# Evaluation of hydrostatic pressure variation in a curved layer seismic reservoir.

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## Abstract

The present work aims to evaluate the hydrostatic pressure variation in a seismic reservoir characterized by a model of curved layers that presents regions of synclinals that, by stressing the upper layers, cause deformation in them over millions of years. These synclinal regions are natural traps that have the ability to confine fluids, provided they have porosity, and in this process, the fluids are subjected to hydrostatic pressures. The determination of the hydrostatic pressures to which these fluids are submitted is possible because it is related to the variation in velocities of the seismic waves P and S that propagate inside the reservoirs, which in turn are related to the elastic and petrophysical parameters of the rocks, such as the shear modulus and the incompressibility modulus. From the determination of these parameters, one can determine variations in the hydrostatic pressure of fluids and plot the results on color maps. The importance of this study lies in the fact that these pressures cause folding in these layers, explaining the presence of curved layers in complex media and also the presence of hydrocarbons in these reservoirs. The initial step in the processing flow of this work was to create synthetic models that depict the geology of the curved layers and assign constant density values and P and S wave velocities for each layer. The software used to create these geological models is the free seismic data processing software called Seismic Un*x, developed by Central Wave Processing (CWP) at the Colorado School of Mines

## Introduction

Seismic method is the geophysical method that uses the propagation of seismic waves in subsurface to obtain data and images of its complex geological features (ROKSANDIC (1978); AKI & RICHARDS (1980); SHERIFF & GELDART (1982)). Hydrocarbon exploration is the most important application of seismic, and in this regard, many studies have been published to construct theoretical rock models that can physically represent complex real models (GASSMANN (1951); BIOT (1956a); BIOT (1956b)). In these studies, the physical and petrophysical properties of rocks have been the subject of study to make it possible to infer values of elastic rock properties when saturated by fluids. The presence of folds in the buckling sedimentary layers, in response to compressive forces, creates a natural trap for hydrocarbon confinement, with the possibility of anticlinal and sinclinal formations occurring, depending on the rock properties and the magnitude and direction of the forces applied. The most important example is a dome, which is an anticlinal structure featuring a circular or oval overhang of rock layers. The layers on the flanks of a dome surround it at a central point and dip radially from this point, and certain domes can be attributed to bodies of less dense material pushing the overlying sediments upward. As the seismic wave beam strikes the interface separating the fluid contained within these structures, it deflects and indicates the change in velocity and density of the medium. The observation of the decrease in P-wave velocity indicates that the new medium in which the seismic wave is propagating is less dense than the previous medium, and may even be fluid.

The relationships between applied forces and deformations are expressed in terms of the concepts of stress and strain (TELFORD et al. (1990)).

### Stress

When a force is applied to an area element, stress is defined as force per unit area to which it is applied. If the force varies from point to point, the stress also varies and its value is given by:

$$\sigma_{ij} = \lim_{A_j \to 0} \frac{F_i}{A_j} \qquad (i, j = 1, 2, 3).$$

The stress is normal or pressure when the force is perpendicular to the area; and shear or shear when the force is tangential to the area element. Normal stresses are indicated by $\sigma_{ij}$ and shear stresses by $\tau_{ij}$.

When two subscripts are equal (like $\sigma_{xx}$), the stress is said to be normal stress; and when the subscripts are different (like $\tau_{xy}$) the stress is said to be shear or shear.

The nine stress components completely define the state of stress to which the body is subjected and can be conveniently described by the stress matrix:

$$\sigma_{ij} = \begin{bmatrix} \sigma_{xx} & \tau_{xy} & \tau_{xz} \\ \tau_{yx} & \sigma_{yy} & \tau_{yz} \\ \tau_{zx} & \tau_{zy} & \sigma_{zz} \end{bmatrix}$$

If the forces acting on the body are compensated so that they do not cause rotations, this matrix is symmetric (i.e. $\sigma_{xy} = \sigma_{yx},\ \sigma_{xz} = \sigma_{zx},\ \sigma_{yz} = \sigma_{zy},$ ) and contains only independent elements.

