Changes in attenuation with depth in an ocean carbonate section: Ocean Drilling Program sites 806 and 807, Ontong Java Plateau

L. Neil Frazer, Xinhua Sun, and Roy H. Wilkens
School of Ocean and Earth Science and Technology, University of Hawaii at Manoa, Honolulu

Abstract. Changes of in situ seismic attenuation with depth are estimated from full waveform acoustic logs in a soft formation (rock with shear velocity less than borehole fluid velocity). The relative attenuation $Q_p^{-1}$ is computed from the variation with depth of the $P$ wave amplitude spectrum of a single source receiver pair. The method is applicable when unknown source and receiver responses of the sonic tool make other methods difficult to apply. Tests with synthetic data generated by a full waveform method verify that in a soft formation the decay of the $P$ wave amplitude spectrum depends mainly on attenuation. The method is applied to Ocean Drilling Program data from holes 806B and 807A on the Ontong Java Plateau in the western Pacific Ocean. Four attenuation logs were computed independently using data from four source receiver pairs. Agreement of the four logs at each site and agreement of results from the two sites suggest that the method is robust and practical. The attenuation logs show a maximum attenuation between 300 and 500 m below the seafloor. They agree well with compilations of high-frequency attenuation versus porosity and frequency in marine sediments but are somewhat greater than results from seismic experiments, possibly owing to the presence of sedimentary microbeds and the three-decade difference between the seismic and sonic frequency bands.

Introduction

The profile of attenuation of seismic waves in the sediments covering the ocean floor is poorly known. Attenuation measurements made in the laboratory at ultrasonic frequencies on often disturbed sediment samples or estimated from seismic experiments over long wavelengths reveal little about the evolution of sediment attenuation characteristics as those sediments transform through compaction and diagenesis from loose grains at the seafloor to sedimentary rocks buried far below the surface. Models of seismic wave propagation, notably those of Biot [1956a, b] and derivatives, predict compressional velocities well, but their attenuation predictions have not been tested because of a lack of in situ data for comparison. Detailed profiles of attenuation are valuable, not only for interpreting seismic data, but also for understanding the rock physics consequences of the diagenetic process. In this study we describe an effective method of estimation of sediment attenuation and apply the method to pelagic carbonates of the western Pacific Ocean at two locations.

For many years the Ocean Drilling Program (ODP) has recorded full waveform data using the Schlumberger long-spaced sonic (LSS) tool. The LSS has two sources and two receivers arranged as shown in Figure 1. Every 0.152 m (6 inches), four microseismograms, with three source receiver spacings, are recorded from the different source receiver pairs (Figure 2). Those data provide an excellent record of the in situ physical properties of the ocean bottom. This study presents a method of calculating relative $P$ wave attenuation from such data. Existing methods of attenuation determination have been reviewed by Jannessen et al. [1985] and Tonn [1991]. Methods relevant to borehole data include spectral ratios [Cheng et al., 1982; Sams and Goldberg, 1990; Tonn, 1991], centroid frequency shift [Quan et al., 1994], filter correlation [Courtney, 1993], partition coefficients, and full waveform inversion [Paillet and Cheng, 1991]. Our attempts to apply some of these methods to data from ODP sites 806 and 807, pelagic carbonate sites on the Ontong Java Plateau in the western Pacific Ocean, gave poor results because the source and receiver responses of the LSS are not known and are not constrained by the data. Also, differences between the frequency spectra of the sources and an unknown formation $S$ wave velocity made it difficult to accurately determine attenuation by waveform modeling. Unable to determine absolute
attenuation, we resolved instead to determine relative attenuation, i.e., the change in $Q_n^{-1}$ with depth. The simple method that we developed to do so can also be used to make an absolute depth profile of attenuation if one is willing to assume attenuation at one depth.

The Ontong Java Plateau is a broad oceanic subsea rise located just north of the equator and paralleling the Solomon Island chain to the east [Shipboard Scientific Party, 1991]. Rising several thousand meters above the surrounding seafloor, it is an ideal locality to study physical properties of pelagic carbonate sediments because (1) it is almost completely free of contamination by terrestrial detritus and (2) its location near the equator assures high carbonate productivity, swamping biogenic silica production. Indeed, for virtually all of the sediment drilled during ODP leg 130, calcium carbonate contents were 80% or greater [Shipboard Scientific Party, 1991; Urmos et al., 1993]. Site 806 is located in 2,520 m of water, and site 807 is in 2,805 m. Site 806 lies within 50 km of the equator, and site 807 is about 450 km to the NNW. Both sites are now and have historically been at depths at or above the calcium compensation depth and have experienced sedimentation rates of the order 20 to 60 m.y.$^{-1}$ [Berger et al., 1991].

