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ATTENUATION FROM VSP DATA COLLECTED ON MELVILLE ISLAND[1]

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ABSTRACT

Seismic waves are observed to attenuate while propagating through anelastic inhomogeneous media. In this study, the variation of intrinsic attenuation with depth is investigated in the neighborhood of a 5-km-deep borehole located on Melville Island in the Canadian Arctic. The data from a vertical seismic profiling (VSP) survey performed in the well is analysed with the spectral ratio method. The results indicate good correlation between a highly attenuating zone and a sandstone unit of the Lower Triassic Series, located between 1400 m and 2500 m.

The specific attenuation for the region is found to be approximately 0.045 ($Q \simeq 22$) with a standard deviation of 0.040. The lack of source normalization during recording is the reason for the large standard deviation. However, in the Lower Triassic layer the sandstone has an anomalously high specific attenuation of 0.105 ($Q \simeq 10$) with a standard deviation of 0.058. Immediately above 4000 m there exists a second low-Q zone where the specific attenuation is 0.083 ($Q \simeq 12$) with a standard deviation of 0.042. The Q values are lower than those found in the North Sea basin.

From an analysis of seismic velocity dispersion it was concluded that scattering of seismic energy and the effects of intrabed multiples may be the cause of at least 60% of the observed attenuation. Nonetheless, excellent agreement between the lithology and Q profile indicate that intrinsic attenuation is being observed.

INTRODUCTION

The crust of the earth has many properties that are of interest to researchers in geophysics, geology and geochemistry. In exploration geophysics the important properties include the compressional and shear seismic velocities, the density of a rock, its magnetic characteristics, electrical conductivity, porosity, and levels of gas or fluid saturation. A property that has recently been shown to be useful in the exploration for oil is the degree of anelasticity of a rock. The more anelastic a material the greater is the intrinsic attenuation of an elastic wave passing through it. The intrinsic attenuation, $a(\omega)$, is given by,

$$a(\omega) \sim \frac{\omega}{2\,v\,Q} \qquad (Q >> 1)\ , \qquad (1)$$

where the specific attenuation is proportional to the ratio of the energy lost in one cycle, ΔE, to the peak strain energy of the cycle, E, as,

$$\textbf{Specific Attenuation:} = \frac{1}{Q} = \frac{\Delta E}{2\pi E}\ . \qquad (2)$$

v is the speed of the seismic wave, ω is angular frequency, and Q is the dimensionless quality factor.

To date, many laboratory experiments have demonstrated that intrinsic attenuation correlates well with rock properties such as the level of fluid saturation, microstructure, the viscosity of pore fluid, and porosity. These parameters are difficult to obtain from in-situ rocks. However, there are indications that the amount of intrinsic attenuation undergone by a travelling wave will yield information regarding these parameters.

In exploration geophysics waves of seismic frequencies are used because they are easy to generate and they propagate large distances within the earth. Laboratory attenuation experiments use sonic to ultrasonic frequencies because of the small size of laboratory rock samples. Seismic and sonic signals do not attenuate in the same manner, thus, in-situ measurements of seismic attenuation are required for direct application in exploration geophysics. Very few such surveys have been conducted. Consequently, it remains to be shown that in-situ experiments can reliably yield intrinsic attenuation measurements.

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This study attempts to extract values of intrinsic attenuation from seismic data collected in situ, and to correlate the results with the local lithology. The lithology of the study area is known, and it is important to be able to associate anomalous attenuation measurements with specific rock types known to have certain rock properties.

A number of experiments have been conducted in an attempt to find Q values for a variety of rock types under different levels of porosity, fluid and gas saturation, degree of consolidation, temperature, and level of hydration. The experiments have been conducted both in the laboratory and in the field and, while a great deal more work is required in this area, the likelihood of being able to discern the lithology of a borehole, or the level of fluid/gas saturation, via its Q-log is high (Stainsby and Worthington, 1985).

Winkler and Nur (1979) suggest that one would be able to monitor steam injection zones by measuring attenuation since this quantity is very sensitive to the level of fluid saturation in porous media. In their studies on Massilon sandstone they found P-wave velocities to vary approximately 15% between fully saturated and dry rock. However, the attenuation was noted to vary by 500% or more depending on the confining pressure. In addition, they speculate upon the usefulness of attenuation studies in earthquake prediction. Winkler and Nur (1982) indicate that with knowledge of the velocity profile, the porosity, and other rock parameters including specific attenuation, it may be possible to determine fluid content, permeability, and microstructure of the lithology, information which is of interest in exploration.