This occurs when the medium is in static equilibrium. In this case, the forces must be balanced. This medium has three stresses $\sigma_{xx},\ \tau_{yx}$ and $\tau_{zx}$ acting that must be equal and opposite to the corresponding stresses acting in the opposite direction, with similar relationships to the remaining four faces. In addition, a pair of shear stresses, such as $\tau_{yx}$, constitute a pair tending to rotate the element about $z$. The magnitude of the pair is the moment, defined by:

$$F \times b = (\tau_{yx} dydz)dx$$

where $F$ is the magnitude of the force and $b$ is the arm.

If we consider the stresses on the other four faces, we conclude that this pair is opposed only by the pair $\tau_{xy}$ with magnitude $(\sigma_{xy} dxdz)dy$. Since this element is in equilibrium, the total momentum must be zero; so $\tau_{xy} = \tau_{yx}$. In general, in the case of static equilibrium:

$$\sigma_{ij} = \sigma_{ji}$$

**Strain**

When an elastic body is subjected to stresses, variations occur in its dimensions and shape. These variations, which are called deformations. These deformations can be longitudinal or shear. Deformations are indicated by $\varepsilon_{ij}$.

Longitudinal strains are deformations produced by the components of the normal stresses $\sigma_{xx}$, $\sigma_{yy}$ and $\sigma_{zz}$. The longitudinal strains are $\varepsilon_{xx}$, $\varepsilon_{yy}$ and $\varepsilon_{zz}$.

Shear strains are strains produced by the components of the shear stresses $\sigma_{xy}$, $\sigma_{xz}$ and $\sigma_{yz}$. The shear strains are $\varepsilon_{xy}$, $\varepsilon_{xz}$ and $\varepsilon_{yz}$.

The longitudinal and shear strains define a symmetric 3×3 matrix, called the strain matrix.

$$\varepsilon_{ij} = \frac{1}{2}\left(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i}\right) = \begin{bmatrix} \varepsilon_{11} & \varepsilon_{12} & \varepsilon_{13} \\ \varepsilon_{21} & \varepsilon_{22} & \varepsilon_{23} \\ \varepsilon_{31} & \varepsilon_{32} & \varepsilon_{33} \end{bmatrix} \qquad (i, j = 1, 2, 3)$$

**Poisson's ratio**

In elastic behavior the deformations $\varepsilon_{yy}$ and $\varepsilon_{zz}$ are not independent of $\varepsilon_{xx}$.

The elongation in the direction parallel to $x$ is accompanied by a contraction in the directions parallel to the $yy$ and $zz$ axes (the latter is not represented in the figure, as it only represents what is happening in the $x - y$ plane). The deformations $\varepsilon_{yy}$ and $\varepsilon_{zz}$ have an opposite sign, but are proportional to the extension $\varepsilon_{xx}$, being given by:

$$\varepsilon_{yy} = -\nu\varepsilon_{xx} \qquad \varepsilon_{zz} = -\nu\varepsilon_{xx} \tag{1}$$

$$\nu = -\frac{\varepsilon_{yy}}{\varepsilon_{xx}} = -\frac{\varepsilon_{zz}}{\varepsilon_{xx}} \tag{2}$$

The proportionality constant $\nu$ is called Poisson's ratio.

**Rotation tensor**

In addition to stresses and strains, the body is subject to rotations about the three axes $(x, y, z)$, which are given by:

$$\begin{cases} \theta_x = & \frac{1}{2}\left(\frac{\partial w}{\partial y} - \frac{\partial v}{\partial z}\right) \quad \text{Rotation about the x-axis} \\ \theta_y = & \frac{1}{2}\left(\frac{\partial u}{\partial z} - \frac{\partial w}{\partial x}\right) \quad \text{Rotation about the y-axis} \\ \theta_z = & \frac{1}{2}\left(\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}\right) \quad \text{Rotation about the z-axis} \end{cases}$$