Numerous studies have been carried out on both surface and subsurface sediments from the Ontong Java Plateau, including four drilling legs Deep Sea Drilling Project (DSDP) legs 7, 30 and 89, and ODP leg 130. In general, the sediments consist of a matrix of small calcareous fossils (coccoliths, discoaster plates) and fragments of the order of 10 $\mu$m and smaller surrounding larger, mostly hollow foraminifer tests that range from a few tens to 100+ $\mu$m in diameter (Figure 3 and Johnson et al., 1977; Hamilton et al., 1982). Fulthorpe, et al. [1989] found good agreement between sonic logs recorded during DSDP leg 80 and seismic velocity structure of the plateau inverted from sonobuoy records. Mithholland et al. [1980] and Kim et al. [1985] included limestone samples from the Ontong Java Plateau in their laboratory studies of P and S wave velocity and attenuation. Results of Mithholland et al. [1980] and Kim et al. [1985] were later modeled using the Biot-Stoll model [Stoll, 1989] by Hurley and Manghnani [1991]. Urmos et al. [1993] and Urmos and Wilkens [1993] found that density logs from the Ontong Java Plateau agreed well with laboratory determinations from drilling samples and developed a new algorithm for correcting laboratory velocity measurements to in situ values in pelagic carbonates.

**Theory**

In this section we consider a single source receiver pair; the four individual source receiver pairs are processed similarly, and the degree of their agreement is taken as a measure of the reliability of the results. The P wave train is the first arrival on the sonogram (Figure 2). We separate it from later arrivals using a time window tapered at both ends with a half cosine and then estimate its amplitude spectrum with a fast Fourier transform (FFT). In the sequel the term spectrum, appearing by itself, denotes an amplitude spectrum. Our processing procedure is based on the following simple model for the spectrum $X$ of the $P$ wave train:

![Figure 2. Four waveforms recorded by the Schlumberger LSS tool at 240 m below seafloor (mbsf) in hole 807A, Ontong Java Plateau. In this slow part of the formation the $P$ wave arrival is followed quickly by a higher-amplitude borehole fluid arrival (airgy phase). Waveform differences are due, in part, to differences in source receiver transfer function.](image-url)
\[ X(z_r, z_s) = S C_s(z_s) R C_r(z_r) G(z_r, z_s) \]

\[ \exp \left[ -\frac{\omega}{2} \int_{z_p}^{z_s} \frac{dz}{Q_p v_p} \right], \]

where \( \omega \) is radian temporal frequency, \( S \) is the spectrum of the source (transmitter) at depth \( z_s \), \( C_s \) is the coupling of the source to the borehole wall, \( R \) is the spectrum of the receiver at depth \( z_r \), \( C_r \) is the coupling of the receiver to the borehole wall, \( G \) is geometrical spreading, \( v_p \) is formation \( P \) velocity, and \( Q_p \) is formation quality factor.

The coupling functions \( C_s \) and \( C_r \) depend on frequency, but in a uniform hole they are independent of depth. They include the attenuation of the \( P \) wave during transmission through the fluid between tool and borehole wall. The geometrical spreading \( G \) depends strongly on spacing \( \Delta z = |z_s - z_r| \), moderately on frequency [Sams and Goldthry, 1990], and weakly on depth \( z = (z_s + z_r)/2 \). The integral in the exponent is approximated by \( Q_p^{-1} \Delta t \), where \( \Delta t \) is the traveltime of the \( P \) wave in the formation, so \( Q_p^{-1} \) is an average over the source receiver interval. Note that our \( \Delta t \) is a travel time, not a reciprocal velocity as in some papers on well logging.