In addition to the above, seismic inversion, deconvolution, and the quality of synthetic seismograms would be improved with a more complete understanding of attenuation (Kan *et al.*, 1982).

The vertical seismic profiling survey (VSP) is the best approach for analysing the elastic properties of in-situ rocks because it samples the waveform at many depths and can, therefore, spatially monitor the attenuation. Few attempts involving field data have so far been successful in obtaining attenuation profiles from VSP surveys. Hauge (1981) used the spectral ratio approach to calculate a cumulative attenuation profile for five wells in the southern United States with the aid of a VSP survey. Hauge's profile showed a correlation of high attenuation with porous sands in contrast with neighbouring shales.

Ganley and Kanasewich (1980) analysed the results of a check-shot survey conducted in the Beaufort Sea. They applied the spectral ratio method over two independent depth intervals arriving at a value for Q between 62 and 73 for the depth interval 945 to 1311 m and between 42 and 45 for the more shallow interval of 549 m to 1193 m. Other studies include those of Kan *et al.* (1982) and Dietrich and Bouchon (1985). Stainsby and Worthington (1985) obtained a Q value of 25±3 for the zone lying between 1500 m and 1700 m in a North Sea well.

It is apparent from the literature that there is a lack of substantial evidence for the ability of in-situ seismic attenuation measurements to yield information about lithology and other rock parameters. However, the potential exists for obtaining such information, and it is hoped that the present study will exhibit this.

The consensus among current researchers favors the viscous squirt flow mechanism as being the major cause of intrinsic seismic attenuation. This is a fluid flow mechanism. Consequently, it is believed that saturation, pore aspect ratio, and porosity of a medium are the dominant parameters governing intrinsic attenuation.

The spectral ratio method

Data collected from a vertical seismic profiling survey is analysed in this study with the spectral ratio method. The essential relation of this method is,

$$Q \approx \frac{-\pi\,(z'' - z')}{v\,m} \tag{3}$$

for $Q >> 1$. Here, z' and z'' are the depths of the upper and lower boundaries of a zone respectively, v is the interval velocity of the zone, and m is the slope of the log of the spectral ratio versus frequency curve, given by,

$$m = \frac{1}{\Delta f}\,\Delta\{\ln\,[\,\frac{U''(\omega)}{U'(\omega)}\,]\}\ . \tag{4}$$

$U'(f)$ and $U''(f)$ are the amplitude spectra of the pulses recorded at depths z' and z'' respectively.

In essence then, a value for Q can be obtained for a depth interval if one obtains the amplitude spectrum of a pulse recorded at the top of the interval and one for the same pulse that has travelled to the bottom of the interval with known phase velocity, v.

The spectral ratio method is successful when applied to pulses propagating in a theoretical homogeneous anelastic medium, as demonstrated by Ganley and Kanasewich (1980). However, when interfaces were incorporated into the medium such that pulses were reflected and transmitted at each boundary, situations occurred whereby other pulses, *i.e.*, intrabed multiples (or short-path multiples), were superimposed on the pulse being analysed, thus forming a wavelet. The wavelets at the top and bottom of a depth interval displayed unique interference patterns because the intrabed multiples are different at each depth. When the resulting wavelets were analysed the plot of the logarithm of the amplitude ratio versus frequency (hereafter called the log plot) oscillated about a straight line. Figure 1 shows examples of oscillations based on field data from this study, where the solid line is the least-squares linear regression curve. Ganley and Kanasewich (1980) found that they could suppress the effects of intrabed multiples (*i.e.*, the oscillations and alteration of the slope of the regression curve) and could obtain a more reliable value for Q if they divided the amplitude ratio of the wavelets by the amplitude ratio of two corresponding synthetic wavelets, $S'(\omega)$ and $S''(\omega)$, that did not include the effects of absorption and dispersion but did include the effects of the intrabed multiples (see Ganley (1981) for their algorithm). Hence,

$$m = \frac{1}{\Delta f}\,\Delta\left\{\ln\left[\frac{U''(\omega)}{U'(\omega)}\,\frac{S'(\omega)}{S''(\omega)}\right]\right\} \tag{5}$$

where equation (3) remains valid. The logarithm of the quotient, when plotted against frequency, yielded a more reliable value for the slope.

