SEDIMENT PROPERTIES IN SUBMARINE MASS TRANSPORT DEPOSITS USING SEISMIC AND ROCK-PHYSICS OFF NW BARENTS SEA

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**Supplements: Details on the methods**

**1. Traveltime tomography**

Traveltime tomography provides an adequate elastic velocity model in depth, allowing resolution of vertical as well as lateral velocity gradients and a reliable reconstruction of the geological structures (see Böhm et al., 1999; 2000; Vesnaver and Böhm, 2000; Rossi et al., 2001; 2011).The input of the tomographic algorithm encompasses the reflected traveltimes picked on the prestack seismic data. In the present work, we picked five reflections (R1 to R4 plus the sea-floor) on the stacked section of profile IT-EG08B (Fig. 2). Then, to facilitate identification of target seismic events in the prestack data, we used an initial velocity field obtained by interpolating stacking velocities to estimate traveltimes at the different offsets. Hence, we picked five reflections (R1 to R4 plus the sea-floor), and the direct arrivals of the waves travelling in the sea-water. The traveltime inversion required about 40 iterations to minimize time-residuals and to resolve each layer. Finally, we used the staggered grid procedure of Vesnaver and Böhm (2000) to refine the velocity model, shifting base grids (with a cell size of 1 km) three times along x, obtaining a final grid with cells of 0.3 km. From the comparison between the picked traveltimes and the ones calculated based on the final tomographic model, it appears that more than 90% of the travel time residuals are below 1% of the picked traveltimes. We evaluated the reliability of our inversion by checking time residuals and by applying a picking error of 6 ms to our traveltimes, i.e., the average distance between two adjacent distinguishable reflections in our record, and evaluated the differences in the models, to understand which part of the model is reliable. The picking error of 6 ms implies an error in the velocity estimate lower than ±30 m/s in the whole model, except <10 pixels showing a velocity error close to 80 m/s. The more significant errors are located near the borders of the illuminated area, because of lower ray coverage.

**2. Attenuation tomography**

Seismic waves travelling through the Earth are attenuated, preferentially in the highest frequencies, so that there is a loss of resolution as the seismic wave propagates. The attenuation of P waves is usually quantified by the seismic quality factor (*Qp*), which is proportional to the inverse of the loss of energy per wavelength:

$Q\_{p}=2π\frac{Energy stored}{Energy dissipated per cycle}$.

 (e.g., Sheriff and Geldart, 1995). The larger the quality factor, the better the rock acts as a transmitter of seismic energy (Picotti et al., 2007). The *Qp* value for sedimentary rocks generally ranges between 20 and 200 (Sheriff and Geldart, 1995), and depends on saturation and viscosity of the pore fluids, permeability, and porosity, being even more sensitive than velocity (Best et al., 1994).The reflection coefficient at an interface also depends on the *Qp* of the two adjacent layers so that a high *Qp* contrast can generate slight reflections even in the absence of velocity differences (e.g., Lines et al., 2014). Usually, the seismic quality factor is estimated through the spectral ratio method (e.g., Toksöz et al., 1979), based on the ratio between the spectral amplitudes of the original and attenuated wavelets.

However, the loss in frequency of the seismic wave is bound through a line integral along the ray-path to the seismic wave attenuation factor, as traveltimes to seismic wave velocities, enabling a tomographic approach that provides models of the seismic attenuation in depth (Quan and Harris, 1997; Rossi et al., 2007). The reliability of attenuation tomography, based on the P-wave quality factor *Qp*, is less dependent upon the offset length than another kind of analyses as AVO (amplitude versus offset) or AVA (amplitude versus angle), providing similar information. The method used in this study requires frequency analysis of the same picked events as the travel time tomography so that the velocities previously determined are used both for the ray-tracing and to convert the attenuation factors into *Qp* values. For further details on the technique, we refer to Rossi et al. (2007) and Madrussani et al. (2010). In the absence of a calibrated signature for seismic

sources, we derived a source signature from the sea-floor reflection. In fact, due to the very high Q factor of water, the wavelet’s frequency content is preserved, as reveals a comparison of the spectrum of the sea-floor reflection with the source’s theoretical one. The analysis was done on the whole seismic section, providing a depth model of seismic attenuation within the MTDs and in the surrounding sediments. Frequency decay is calculated between the top and bottom reflections of each layer. Hence, only if the top and bottom of the MTD coincide with the top and bottom of the layer, the obtained *Qp* values represent the intrinsic attenuation of the MTD material. Otherwise, the values are a weighted average of the properties of the MTD and the embedding sediment. On the other hand, above an MTD, the *Qp* can be affected by the scattering at the top of a rough surface. The coarser grid used in *Qp* inversion in respect to the *Vp* one (2.9 km for the base grid, and 1 km for the final staggered grid) allows a smoother and more reliable result (e.g., Madrussani et al., 2010).