Rearranging (1) and taking logarithms gives

\[ \frac{2 \ln X}{\omega} - Q_p^{-1} \Delta t + \frac{2 \ln(SC_s RC_r G)}{\omega}. \]

On the right side of this relation the first term depends strongly on \( z \) but only weakly on \( \omega \), while the second term depends strongly on \( \omega \) but only weakly on \( z \). Neglecting the weak dependencies, we rewrite the last relation as

\[ \Phi(z, \omega) = \phi(z) + A(\omega), \]

where \( \Phi = 2\omega^{-1} \ln X \), \( \phi = -Q_p^{-1} \Delta t \), and \( A = 2\omega^{-1} \ln(SC_s RC_r G) \). The left side of (3), \( \Phi(z, \omega) \), is the data; on the right side of (3) the depth profile \( \phi(z) \) is what we wish to extract, and \( A(\omega) \) is an unknown depth-independent shift of \( \phi \).

Regarding \( z \) and \( \omega \) as the row and column indices of the data \( \Phi \), the basic idea of the method is to average the columns of \( \Phi \) so as to obtain a maximally accurate estimate of the shape of \( \phi \) without concern for any depth-independent shift in \( \phi \). Later, the best estimate shape of \( \phi \) is shifted so as to give a particular value of attenuation at some reference depth. The attenuation profile is then recovered as \( Q_p^{-1}(z) = -\phi(z)/\Delta t(z) \), where \( \Delta t(z) \), is obtained from the velocity log.
In processing the columns of $\Phi$, we use a median rather than a mean in order to reduce errors due to the weak depth dependence of $A$. Care is required because the median is more vulnerable to variations in $\Phi$ with $\omega$. To see why this is so, imagine taking the median of three depth profiles that do not intersect: the middle profile is always selected by the median; but if that middle profile is given a constant shift so that it becomes an outer profile, then another profile will be selected. To avoid such problems, we proceed as follows. First, we average the columns of $\Phi$,

$$\bar{\Phi}(z) = \frac{\text{mean}}{\omega} \{\Phi(z, \omega)\}. \quad (4)$$

Then we calculate the median offset of each column from the average,

$$\delta\Phi(\omega) = \text{median}_z \{\Phi(z, \omega) - \bar{\Phi}(z)\}. \quad (5)$$

Next we shift each column to bring it closer to the average

$$\Phi_s(z, \omega) = \Phi(z, \omega) - \delta\Phi(\omega). \quad (6)$$

Finally, we take the median of the shifted columns

$$\hat{\Phi}(z) = \frac{\text{median}}{\omega} \{\Phi_s(z, \omega)\}. \quad (7)$$

The resulting $\hat{\Phi}(z)$ is an estimate of $-Q_p^{-1}(z)\Delta t(z)$ plus an unknown constant $\hat{A}$

$$\hat{\Phi}(z) = -\Delta t(z)Q_p^{-1}(z) + \hat{A}. \quad (8)$$

In order to obtain the value of the unknown constant $\hat{A}$, let $\zeta$ be some reference depth at which we are willing to guess at the value of $Q_p$. Setting $z$ equal to $\zeta$ in (8) and solving for $\hat{A}$ gives

$$\hat{A} = \hat{\Phi}(\zeta) + \Delta t(\zeta)Q_p^{-1}(\zeta). \quad (9)$$

On the right side, $\hat{\Phi}(\zeta)$ is known from our processing so far, $\Delta t(\zeta)$ is known from the well log, and we have specified the value of $Q_p^{-1}(\zeta)$; hence $\hat{A}$ is now also known.

Next we recover the profile of $Q_p^{-1}\Delta t$ by use of (8) in the form

$$Q_p^{-1}(z)\Delta t(z) = \hat{A} - \hat{\Phi}(z) \quad (10)$$

Substituting values of $\Delta t(z)$ from the velocity log and solving for $Q_p^{-1}$ gives the attenuation profile.

Figure 4. Verification of (3) using synthetic data. Here $\Phi$ calculated from fullwave synthetics is plotted as a function of relative attenuation $Q_p^{-1}$ at different frequencies for a source receiver separation of 3.05 m (10 feet). The similar trend of the $\Phi$ curves at different frequencies validates the noise reduction procedure of median averaging $\Phi(z, \omega)$ profiles over different values of $\omega$ to obtain $\Phi$. Parameters $v_f$, $Q_f$, and $\rho_f$ are the velocity, $Q$, and density of the borehole fluid, respectively.
\[ Q^{-1}_p(z) = \frac{\hat{A} - \hat{\Phi}(z)}{\Delta t(z)} \]
\[ = \frac{\hat{\Phi}(-\xi) - \hat{\Phi}(z) + \Delta t(\xi)Q^{-1}_p(\xi)}{\Delta t(z)} \tag{11} \]

Note that a lower bound for the attenuation profile is obtained by zeroing the reference attenuation \( Q^{-1}_p(\xi) \). Also note that the attenuation profile was obtained from a single source receiver pair.