The program used to generate the synthetic seismograms required that the Earth model be horizontally layered and that the rays reflect from, and transmit through, each interface at normal incidence. These restrictions are not too limiting because a great number of geophysically interesting areas and surveys meet these conditions.

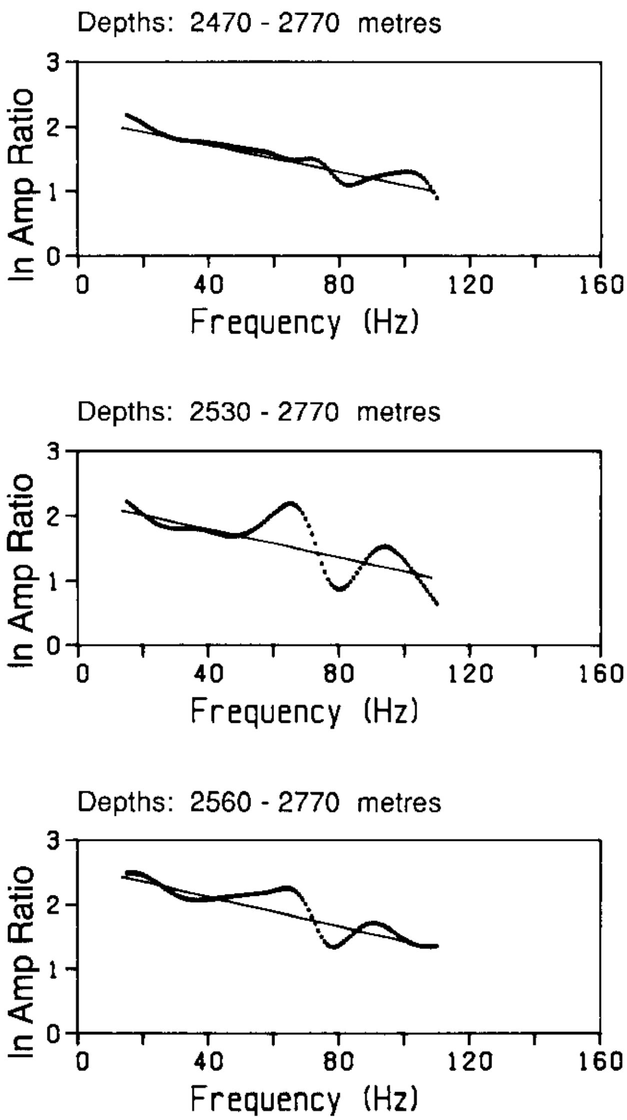

Fig. 1. Example of the slope and oscillatory patterns observed in this spectral ratio study, (top) slope = -0.0104, interval velocity = 4399 m/s, Q = 20, (middle) slope = -0.0108, interval velocity = 4446 m/s, Q = 16, (bottom) slope = -0.0116, interval velocity = 4464 m/s, Q = 13.

THE FIELD DATA

Panarctic Oils Ltd. *et al.* conducted a VSP survey in a well on Melville Island which is located on the southern side of the Sverdrup basin in the Canadian Arctic. The name of the well is Sherard Bay F-34 and a map of its location is shown in Figure 2. Unfortunately, the survey was conducted on three separate occasions during 1983 and 1984. The depth interval 397 m to 1505 m was shot on November 1, 1983; the traces between 1505 m and 3819 m were shot on January 8, 1984; and those below 3819 were recorded on April 6, 1984.

There were several appealing features of the survey which made obtaining an attenuation profile feasible. The trace separation was of the order of 30 m for most of the well allowing for a relatively high depth resolution. The sources were located 150 m from the head of the well allowing for a vertical travel path of the seismic energy to the receiver. This, in combination with the horizontal nature of the geologic layers, resulted in energy approaching all horizons at near-normal incidence. Reflection and transmission coefficients are frequency-dependent at non-normal incidence. Therefore, the geometry of the survey was suitable for a Q analysis. Figure 3 illustrates the lithology of the well where all major rock types are included. Zones of multiple rock types are shown as a sum of the components. The geologic periods are also indicated in Figure 3. The well extends to a depth of 5.4 km below the surface allowing for an extensive survey. The length also lends itself to a high signal-to-noise ratio at depth, since Rayleigh waves resulting from the near-offset shot attenuate rapidly with depth.