The rotation tensor is described in a similar way to the strain tensor by the components of the rotation in the form:

$$\theta_{ij} = \frac{1}{2}\left(\frac{\partial u_j}{\partial x_i} - \frac{\partial u_i}{\partial x_j}\right) = \begin{bmatrix} \theta_{11} & \theta_{12} & \theta_{13} \\ \theta_{21} & \theta_{22} & \theta_{23} \\ \theta_{31} & \theta_{32} & \theta_{33} \end{bmatrix} \qquad (i, j = 1, 2, 3)$$

In the matrix of rotations, $x_1 = x$, $x_2 = y$, $x_3 = z$, $u_1 = u$, $u_2 = v$ and $u_3 = w$. Notice that the tensor of rotations is not symmetric.

**Dilatation**

The variations in dimensions given by normal stresses result in volume variations; the volume variation per unit area is called dilation and is represented by $\Delta$.

$$\Delta = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = \varepsilon_{xx} + \varepsilon_{yy} + \varepsilon_{zz}$$

**Hooke's Law**

Hooke's law states that a given stress is directly proportional to the strain produced. In general, Hooke's law leads to complicated relationships, but when the medium is isotropic, that is, when its properties do not depend on direction, it can be expressed in the following form (TELFORD et al. (1990))

$$\sigma_{ij} = \lambda'\Delta\delta_{ij} + 2\mu\varepsilon_{ij} \qquad (i = 1,\ 2,\ 3) \tag{3}$$

where $\delta_{ij}$ is called the Kroenecker symbol, defined by:

$$\delta_{ij} = \begin{cases} 1 & \text{se } i = j \\ 0 & \text{se } i \neq j \end{cases}$$

The quantities $\lambda'$ and $\mu$ are known as Lamé constants.

If $i \neq j$ the Eq.(3) reduces to $\sigma_{ij} = 2\mu\varepsilon_{ij}$ and we write $\varepsilon_{ij} = \sigma_{ij}/2\mu$, it is evident that $\varepsilon_{ij}$ is smaller the larger $\mu$ is. So $\mu$ is a measure of resistance to shear deformation and is called the stiffness or shear modulus (since $i \neq j$).

**Elastic Constants**

No intervalo de deformação elástica a lei de Hooke nos diz que existe uma relação linear entre a tensão e a deformação, sendo que o quociente entre estas duas grandezas define uma constante elástica. Como por sua vez as deformações já são dadas por quocientes entre comprimentos (por isso são adimensionais) as constantes elásticas têm as mesmas dimensões que a tensão (N/m$^2$). As constantes elásticas, definidas para diferentes tipos de deformações são, o módulo de Young, o coeficiente de rigidez e o módulo de volume ("bulk modulus").

**Young modulus**

Young's modulus (E) is defined from the longitudinal strain. Each longitudinal strain is proportional to the existing normal stress component, i.e:

$$\sigma_{xx} = E\varepsilon_{xx}, \quad \sigma_{yy} = E\varepsilon_{yy}, \quad \sigma_{zz} = E\varepsilon_{zz}$$

$$E = \frac{\sigma_{xx}}{\varepsilon_{xx}} = \frac{\sigma_{yy}}{\varepsilon_{yy}} = \frac{\sigma_{zz}}{\varepsilon_{zz}}$$

**Stiffness Modulus**

The stiffness modulus $\mu$ is defined from the shear or shear deformation. Each shear deformation is proportional to the existing shear stress component, i.e:

$$\sigma_{xy} = \mu\varepsilon_{xy}, \quad \sigma_{xz} = \mu\varepsilon_{xz}, \quad \sigma_{yz} = \mu\varepsilon_{yz}$$

$$\mu = \frac{\sigma_{xy}}{\varepsilon_{xy}} = \frac{\sigma_{xz}}{\varepsilon_{xz}} = \frac{\sigma_{yz}}{\varepsilon_{yz}}$$

