tr>
</tbody>
</table>

**Table 1.** Tidal coefficients relative to the predicted tidal theoretical tidal volume deformation in Aigio. Barometric pressure effects were removed with the remove-restore module of ETERNA3.3 [Wenzel, 1996a]. The volume deformation is also inverted with ANALYZE to check the quality of the inversion. This is to be compared with the expected results, indicated in brackets.
Figure 5. The signals are first detrended thanks to a four-pole Butterworth high-pass filtering with a corner period of 4 days. We used forward and reverse filtering to minimize phase shifting induced by this operation. The process highlights the tidal variations, as shown in figure 4. A linear combination of the data of figure 4 fits fairly well the borehole pressure data.
Figure 6. Quality of the linear regression of the tidal variations of the borehole pressure from the theoretical Earth tides and the diurnal and semi-diurnal oceanic tides, reconstructed from the tidal coefficients of table 1. The last two tidal variations are time shifted before each regression and the determination coefficient $R^2$ is computed each time. The regression is performed for data comprised between November 20th and December 10th 2003. A close-up view around the maximal $R^2$ coefficient is presented on the lower graph.
Figure 7. Graph similar to figure 6, except that the data has been acquired 6 months later, between May 29th and June 17th 2004.
Figure 8. Comparison of the variations recorded by the Temeni atmospheric pressure gauge (dashed line) and the Aigio borehole pressure sensor (solid line) during December 2003 and January 2004.

Figure 9. Conventions used in equation 17 to describe Boussinesq strip surface loading.
Figure 10. Loading profile derived from the Boussinesq equation 17, at a depth of 700 m for two different values of the undrained Poisson ratio $\nu_u$. The borehole is located at 500 m from the seashore, and in a transition zone separating the loaded and free regions of the van der Kamp model.
Figure 11. Response of the borehole pressure in a borehole regarding to the oceanic load, computed in a unidimensional model, for a depth of $z_0 = 1000 \text{ m}$ in an aquifer of north-south extension $L = 5000 \text{ m}$, with a distance to seashore $x = 500 \text{ m}$.
Various studies show that the Aigio fault is impervious at its intersection with the borehole. The smearing of a 10 m thick radiolarite layer has been proposed to explain its impermeability. However, given the 150 m of vertical offset of the fault, this phenomenon affects only the upper part of the aquifer.
Figure 13. Map of phase lag relatively to the oceanic load for diurnal (upper graph) and semi-diurnal waves (bottom graph). We assumed a hydraulic diffusivity $D = 20 \text{ m}^2/\text{s}$ and a Skempton coefficient $B = 1$. 