Seismic Attenuation Assessment in Passive Seismic Investigations Using Pulse Width Analysis and Considering Source Effects

G.A. Tselentis (University of Patras), P. Paraskevopoulos (University of Patras), N. Martakis (LandTech Enterprise SA), A. Serpetsidaki (University of Patras) & E. Sokos (University of Patras)*

Summary

Knowledge of attenuation can be very useful in many stages of conventional seismic data processing as its removal increases resolution. Recent advances in passive seismic investigations involving microearthquakes indicate that attenuation assessment has a great potential as a direct hydrocarbon indicator and reservoir structural delineation. The purpose of this investigation is to assess the use of pulse width analysis to invert for attenuation structure.

Introduction

Geological structures of hydrocarbon interest are commonly strongly heterogeneous. This heterogeneity affects the seismic wave propagation and may produce phenomena like focusing and defocusing of seismic rays, wave scattering, amplitude absorption etc. Furthermore, a reservoir full of hydrocarbons tends to be acoustically softer than if it is full of an incompressible fluid such as water. All these phenomena are parameterized by the Q-parameter (or quality factor) which accounts overall for the energy decay as a function of distance.

Theoretical models of seismic wave attenuation predict that a wave passing a hydrocarbon reservoir should suffer more attenuation than in surrounding geological materials. Attempts to extract attenuation from conventional seismic data have met with little success since amplitude spectra of the seismic record contain the imprint of the amplitude spectrum of the earth's reflectivity as well as the amplitude spectrum of the seismic wavelet.

The aim of the present investigation is to extend the application of passive seismic methodologies towards evaluating accurate values of the attenuation structure correcting for the seismic source.

Theory

The risetime τ and pulse width ΔT of a seismic signal generated by a finite source duration can be represented by the following nonlinear relations (DeLorenzo et al. 2004)

$$\tau = \tau_0 + \gamma(r_0, V_p, \theta)\frac{T}{Q} + \lambda(r_0, V_p, \theta)$$

$$\Delta T = \Delta T_0 + C\frac{T}{Q}$$

Where $\tau_0 = \frac{r_0}{V_r} - \frac{r_0}{V_p}\sin\theta$;

$$\Delta T_0 = \frac{r_0}{V_r} + \frac{r_0}{V_p}\sin\theta$$

θ is the take off angle and ro the fault radious (assuming circular crack), Vr the rupture velocity, Vp the p-wave velocity, C has been experimentally determined to be 0.5 (Gladwin and Stace, 1974) and the parameters γ, λ are those given by DeLorenzo et al. 2004).

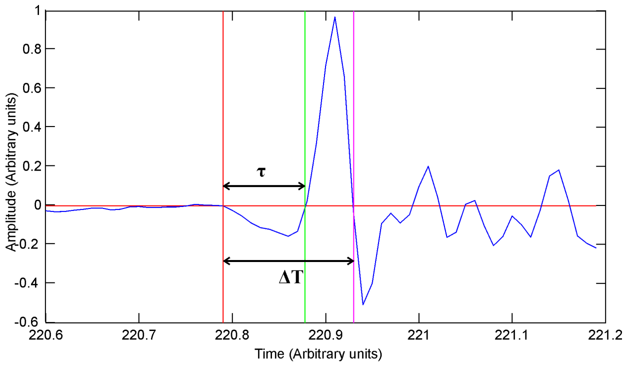


Figure 1: Rise time ΔT and pulse width τ of a seismic phase on a velocity seismogram

a velocity seismogram, τ represents the time duration of the first half cycle of the p-wave phase, while ΔT is the time duration of the complete cycle of the wave (Figure 1). Given N microearthquakes recorded at a network of M receivers, the data space consists of the measurements of rise times τij and pulse widths ΔTij, with i=1,N and j=1,Mi, where Mi the number of receivers that recorded the ith event. The take off angles θij can be expressed in terms of the dip δi and strike φi directions of the plane containing the circular fault crack of the i directions of the plane containing the circular fault crack of the ith event (Aki and Richards 2002).