From the previous expression we have:

$$\sigma_{ij} = \begin{cases} \sigma_{xy} = \mu\varepsilon_{xy} \\ \sigma_{yz} = \mu\varepsilon_{yz} \\ \sigma_{zx} = \mu\varepsilon_{zx} \end{cases}$$

**Pressure Field**

The construction of a pressure field is important for analyzing low pressure zones, which are favorable for fluid accumulation. The vertical and horizontal pressures are key parameters, also called invariant scalars of the stress tensor. Rewriting the stress-strain relationship for the isotropic medium from Eq.(3) one has:

$$\sigma_{ij} = \lambda'\Delta\delta_{ij} + 2\mu\varepsilon_{ij} \tag{4}$$

$$\Delta = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = \varepsilon_{xx} + \varepsilon_{yy} + \varepsilon_{zz}$$

Considering Eq.(4) we calculate the normal stresses in the form of

$$\sigma_{xx} = \lambda'\Delta\delta_{xx} + 2\mu\varepsilon_{xx}$$

$$\sigma_{yy} = \lambda'\Delta\delta_{yy} + 2\mu\varepsilon_{yy}$$

$$\sigma_{zz} = \lambda'\Delta\delta_{zz} + 2\mu\varepsilon_{zz} = (\lambda + 2\mu)\varepsilon_{zz}$$

where $\delta_{xx} = \delta_{yy} = \delta_{zz} = 1$, and the preceding expressions take the forms:

$$\sigma_{xx} = \lambda'\Delta + 2\mu\varepsilon_{xx}$$

$$\sigma_{yy} = \lambda'\Delta + 2\mu\varepsilon_{yy}$$

$$\sigma_{zz} = \lambda'\Delta + 2\mu\varepsilon_{zz} = (\lambda + 2\mu)\varepsilon_{zz} \tag{5}$$

$$\varepsilon_{zz} = \frac{1}{\lambda + 2\mu}\sigma_{zz}$$

The stress is non-hydrostatic even in horizontal layer media subject only to compaction due to vertical pressure without horizontal displacement, in such cases the vertical stress is defined as equal to the load in the following form

$$\frac{d\sigma_{zz}}{dz} = \rho(z)g(z)$$

$$d\sigma_{zz} = \rho(z)g(z)dz$$

$$\int d\sigma_{zz}(z) = \int_{z=0}^{z} \rho(z)g(z)dz$$

$$\sigma_{zz}(z) = \int_{z=0}^{z} \rho(z)g(z)dz$$

For $z = 0$, you get the vertical pressure at the surface of the layer, given by:

$$\sigma_{zz}(0) = \sigma_{zz} = \rho g z = P_z = P_0 \tag{6}$$

The corresponding horizontal pressure is described in the form:

$$\sigma_{xx} = P_x = P_0(1 - 2\gamma^2) \tag{7}$$

In Eq.(7), the constant $\gamma$ is given by the ratio of the S- and P-wave velocities Landau (1988).

$$\gamma = \frac{\beta}{\alpha} \text{ e } \sigma_{xx} = \sigma_{yy}. \tag{8}$$

The overburden pressure is one third of the sum of the vertical and horizontal stresses, considering the stress relationships in Eq.(5), if the vertical stresses are equal, the loading or hydrostatic pressure is simply the pressure of the overlying layers Sibiryakov (2004):

$$P = \frac{\sigma_{xx} + \sigma_{yy} + \sigma_{zz}}{3}$$

$$P = \frac{1}{3}\left(\frac{3\lambda + 2\mu}{1}\right)\varepsilon_z z$$

$$P = P_0\left(1 - \frac{4}{3}\gamma^2\right) \tag{9}$$

**Data and Methods**

The model used in this work was generated using the free software Seismic Un*x (SU) Cohen & Stockwell (2005) with 10 layers and that presents anticlinal and synclinal geological structure, presenting fault regions, as illustrated in Figure 1.