The final step is to apply the procedure just given to the remaining three source receiver pairs and then median average the four profiles into a single profile. Note that depth shifting of the profiles before averaging is unnecessary, as the 2.44- and 3.66-m pairs have the same center and the two 3.05-m pairs are offset from the center by the same amount in opposite directions. Deviations of the four individual profiles from the average profile provide an estimate of error at each depth.

Tests on Synthetic Data

Before proceeding further, we test the validity of our simple model for the \( P \) wave process using full wave synthetic seismograms with a realistic source and receiver. Synthetic \( P \) wave arrivals were processed in the same manner as the real data. In particular, a complete synthetic seismogram including all arrivals was computed using equations from Cheng [1989] for an unceded, fluid-filled hole in a solid formation with appropriate causal frequency-dependent attenuation and phase velocities. The \( P \) wave train was extracted from the synthetic seismograms by time windowing in the same way as for the data. We used the source function described by Tsang and Rader [1979] with complex spectrum \( S \) given by

\[ S(\omega) = \frac{8\omega_0(\alpha - i\omega)}{[(\alpha - i\omega)^2 + \omega_0^2]^2} \tag{12} \]

in which \( \omega_0 = 2\pi \times 15 \text{ kHz} \) and \( \alpha = 0.5\omega_0/\pi \). Shear wave \( Q \) was kept fixed at the value 30.

Equation (3) is verified by showing that \( \Phi \) varies linearly with attenuation when frequency and the values of the other formation parameters are kept fixed (Figure 4). The formation parameters on Figure 4 were chosen as typical of soft marine sediments. (A formation is said to be soft if its shear wave velocity is less than the velocity of the borehole fluid.) Although we do not know their exact values, shear wave velocities equal to or greater than seawater velocity would result in unrealistically low Poisson's ratios for the sediments. Figure 5 shows contours of \( \Phi \) as a function of attenuation and

![Figure 5](image_url)

**Figure 5.** Insensitivity of \( \Phi \) to shear wave velocity: contours of \( \Phi \) in (3) as a function of \( Q^{-1}_p \) and \( v_s^{-1} \) at 10 kHz for synthetic data. Formation parameters other than \( Q_p \) and \( v_s \) were fixed as shown in Figure 4.
shear slowness $v_s^{-1}$ at frequency 10 kHz for fixed values of the other formation parameters. It was feared that $\Phi$ might be as sensitive to shear wave slowness as it is to compressional attenuation, but Figure 5 illustrates that this is not the case for the soft formations of interest here. The calculations shown in Figures 4 and 5 indicate that the simple $P$ wave propagation model of (1)-(3) is adequate for an uncased hole.

The next set of tests examines whether our procedure will work over the range of formation parameters of ODP hole 806B. The actual $v_p$ and density logs from hole 806B are shown in Figure 6. For synthetic tests these were approximated by linear increases with depth: $v_p$ increased linearly with depth from 1.8 km s$^{-1}$ at 200 m below seafloor (mbsf) to 2.6 km s$^{-1}$ at 700 mbsf, and density increased linearly from 1.6 Mg m$^{-3}$ at 200 mbsf to 1.9 Mg m$^{-3}$ at 700 mbsf. Smooth shear wave velocity profiles (not shown) were obtained by fixing Poisson’s ratio. In making the synthetics we used the sine wave $Q_{P}^{-1}$ log indicated by solid lines in Figures 7a and 7b. Other model parameters are the same as shown in Figure 4. For each source receiver spacing we calculated synthetic waveforms at 5-m depth intervals from 200 to 700 mbsf. The time-sampling interval of the synthetic waveforms was 10 $\mu$s, the same as that of the data. Finally, we inverted for attenuation using the procedure given above. Inversion results are shown as dashed, dotted, and dash-dotted lines in Figure 7.