Let us assume that by applying conventional passive seismic tomographic inversion on our dataset (Tselentis et al. 2007), we have a detailed 3D p-wave velocity model of the region under investigation. By dividing the region into blocks the whole path attenuation term T/Q can be expressed as a sum of the contributions Ti/Qi of each block crossed by the p-wave rays

$$\frac{T}{Q} = \sum_{i}^{P} \frac{T_i}{Q_i}$$

The nonlinear system of equations to be solved is therefore given by

$$\tau_{ij} = \frac{r_{0,i}}{V_r} - \frac{r_{0,i}}{V_p} \sin \Omega(\delta_i, \phi_i) + \gamma_{i,j} \sum_{m=1}^{P} \frac{T_{i,j,m}}{Q_m} + \lambda_{i,j}$$

$$\Delta\mathrm{T}_{ij} = \frac{r_{0,i}}{V_r} + \frac{r_{0,i}}{V_p} \sin \Omega(\delta_i, \phi_i) + 3.9 \sum_{m=1}^{P} \frac{T_{i,j,m}}{Q_m} + \lambda_{i,j}$$

With i=1,N and j=1,M and Ti,j,m representing the travel time of the first p-wave generated by the ith event recorded at the jth station passing through the mth block and Ω given as a function of strike and dip of the fault plane (Aki and Richard 2002)

Since the above equations show a nonlinear dependence in source and attenuation parameters we define the following cost function cf

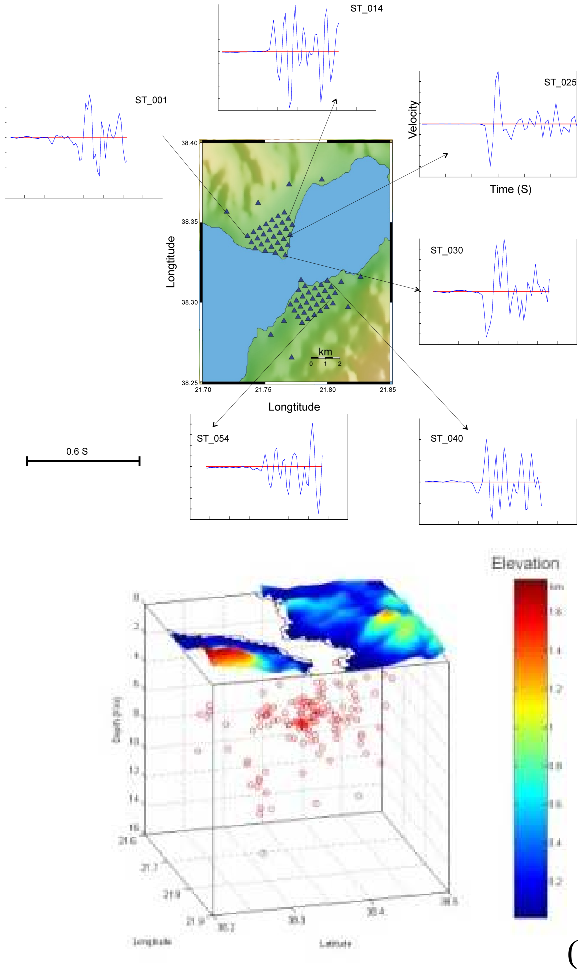
$$cf = \sum_{i=1}^{N} \sum_{j=1}^{M} \left\{ W_{\tau i,j} \left(\tau_{i,jobs} - \tau_{i,jest} \right)^2 + W_{\Delta \mathrm{T} i,j} \left(\Delta\mathrm{T}_{i,jobs} - \Delta\mathrm{T}_{i,jest} \right)^2 \right\}$$

Where $W_{\Delta \mathrm{T} i,j}$ and $W_{\tau i,j}$ are the weight factors.

The next stage is to solve this equation by using a global optimization technique for searching for its minimum, employing the downhill Simplex optimization method (Ichinose et al. 1997). We assume some a priori constraints of the Q for each discretised subsurface block derived from the minimum and maximum slope of the relation of rise time and pulse width versus travel time data.

The data

In this investigation we used the source and receiver configuration of the Rio-Antirio 2004 passive microearthquake experiment (Tselentis et al. 2007). The data set consists of 150 well-located microearthquakes with ML ranging from 0.5 to 3.0 recorded on a network of 70 stations (Figure 2).



(a)

(b)

Figure 2: (a) Seismic station layout (triangles) and some velocigrams of the the first breaks recorded from a local microearthquake (shown in 6 stations) The time segment shown is the same for all velocigrams (b) 3D distribution of microearthquake hypocenters recorded by the local network

Each station consists of a 24bit EarthData PR24 digitizer supplemented with LandTech's LT-S01 three component velocity sensors. The microearthquake hypocenters (figure 2b) and the 3-D P-wave velocity model were obtained from a joint tomographic inversion and are described in Tselentis et al. 2007 (Figure 3).