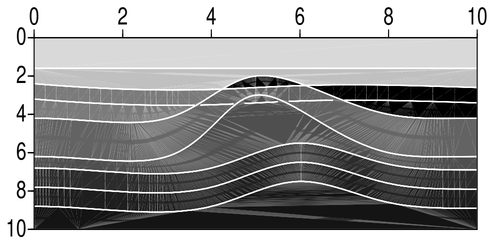

*Figure 1 – Curved Layer Model.*

Because the model is marine, it is natural that the P-wave velocity of the water would be in a range between 1400 to 1500 $m/s$. As it is referred in the literature, the occurrence of basin embayment around 6 $km$ depth. Thus factors such as the deposition of sedimentary layers, natural vertical compression or the weight of each layer itself, the influence of tectonics that is often manifested by a compression or distension of the stress field and that causes a deformation that can occasionally give rise to the creation of other structures.

The above model was made using SU's trimodel function that uses a triangularization using the vagarosity (sloth), and this vagarosity is calculated by $s = \frac{1}{v^2}$. The physical parameters of this model are shown in Table 1.

| Layers | Vagarosities | $V_p(m/s)$ |
|---|---|---|
| 1 | 0.44 | 1507 |
| 2 | 0.39 | 1601 |
| 3 | 0.36 | 1666 |
| 4 | 0.25 | 2000 |
| 5 | 0.20 | 2236 |
| 6 | 0.16 | 2500 |
| 7 | 0.09 | 3333 |
| 8 | 0.06 | 4082 |
| 9 | 0.04 | 5000 |
| 10 | 0.03 | 5773 |

Tabel 1 – P-wave velocity gradient in $m/s$ for each layer.

In this way it is possible to determine several physical parameters that help in locating a possible petroleum system (generator: migration route (when it exists) reservoir; sealant), and that help later through the calculation of each of these parameters for the respective mesh element. The parameters that were used in this work were:

- Shear modulus $(\mu)$;
- First Lamé Parameter $(\lambda)$;
- Relationship between velocities S e P $(\gamma)$;
- RMS S-wave velocity $(Vs_{rms})$;
- RMS P-wave velocity $(Vp_{rms})$;
- Vertical pressure $P_x$;
- Horizontal pressure $P_z$;
- Hydrostatic pressure $P$;

**Results**

Figure 2 shows the interval velocity model of the $S$ waves, varying with velocity from 0 $m/s$ for the upper layer to a velocity of 3800 $m/s$ for the lower layer. In this figure, it can be seen that the S-wave velocity is zero at layer 1, as it does not propagate in the water. S-wave velocities are not as widely used to identify petroleum systems as P-wave velocity.

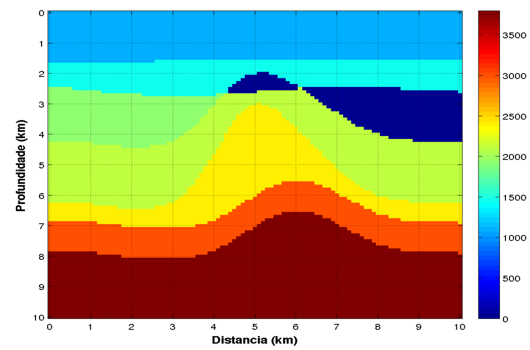


*Figure 2 – S-wave velocity distribution.*

On the other hand, Figure 3 shows the interval velocity model of the $P$ waves, it is observed that the velocities of the layers vary from the top layer with a velocity of 1500 m/s, to the bottom layer with a velocity of 5500 m/s. The velocity of this type of wave tends to increase with depth. Typically the velocity of the P-wave in a reservoir or in a generating layer (provided it has the hydrocarbon fluid) is around 2200 m/s.