The uncertainty in *Qp* values is between 11 and 18. More than 90% of the frequency shift residuals are below 1% of the input values.

We estimated the *Qp*-factor field by applying a frequency shift error of 5 Hz to our input files and evaluated the differences in the models, to understand which part of the model is reliable. The error in the frequency shift analysis of 5 Hz implies an error in the *Qp*estimate lower than ±18 for the whole model.

**3. Physical properties modeling**

To estimate the porosity distribution, we used a velocity-porosity relationship proposed by Carcione et al., (2005). These authors obtained the generalized Gassmann modulus for a multi-phase system consisting of *n* solids (mineralogical components, e.g., quartz, feldspar, and clay) and a saturating fluid. We performed a four-step process: 1) predict effective elastic grain moduli by using Hashin-Shtrickman bounds formula (Hashin and Shtrickman, 1963; Carcione et al., 2006); 2) estimate the dry-rock bulk moduli through Krief’s equations (Krief et al., 1990; Mavko et al., 1998); 3) compute the properties of brine as a function of temperature, pressure and salinity (Batzle and Wang, 1992); and, 4) apply the generalized Gassman’s relation (Carcione et al., 2005, 2006; Gassman, 1951) to obtain P- and S-wave velocities of the fully brine saturated sediments.

The exponential dimensionless parameter A in the Krief relationship (equation 59 of Carcione et al., 2006) is a pore compliance coefficient that depends on the pore shape and Poisson ratio of the matrix. Several types of porosity can be recognized in rocks including rounded pores and thin, elongated ellipsoidal (crack-like) pores which align sub-parallel to grain bedding. Rounded pores are typical of sandstones, while crack-like pores are typical of shales with high clay content. The parameter A takes a value of about 2 for spherical pores, increasing as the pores become ellipsoidal (Le Ravalec and Gueguen, 1996; David and Zimmerman, 2011). Therefore, this coefficient should be calibrated using velocity measurements on core samples or sonic-logs. The dry-rock shear modulus is obtained from the dry-rock bulk modulus assuming that the Poisson ratio of the dry porous rock is equal to the Poisson ratio of the mineral forming the rock frame. This is generally not the case (Le Ravalec and Gueguen, 1996; David and Zimmerman, 2011), but such simplification is used in the absence of further data to calibrate the model. The grain (*ρs*) and bulk (*ρ*) density is the arithmetic average of the densities of the single constituents weighted by the corresponding volume fractions. White (1975) and White et al. (1975) were the first to introduce the mesoscopic-loss mechanism in the framework of Biot’s theory, based on the approximation of a regular distribution of patches. Even if their models were originally introduced to study the attenuation due to partial saturation, where the pore space is saturated both by a fluid and a gas, they could also be used to study the case where the attenuation is due to mesoscopic heterogeneities in the petrophysical properties of the hosting rock. In this work, we employed the first model introduced by White et al. (1975), which is extensively outlined in Carcione and Picotti (2006), but similar results can also be obtained using the second model.

Consider a periodic layered system composed of porous media 1 and 2 with proportions *p1* and *p2*, such that *p1*+*p2*=1. White et al. (1975) obtained the frequency-dependent complex modulus *E* for a P-wave traveling along the direction perpendicular to the stratification. We do not report here the entire theoretical description of White's model because it can be found in Carcione and Picotti (2006). The complex P-wave modulus *E (ω)* is a function of the angular frequency *ω*, the parameters of the saturating fluid (bulk modulus, density, and viscosity) and the parameters of the two rock phases (porosity, permeability, dry-rock moduli, and grain moduli).

Variable porosity implies changes in permeability and dry-rock moduli. Porosity *ϕ* and permeability are related together by a simplified version of the classical Kozeny-Carman relation (Carcione and Picotti, 2006; Mavko et al., 1998). The dry-rock moduli follow from the Generalized Gassmann’s approach described in the 3.2 subsection, whereas the fluid (brine, in our case) viscosity can be obtained using the empirical relations suggested by Batzle and Wang (1992).

Once *E* has been determined, we use the concept of complex velocity to obtain the P-wave phase velocity *Vp* and the quality factor *Qp*. If *ρ1* and *ρ2* are the bulk densities of the two materials, the complex velocity *v* is defined by the relation $E=\overbar{ρ}v^{2}$, where $\overbar{ρ}=ρ\_{1}p\_{1}+ρ\_{2}p\_{2}$ is the averaged density. Then, *Vp* and *Qp* are given by

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(Carcione and Picotti, 2006).

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