Here are some of the details of the inversion procedure. We picked $P$ wave trains from the data using a time window beginning at the first arrival of the $P$ wave refraction, as predicted by the following formula:

$$t(z) = \frac{2r}{v_f \cos(\sin^{-1}(v_f/v_p))} \frac{\Delta z - \tau \tan[\sin^{-1}(v_f/v_p)]}{v_p},$$

where $r$ is the radius of the borehole, $\Delta z$ is the source receiver spacing, and $v_f$ is the velocity of the borehole fluid. The window size is important because in soft marine carbonate sediments the $P$ wave velocity is very low, approaching the borehole fluid velocity, and so the pure $P$ wave train is short. In our application a time window of 20 data points (200 $\mu$s) was sufficient to characterize the $P$-wave.

The spectra of the $P$ wave trains were calculated by an FFT after padding the time series with zeros to 128 points. We computed $\Phi$ in (3) at 26 frequencies in the interval from 5 to 25 kHz, the band of the $P$ wave source spectrum. We then used (4)-(8) to calculate $\Phi$. The reference depth was chosen automatically by a rule in which the maximum value of $\Phi$ corresponds to a maximum $Q_{P}^{-1}$, in this case, set to 100. Then we used (9) to calculate $A$ and (11) to recover the $Q_{P}^{-1}$ profile. Figure 7a compares the results of our procedure (dashed, dotted, and dash-dotted lines) with the true attenuation profile (solid lines); it can be seen that the true $Q_{P}^{-1}$ log and the calculated $Q_{P}^{-1}$ logs are very similar and that the calculated $Q_{P}^{-1}$ logs for different source receiver pairs are almost the same. We calculated the correlation between the true $Q_{P}^{-1}$ log and the calculated $Q_{P}^{-1}$ logs for different source receiver spacings and for different Poisson’s ratios. Roughly speaking, the correlation between the calculated and true profiles increases slightly with increasing source receiver spacing and with decreasing Poisson’s ratio and is generally greater than 0.85. The relative insensitivity of the method to Poisson’s ratio, consistent with Figure 5, is shown in Figure 7b.

Finally, in order to test the stability of our method, we added random noise to the synthetic data. The noise was generated by taking the spectrum of the complete synthetic seismogram and randomizing its phase. Such
noise is an approximation to signal-generated noise, as the signal and the noise have the same amplitude spectrum. The noise level is expressed as the ratio of the noise waveform root-mean-square amplitude to the $P$ wave maximum amplitude. For given noise this ratio tends to be smaller than many other so-called signal to noise measures. The calculated $Q^{-1}$ logs for the noise-free case, the 15% noise case, and the 15% noise case after smoothing over depth with a 19-point median filter are illustrated in Figure 8. Depth points are 0.152 m apart, so 19 depth points correspond to one 3.05-m depth interval. It can be seen that the method is robust up to the 15% noise level, which is adequate for our purposes, as the noise levels in the actual ODP sonic data are less than 5%.

Results from Data

We first applied our technique to data from ODP hole 806B. Hole 806B was logged in 2520 m of water from 120
to 700 mbsf. (During ODP logging operations the bottom of the drill pipe is typically left at about 100 mbsf to avoid the possibility of the pipe dragging off the hole in poorly consolidated near-seafloor sediments.) We used a fixed 200-μs window length to cut the $P$ wave trains from the full waveforms. The edges of the window were half cosine tapered over six points (60 μs).

We calculated $P$ wave spectra for each waveform from 120 to 720 mbsf at the depth interval of 0.152 m. As noted above, we used 26 frequencies from the $P$ wave spectrum in the band 5 to 25 kHz. As these data are from soft carbonate sediments, in which shear wave velocity is less than fluid velocity, we assumed that the minimum attenuation would be in the upper part of the profile, where the sediment behaves much like a fluid. From other experiments [e.g., Fu et al., 1996], $Q_p$ of shallow ocean bottom carbonate sediments is of the order of 80 – 100. Thus for each source receiver pair the shallow half of the $\Phi(z)$ profile was searched for its maximum (corresponding to a minimum of attenuation). Then a 19-point median of $\Phi(z)$ was taken, centered about this maximum. The depth selected by the median was the reference depth $\zeta$, and the profile $\Phi(z)$ was shifted to have the reference attenuation at depth $\zeta$. The four source receiver profiles were then combined at each depth by discarding the two extreme values of $Q_p^{-1}$ and averaging the two interior values of $Q_p^{-1}$. The resulting attenuation profiles, nearly identical for the different source receiver pairs, are shown in Figure 9.