The traces provided by Panarctic Oils Ltd. *et al.* were all single-shot and generated by one of two types of dynamite sources, Aquaflex or Geogel. The VSP section from Sherard Bay is shown in Figure 4. There is a trace approximately every 30 m down the length of the

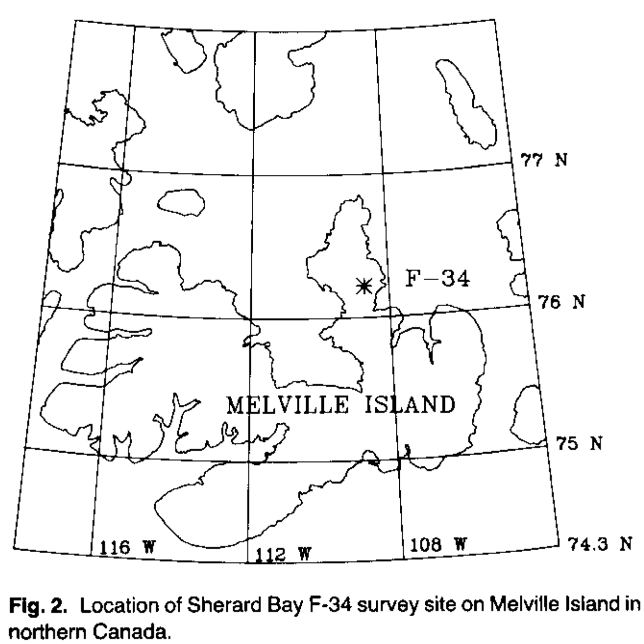

Fig. 2. Location of Sherard Bay F-34 survey site on Melville Island in northern Canada.

hole except between 2800 and 3800 m, where the separation generally varies between 100 and 200 m. The raw data, in many cases, had sufficient power below 160 Hz to provide reliable Q estimates. In addition, only the direct arrivals were analysed, since they contain the widest frequency band of all the pulses within a trace. This is due to their relatively short travel path from shot to receiver, resulting in reduced attenuation of the high frequencies. They also contain less noise, as pointed out by Spencer *et al.* (1982). A sequence of raw first-arrivals from both Aquaflex and Geogel sources is shown in Figure 5. The raw data was recorded for 4 s at a rate of 1 sample per millisecond.

The interference correction factor, referred to earlier, requires the generation of synthetic traces. To calculate synthetics, the variation with depth of layer thickness, P-wave velocity, and rock density, must be known, that is, the Earth model must be specified. The velocity profile is an important part of the Q estimation since the interval velocity is an explicit term in the equation for Q, see equation (1), and its accuracy is important for the determination of the correction factor.