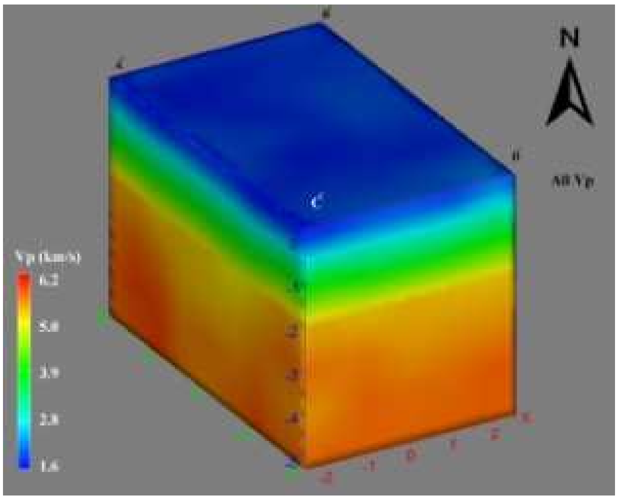


Figure 3: The 3D Vp (P wave) velocity model obtained from travel time tomography.

The basic parameters used in the present study are the rise time τ and the pulse width ΔT of first p-wave arrivals. Figure 4 presents the derived rise times and pulse width from the recorded microearthquakes versus the P wave first break travel time. A general positive trend can be seen, even though a considerable scatter of the data is observed. The increase of τ and ΔT with travel time is usually attributed to ray path attenuation effects. The dispersion of the data around the line of best fit may be due either to directivity source effects and/or spatial variation in the attenuation properties of the sampled by the p-wave rays volume.

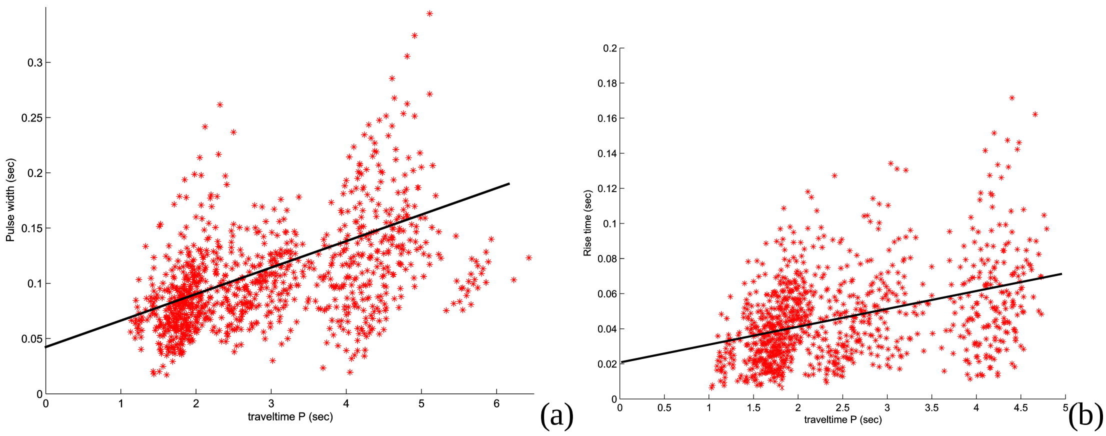


Figure 4: Observed (a) rise times and (b) pulse widths versus travel time.

Characteristic results of the inversion process described previously, as it was applied in the study area are depicted in Figure 5. In this figure only layers with the higher Q values are presented. In this way the “non colored” regions with low Q values can also be delineated. The low Q regions appear to coincide quite well with the earthquake epicenters that were used in the seismic tomography. Also they are in agreement with well known sedimentary and tectonic zones of the region for which we expect higher seismic attenuation.

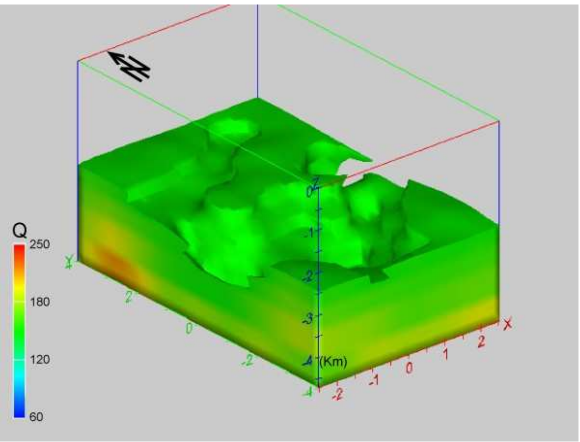


Figure 5: 3D Q structure of the investigated area. Lower Q regions have been removed.

Conclusions

Since the relationship between velocity and attenuation will depend on the composition and crack porosity with depth, in situ measurements of both velocity and attenuation provide important constraints on the lithology and physical properties of hydrocarbon reservoirs and the subsurface in general. Furthermore, the assessment of velocity and attenuation models has important applications to the processing and imaging of seismic reflection data. This includes inverse Q-filtering, Q-phase compensation, amplitude statics and the incorporation of attenuation in migration algorithms.

Acknowledgements

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