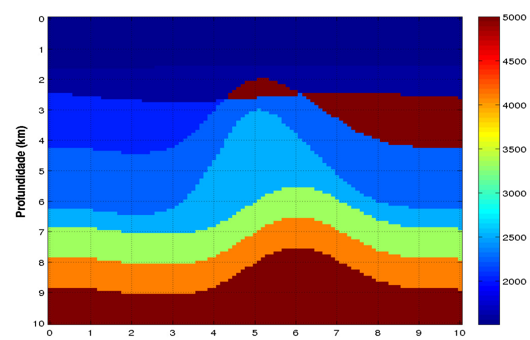


*Figure 3 – P-wave velocity distribution.*

In general, all layers tend to show a large lateral continuity, but an anticlinal structure is observed at coordinate (5,2) in Figure 1 that occurs due to vertical upward pressure from its lower layer. None of the layers is completely horizontal, and this is due to the existence of high strain rates, probably induced by the occurrence of horizontal compressional forces arising from active tectonics, especially around the 3 km extent, as can be confirmed through the graphs of the pressure, and its horizontal and vertical derivatives.

Density is the ratio between the mass and the volume of the soil sample. In this work, density was defined as an input parameter through a vector, containing a set of values compatible with the geological-geophysical model presented in Figure 4. It is also worth noting that the density increases with depth.

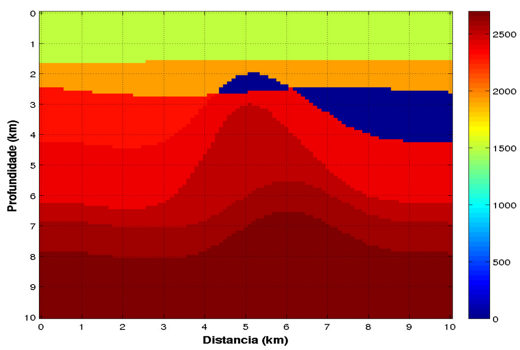


*Figure 4 – Density distribution.*

The shear modulus is calculated using the equation (10):

$$\mu = v_s^2 \rho \tag{10}$$

It is noted that to perform the $5$ calculation, it is necessary to know the set of values of the S-wave velocity and the $5$ density of each layer. This value tends to increase equally with the depth of investigation (See Figure 5).

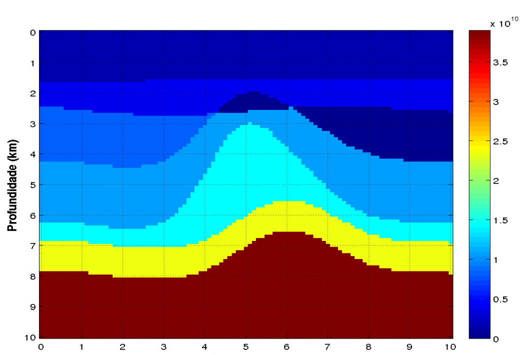


*Figure 5 – Range of shear modulus values.*

To calculate the Lambda parameter, you need to know the values of three parameters for each layer: P-wave velocity, $\rho$ and $\mu$. Therefore, $\lambda$ can be obtained by the form (11):

$$\lambda = v_p^2 \rho - 2\mu \tag{11}$$

This parameter can take either positive or negative values, as shown in Figure 6.

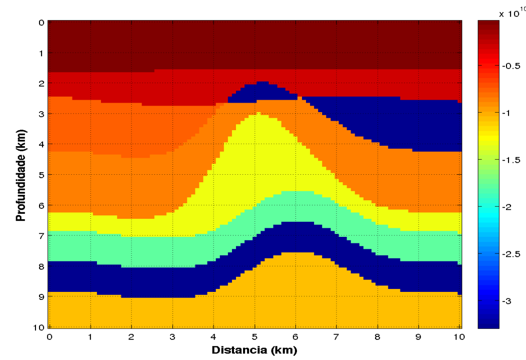


*Figure 6 – Range of values of $\lambda$.*

The $\gamma$ parameter is obtained by the ratio between the S-wave velocity and the P-wave velocity (Eq.7), and is therefore dimensionless - Figure 7. As the value of the S-wave velocity parameter is less than the P-wave velocity, the $\gamma$ value is respectively less than 1.