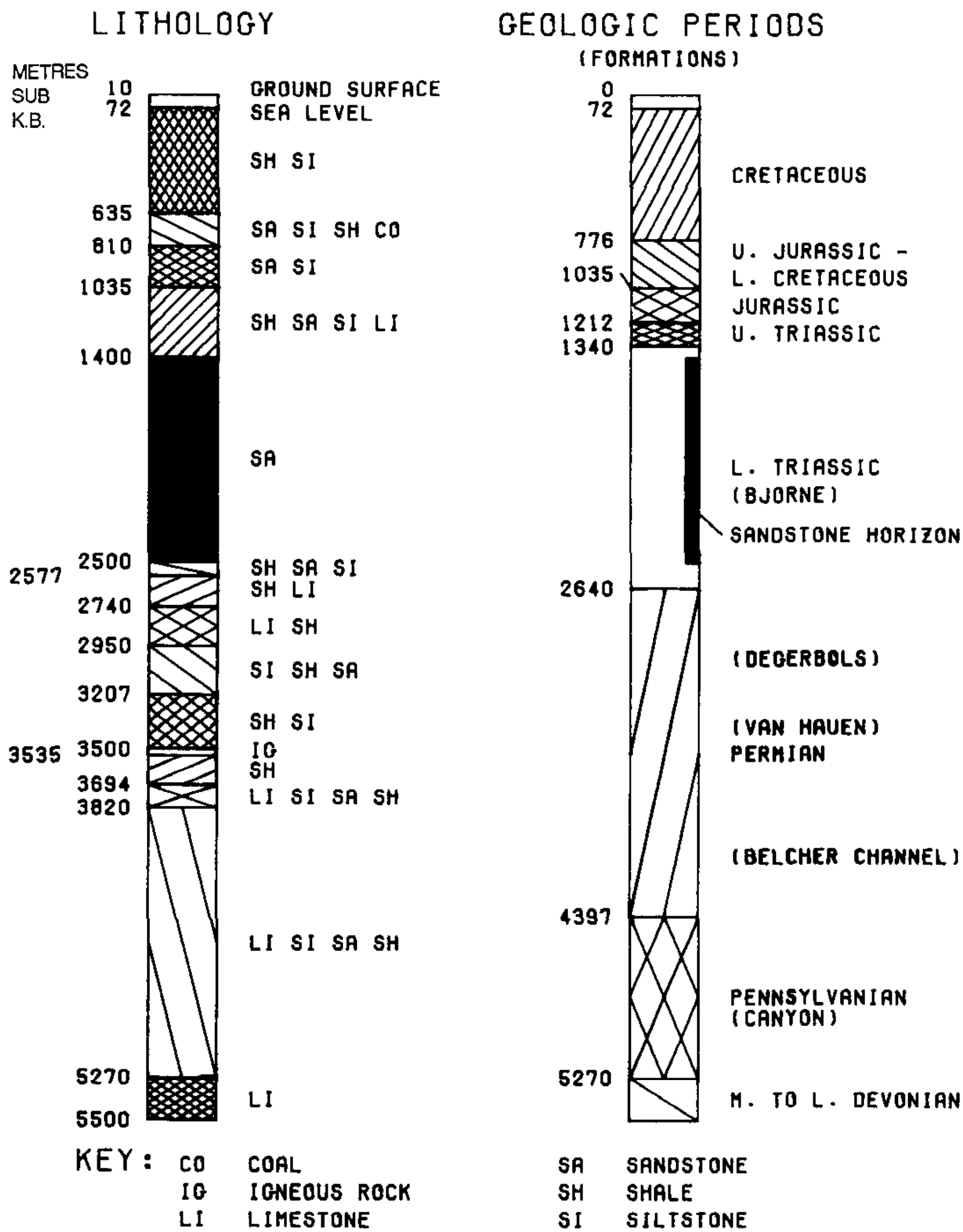

Fig. 3. Lithologic and geologic cross-sections of Panarctic et al. Sherard Bay F-34 survey site.

It is often found that a seismic signal requires more time to travel to a particular depth than is suggested by the integrated sonic traveltime. This indicates that the velocity profile, as given by the sonic survey, does not apply to waves of seismic frequencies. In fact, for the most part, the discrepancy between the sonic arrival time and the seismic arrival time grows with depth, which is a phenomenon called "drift". There have been various explanations for drift as presented by Stewart *et al.* (1984). They quantified the drift caused by variations in layer thicknesses, intrabed multiples, time picking of first arrivals, and attenuation (and dispersion). In their final analysis they concluded that the drift is due mainly to velocity dispersion between the seismic and sonic frequency bands, with the effect of intrabed multiples playing a minor role.