Inversion results are relatively insensitive to $P$ wave window length for windows greater than 150 μs and less than 250 μs; this is verified by Figure 10. The trade-off is that too short a window results in a less accurate estimate of the spectrum, whereas too long a window risks inclusion of non $P$ energy. In shallow ocean bottom sediments, formation $P$ velocity is close to the borehole fluid velocity, so waves in the fluid arrive very soon after the $P$ wave. The effects of varying the value of $Q_p^{-1}$ at the reference depth are shown in Figure 11a. Setting $Q_p^{-1}$ equal to zero at the reference depth (dashed line) gives a lower bound profile of attenuation and thus an upper bound profile of $Q_p$. The solid and dotted lines are for reference attenuations of 0.01 and 0.02, respectively.

An upper bound for the estimated error in $Q_p^{-1}$, assuming $Q_p = 100$ at the reference depth, is shown in Figure 11b. The left curve in Figure 11b is the difference of $Q_p^{-1}$ values obtained from synthetic sonograms with Poisson’s ratios of 0.35 and 0.45. The right curve is the attenuation profile used to compute the synthetics.

### Attenuation Versus Depth

Waveforms from hole 807A were inverted for $Q_p^{-1}$ following the same procedure used for hole 806B data. The results of the two inversions are plotted in Figure 12. The two profiles from different holes separated by 450 km exhibit remarkable similarity. Both begin with relatively high $Q_p$ (40-50) at 100 mbsf, gradually decreas-
ing with depth to a minimum at around 300 mbsf. At both sites most of the decrease in $Q_p$ correlates with the region identified as the ooze-to-chalk transition by shipboard sedimentologists [Berger et al., 1991]. Above the transition, sediments are virtually un cemented and deform easily (ooze). The transition is described as the region within which layers with increased stiffness (chalk) first appear. (It should be noted that this definition of chalk includes materials much softer than the “chalk” we use to write on blackboards.) Within the transition zone the chalk first appears in discrete layers several centimeters thick, separated by ooze layers of tens of centimeters to meters in thickness. Gradually, deeper into the transition zone, the thickness and proportion of chalk layers increase until, at the base of the transition, the formation is mostly made up of stiff material. The transition is primarily due to increasing amounts of calcium carbonate cements in the sediment. Porosity decreases somewhat across the transition zones but does not interrupt the compaction gradient between

Figure 9. $Q_p^{-1}$ logs calculated from four source receiver pairs from ODP hole 806B using a 20-point (200 μs) window. (a) S2-R1 pair (2.44 m), (b) S2-R2 pair (3.05 m), (c) S1-R1 source receiver pair (3.05 m), and (d) S1-R2 pair (3.66 m).
Figure 10. Effects of $P$ wave window length on calculated $Q_p^{-1}$, showing, window lengths of (a) 150 $\mu$s, (b) 200 $\mu$s, (c) 250 $\mu$s, and (d) 300 $\mu$s. Results shown are a median over the four source receiver spacings.

100 and 500 mbsf (see Bassinot [1993] and Urmos et al. [1993], for profiles). Deeper in the section, at approximately 600 mbsf at each site, $Q_p$ begins to increase. At these depths, mechanical consolidation has ceased and porosity decreases slowly by chemical processes of dissolution and precipitation of calcite. These are hard, well-cemented sedimentary rocks, with $P$ wave velocities of the order of 2.5 km s$^{-1}$. Interestingly, a zone of possible enhanced cementation, where chert was seen by core describers in hole 806B, is at the center of an interval of relatively higher $Q_p$ centered around 500 mbsf.