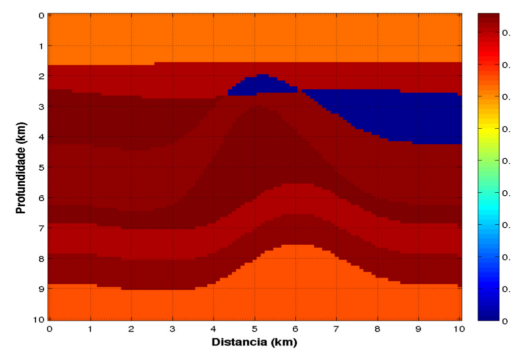


*Figure 7 – $\gamma$ value range.*

For reflection seismic acquisitions where the sedimentary package geometry is practically horizontal between layers and very extensive, it is better to separate and isolate the S-wave arrival time in each of the receivers, due to the reduction of the overlap that appears at the end of the acquisition.

Since the value of the S-wave velocity parameter is practically absent in liquids, it is usually easier to define the interface between the solid material and the liquid, it is easier to define the interface between layer 1 ($V_s = 0$ m/s) and layer 2 ($V_s = 2200$ m/s). Figure 8 shows the result of the RMS velocity calculation for S-waves.

The P-wave RMS velocity parameter (see Figure 9) is one of the main parameters, and is used to perform migration in seismic sections. To calculate this parameter, it was necessary to provide as input variable the S-wave velocity for each layer.

Comparing the values of the RMS velocity parameters of the P-wave with the S-wave, for the same layer the former

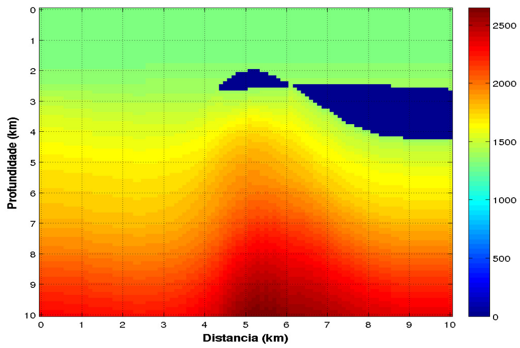


*Figure 8 – Mean squared velocity of the $S$ wave.*

value is always higher than the latter value.

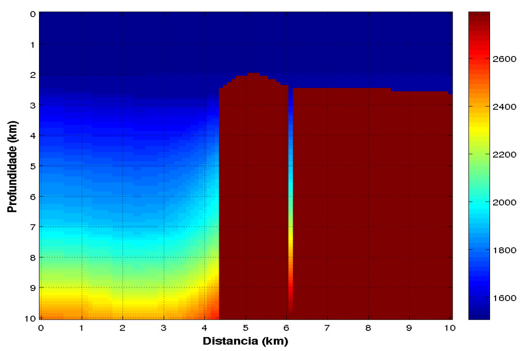


*Figure 9 – Mean squared velocity of the wave $P$.*

The horizontal pressure parameter is calculated using the equation (12):

$$P_x = \gamma^2 P_z, \tag{12}$$

where $P_z$ is the vertical pressure for each layer. The value of $P_x$ is always greater than or equal to 0, being 0 for example in the case when the medium is a fluid. The graph in Figure 10 shows the behavior of the horizontal pressure for each layer of the synthetic model composed of 10 layers presented at the beginning of this section.

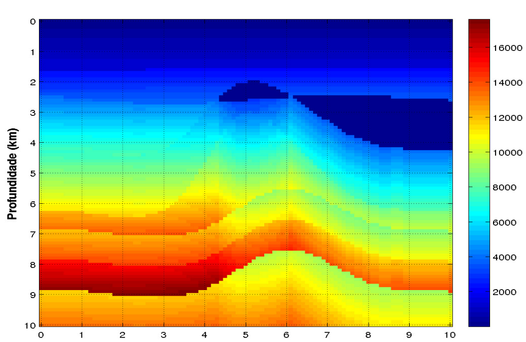


*Figure 10 – Horizontal pressure.*

The calculation of the vertical pressure parameter will be calculated in two steps. The first step will be done for layer 1 as follows (13):

$$P_z = \rho d_z, \tag{13}$$

where $\rho$ is the density and $d_z$ is an increment factor on the z-axis (layer thickness). And the second step will be performed by a pressure accumulation from layer 2, i.e., the vertical pressure of layer 2 will depend on the vertical pressure of layer 1 and so on, so we get the following equation (14):

$$P_z = (P_{z-1}) + \rho d_z, \tag{14}$$

The graph in Figure 11 shows the vertical pressure behavior for each layer of the model in Figure 1.