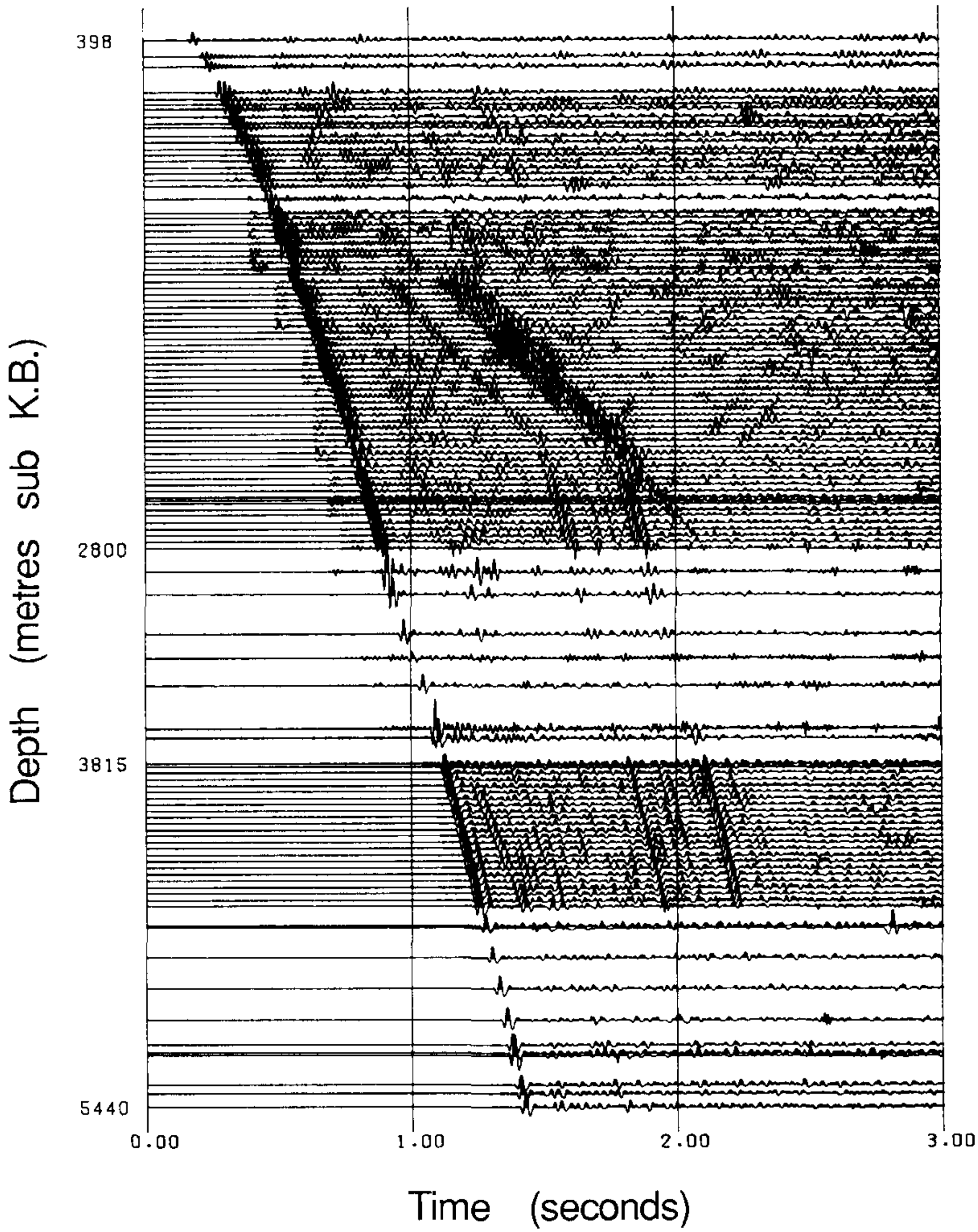

Fig. 4. VSP section from Panarctic et al. Sherard Bay F-34.

The method by which the exploration industry corrects for drift involves performing a check-shot survey. A check-shot survey is the same as a VSP survey except that only the first breaks are recorded and the spacing between receiver depths increases from 30 m to 100 m or more. Panarctic, consequently, used their VSP data to correct for the drift that was present in their velocity data.

Q was calculated using the interference correction factor during an initial trial survey. It was found that, although the correction factor slightly altered the character of the log plot, the oscillations of the curve were not reduced. It was concluded that either the dominant oscillations of the log plot were caused by the differences between the sources used by Panarctic, or the geologic interfaces were rough or contorted in such a manner that the synthetic seismograms did not adaquately resemble the VSP traces. Such contortions would not be apparent from the sonic profile from which the synthetics were calculated and may have scattered energy in an unpredictable fashion. As a result of this, the correction term was not used in any further analysis.

Generation of Amplitude Spectra

The length of the first arrival, or window length, was determined from the field data whether or not synthetic traces were used in the Q estimate. A subjective estimate of the length of the first arrival was required because there was no point along a trace where it was clear that the first arrival had ended. An optimum window length is required such that oscillations in the log plot are minimized while at the same time the entire direct arrival is retained. A number of values ranging from 50 to 200 ms were tested. An optimum value of 125 ms was determined from the degree of linearity of the log plot. When the synthetics were included the same window lengths were applied to them.