The shape of the $Q_p$ profile is not unexpected. Goldberg et al. [1985] and Wilkens et al. [1992] saw a similar decrease in $Q_p$ with depth in acoustic waveform logs us-
ing two different methods. The site they investigated, DSDP hole 612A, exhibited a $Q_p$ decrease from values around 60 at shallow depths to a minimum of about 20 at a little over 400 mbsf. Hamilton [1976] first suggested that an attenuation increase with depth was probable in some situations. He predicted a maximum attenuation at 200 mbsf, based primarily on laboratory measurements of attenuation versus porosity and on profiles of porosity versus depth. In loose, high-porosity sediments the dominant mechanisms of attenuation are probably related to fluid motion relative to grains or to the frame work of loosely connected grains [Stoll, 1989]. With burial, however, points of contact between grains increase and friction becomes a more important contributor to loss of seismic energy [Hamilton, 1976; Stoll, 1989; Wilkens et al., 1992]. With increasing depth, overburden pressure on the sediment framework, as well as increasing cementation, eventually limits relative movement between grains of sediment and decreases attenuation [Hamilton, 1976]. This appears to be the behavior we see in our profiles of $Q_p$. Our in situ data agree with the predictions of Hurley and Manghnani [1991], based on Biot-Stoll theory, but show considerably less attenuation than the Kim et al. [1985] and Mitholland et al. [1980] measurements of carbonate samples from the lower depths of DSDP hole 289 on the Ontong Java Plateau. This may be because those measurements were

**Figure 11a.** Results of the attenuation inversion process for different values of the reference attenuation at the same reference depth. Solid line is for a reference $Q_p^{-1}$ of 0.01; dashed line is for a reference $Q_p^{-1}$ of zero; and dotted line is for a reference $Q_p^{-1}$ of 0.02 at 186 mbsf in the unaveraged profiles. Profiles have been smoothed over depth with a median filter of length 9.15 m to reduce overlap.

**Figure 11b.** How accurate is the relative attenuation profile? The left curve shows the jump in $Q_p^{-1}$ that would result from a jump in the unknown Poisson's ratio from 0.35 to 0.45. The calculation was made using the $Q_p^{-1}$ profile shown on the right.
not performed at elevated confining pressure, allowing friction to play a more significant role in the laboratory than in the field.

The few published profiles of seismic $Q_p$ inverted from marine seismic refraction experiments in fine-grained sediments (summarized by Bowles [1996]), are generally greater than what we see in our data by a factor of 5 to 10. Bowles [1996] discusses the spread of data seen in the profiles and concludes that they do not warrant a general curve fit, citing lack of detail about the nature of the sediments in the published reports and the difficulty of finding $Q_p$ from seismic data in general. All of the profiles shown by Bowles [1996] do, however, show an increase in attenuation with depth (decrease of $Q_p$). In particular, the profiles of Jacobson et al. [1981] and Brienzo [1992] behave very much like our Ontong Java Plateau profiles, although the absolute levels of their attenuation are lower.

The phenomenon of sonic data giving greater attenuation than seismic data is well known in the petroleum industry [S. Leaney, private communication, 1996] where it is often attributed to the frequency dependence of $Q_p$ or to the effects of microbedding. In fact, microbedding causes $Q_p^{-1}$ to increase very rapidly as the seismic wavelength approaches the thickness of the average microbed. Theoretical and numerical model-
ing [Frazer, 1994] show that for almost any distribution of sand/shale microbeds the \( Q_p^{-1} \) due to the microbeds is roughly proportional to \( r^2 f / f_c \), where \( r \) is the sand/shale reflection coefficient and \( f_c \) is the frequency at which wavelength equals the average microbed thickness. The constant of proportionality increases if there is any roughness of the sand/shale interface [Frazer, 1994, Figure 10]. Thus it would be very surprising if sediment \( Q_p^{-1} \) values inferred from sonic data at 5-25 kHz were not considerably greater than sediment \( Q_p^{-1} \) values inferred from seismic data at 10-100 Hz.

**Attenuation Versus Frequency and Porosity**

The effects of frequency on attenuation, and whether it is proper to use a single \( Q_p \) at all frequencies are still open questions extensively discussed in the literature (e.g., Hamilton [1972], Stoll [1989], Bowles [1996], and many others). Our data, in the kilohertz frequency band that lies between laboratory measurements and seismic data, are plotted versus frequency in Figure 13. Also shown are the zone of data compiled by Hamilton [1972, 1976] and the seismic results compiled by Bowles [1996]. The Ontong Java Plateau data lie within the upper bounds of the Hamilton data field, which was determined largely by laboratory experiments or experiments in the field in loose, near-surface sediments, with the exception of a few data in the seismic frequency range (50-250 Hz). The constant slope of unity in the Hamilton zone of data is consistent with a frequency independent \( Q_p \), although some of the more recent seismic data suggest that there may indeed be a frequency dependence of attenuation between seismic and acoustic frequencies.