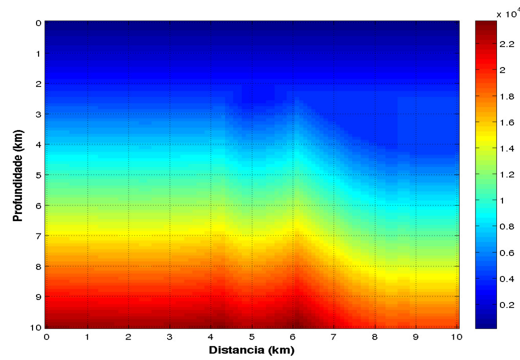


*Figure 11 – Vertical pressure.*

Hydrostatic pressure is a very important parameter in predicting low pressure zones, usually indicative to constitute fluid generating and/or storing lithologies. Fluid generating and/or storing lithologies. The hydrostatic pressure parameter is calculated according to the equation 12 in the previous section.

The information contained in the graph in Figure 12 shows a small variation in the behavior of hydrostatic pressure within each of the layers of the Figure 1. Thus it is not surprising that in the interior of the layers, the values of the hydrostatic pressure parameter increase with depth, so this explains the color gradient within each of the defined layers.

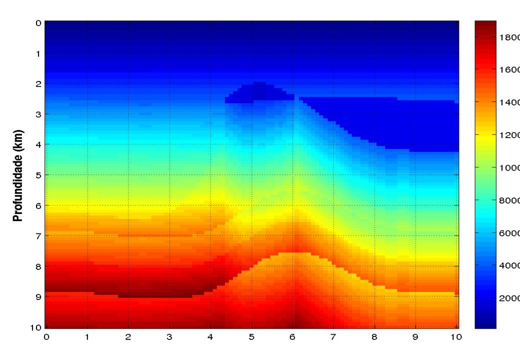


*Figure 12 – Hydrostatic Pressure.*

## Discussion and Conclusions

The present work aimed to calculate the elastic parameters of geological media subjected to the action of tectonic forces from the concepts of the Theory of Elasticity. It is observed that in Figure 3 the values of the p-wave velocities increase with depth, with the exception of the region at the top of the anticlinal structure, which has velocities above 4000 m/s. We observe that in Figure 5 the $\mu$ values are continuous, except in the region of the top of the anticlinal structure, situated at 2 km depth. This anomaly is explained due to the force applied by the upper layer on the walls of the top of the anticlinal. As the values of $\lambda$ depend on the values of $\mu$, it is also observed in Figure 6 an anomaly in the top of the anticlinal, being the value in this region smaller than in the neighborhood, due to the fact that the value of $\lambda$ decreases as the value of $\mu$ increases. Analyzing Figure 10 one notices that the horizontal pressure in the less curved layers is lower than in the layers that present significant curvature. On the other hand, in the regions where the layers present greater curvatures, the horizontal pressure presents greater continuity, differentiating well these layers. Regarding the vertical pressure, Figure 11, it is naturally observed that its value increases as the depth of the layer increases and, moreover, the vertical pressure in the regions of anticlines is lower than in its neighborhood, not allowing a good differentiation of the region just below the anticlines. Figure 12 shows the values of the hydrostatic pressure inside the layers. It can be seen that these values increase with depth and allow for very good differentiation between the different layers, with the exception of the regions at the tops of the anticlines. From these observations, it was possible to identify the low pressure zones of the curved layer model described in the paper.

## Acknowledgements

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