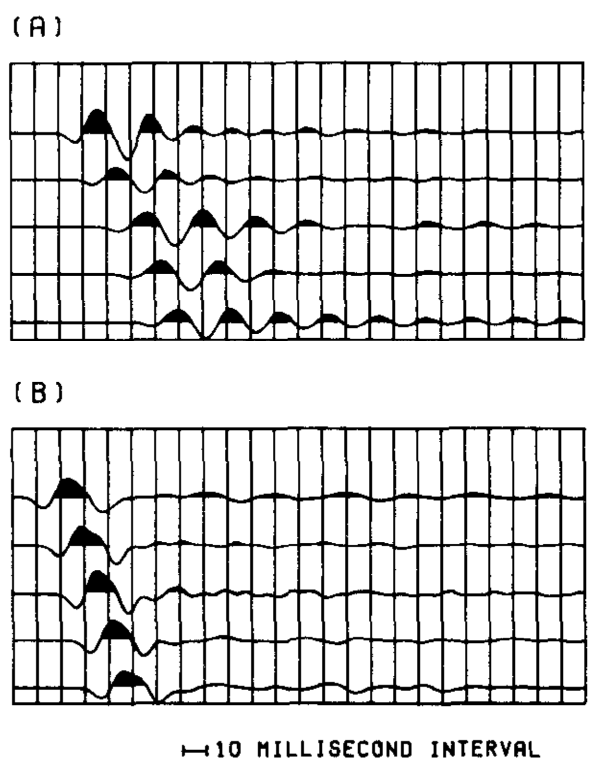

Fig. 5. Sequences of first arrivals with 30-m depth intervals, a) those generated by the Aquaflex source (top trace is 2110 m sub-K.B.) b) those generated by the Geogel source (top trace is 3831 m sub-K.B).

Although it was known that Panarctic employed two types of explosive sources, Geogel and Aquaflex, during the course of the survey, it was not clear from the documentation which traces corresponded to a particular source. Only in the lower portion of the well was there no ambiguity since Geogel was the only source specified. In the process of selecting pairs of traces for analysis, it was important not to compare traces that originated from different sources. Geogel and Aquaflex had quite distinct spectral characteristics which made them incompatible.

After the generation of several spectra with window lengths of 125 ms, the source used in each instance became apparent. Not only were there differences between the spectra of the two sources, but it was found as well that the Geogel source consistently caused first arrivals of shorter duration than the Aquaflex. Consequently, all Geogel-related traces were analysed with a window length of 80 ms. Refer to Figure 5 for a comparison of Aquaflex and Geogel first arrivals and to Figure 6 for a comparison of Aquaflex and Geogel amplitude spectra. The differences between these spectra are also observed when comparing spectra more closely related in depth. Neither the resulting Q values nor their reliability were overly dependent upon the window length. In this study, window lengths may be safely varied within ± 10 ms of the above values. Because the choice of window length is dependent on the sharpness of the pulses obtained, the window used will vary from survey to survey. Hauge (1981) used a 30 ms window, and Ganley and Kanasewich (1980) used a 200 ms time window.

The truncation of a trace after the first arrival, in virtually all cases, will cause a sharp discontinuity between the end of the window and the subsequent zeros. This, in turn, artificially adds high frequencies to the spectrum. It is possible to minimize the effects of truncation by applying a cosine bell to both ends of the window (Kanasewich, 1975). In this study, a 10-point cosine bell was applied to both ends of the data window. The effects of the cosine window on the spectrum of the time window are small as its Fourier transform is very smooth with most of its power restricted to below 20 Hz. Furthermore, the window began 5 ms (5 points) before the first break. In addition to reducing the effect of the cosine bell in an area of the trace where it is of little use, this allowed for errors in picking the first break. Nonetheless it was applied, due to the limitations of the program algorithm.

The resulting time series was Fourier transformed and the amplitude spectrum was calculated. This procedure was carried out on all four traces (or, if no correction factor was used, only the two field traces) and the natural logarithm of the amplitude ratios at each frequency was calculated in accordance with equation (4) or equation (5), as required.

Calculation of the slope, m, of the log plot involved two steps. First, the minimum and maximum frequencies at which both spectra had sufficient power to be reliable were determined. Typical extremes were 15 Hz and 160 Hz, with the minimum upper limit of 80 Hz in a few instances. Secondly, the slope was approximated by finding the slope of the corresponding linear least-squares line as was done by Ganley and Kanasewich (1980) and Hauge (1981).