In predicting that attenuation might well increase with depth in silty sediments down to 300 mbsf, Hamilton [1976] was considering the relationship of his compilation of attenuation to porosity versus depth profiles. Attenuation data from hole 806B plotted with Hamilton’s [1972] best fit curves and data limits are shown plotted versus well log porosity in Figure 14. (To remain consistent with Hamilton [1976] and others who have reproduced his figure, we have used velocities from the well logs to convert \( Q_p \) into \( k = 8.686 k (v_p Q_p) \). The units of \( k \) are \( \text{dBm}^{-1} \text{kHz}^{-1} \).) We have divided our data into ooze, transition, and chalk (Figure 12). Our data are in general agreement with the earlier values. Coincidence of the fields would improve even more if we were to consider only the porosity of the sediment matrix instead of total sediment porosity. Bachman [1984] has shown that empty foraminifer tests in calcareous sediments behave like solid grains and that the porosity within those grains (interTest porosity) does not significantly affect acoustic properties. Lind [1993] and Wilkens and Urmos [1996] have determined from images like Figure 3 that sites 806 and 807 average about 5-10% interTest porosity. Removing that interTest porosity would effectively slide the porosity of our data to lower values.

The fact that chalk attenuation values fall below the Hamilton curves is an indication of the difference in the materials that were measured to provide the data. All of the Hamilton compilation data were soft sedi-

---

**Figure 13.** Attenuation versus frequency. Data from this study fall within the band of data first published by Hamilton [1972]. Individual data points are seismic data from the compilation of Bowles [1996].
ments. The lower-porosity Hamilton data were, in fact, mostly from sand-sized sediments, owing to the empirical relationship that equates grain size and porosity in surficial sediments [Hamilton, 1972]. Our data, on the other hand, come from sediments that do not change grain size but, rather, are reduced in porosity because of both mechanical and chemical compaction. Deeper, lower-porosity materials are under higher pressure and have undergone a greater degree of cementation than those overlying them. It is interesting, however, that porosity (i.e., friction) still seems to be the most important influence in $Q_p$ behavior.

Lack of agreement with seismic $Q_p$ is probably due to the effects of sedimentary microbeds and the three-decade difference between the seismic and sonic frequency bands.

Acknowledgments. X. Sun and R. H. Wilkens were supported by contracts from the National Science Foundation Ocean Drilling Office (OC E0503352) and from a CORE/NRL Distinguished Visitor artist award (R. H. Wilkens) to work with the Coastal Benthic Boundary Layer research group at NRI, Stennis. N. Frazer gratefully acknowledges support from the Office of Naval Research and from the Petroleum Research Fund of the American Chemical Society. SOEST contribution no. 4166.

Conclusions

Relative in situ $P$ wave attenuation can be estimated from full waveform acoustic logs, even when source and receiver responses are unknown. If one is willing to specify the value of attenuation at a single depth, then the relative attenuation profile becomes an absolute attenuation profile. Assigning zero attenuation to the least attenuative depth on the profile gives a profile that is a lower bound for attenuation at each depth, i.e., an upper bound profile for $Q_p$. Tests with full wave synthetic seismograms show that the method is robust with respect to variations in source receiver spacing, window length, and signal-generated noise. It is also robust with respect to unknown formation parameters such as Poisson's ratio, provided the formation is soft.

In the pelagic carbonate section on the Ontong Java Plateau, $Q_p$ decreases with depth to the bottom of the ooze-to-chalk transition, remaining low for the next few hundred meters before beginning to increase again. $Q_p$ values from the inversion process agree well with previous studies of loose sediments, although the few reported values of seismic $Q_p$ are somewhat higher.


L. N. Frazier, X. Sun, Department of Geology and Geophysics, School of Ocean and Earth Science and Technology, University of Hawaii at Manoa, 2525 Correa Road, Honolulu, HI 96822.

R. H. Wilkens, Hawaii Institute of Geophysics and Planetaryology, School of Ocean and Earth Science and Technology, University of Hawaii at Manoa, 2525 Correa Road, Honolulu, HI 96822. (email: wilkens@noest.hawaii.edu)

(Received May 6, 1996; revised October 4, 1996; accepted October 11, 1996.)