During an initial survey, it was found that the log plot was quite noisy, so a 9-point (approximately 9 Hz) running average of the amplitude spectra was included before the ratios were taken (see Figure 6). This resulted in a smooth curve with all major oscillations still present. No smoothing of the log plots themselves was required. The straight line, that the theory predicts, did not materialize in most cases. Instead, an oscillatory curve with a linear trend was observed (see Figure 1).

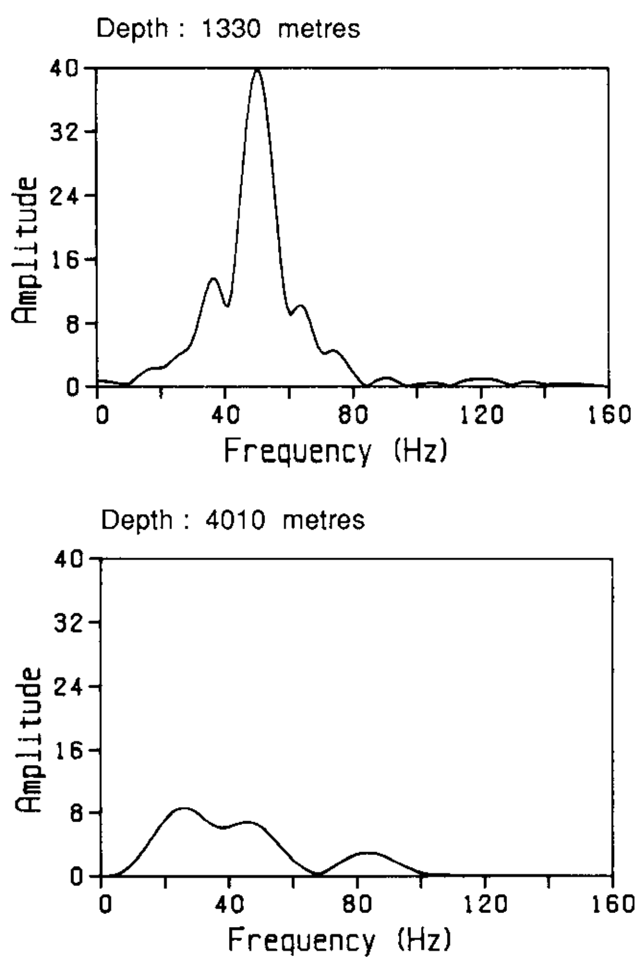

Fig. 6. Comparison of first-arrival amplitude spectra generated from the Aquaflex (top) and Geogel (bottom) sources.

The Q Profile

Very small levels of attenuation are difficult to estimate from seismic data. It was felt that the attenuation of a signal over a distance of 200 to 300 m was measurable and that this provided a high enough depth resolution to be of value. Dietrich and Bouchon (1985) considered a depth interval greater than 100 m as an appropriate figure. At depths greater than 3800 m the separation was increased to 300 to 400 m between traces in case Q increased. In general Q tends to increase with depth reducing the attenuation to levels less than that measurable over small receiver separations. In this study, Q did not increase appreciably, so a depth resolution of 200 to 300 m was added to the data set below 3800 m to maintain the resolution of the more shallow section. Once the more shallow trace was chosen, a compatible trace 200 to 300 m further down the borehole was sought.

Twenty-one different shot holes were used by Panarctic with no indication from the documentation as to which hole was used on a particular occasion. Subtle differences between holes may have been present. Also, the preparation of the dynamite may have varied from shot to shot. Powder densities and density variations will certainly be unique for each shot, and this will result in differences among individual source amplitude spectra. A reliable Q value was obtained by choosing pairs of spectra whereby the deeper spectrum was simply a copy of the more shallow one but multiplied by a ramp with any negative slope. Often this was a difficult condition to assess and several physically nonrealizable values were obtained.

On those occasions where a negative value for Q arose, the curve was dismissed. Negative Q's are physically unrealizable and imply that the source amplitude spectra were not similar. Just as negative Q's were dismissed, so too were extremely atypical Q values. If the majority of results in a depth interval were in close agreement while a few were not, the anomalous values were considered to have no validity. The definition of "anomalous" was taken from Margenau and Murphy (1956) who suggested that residuals exceeding 5 times the probable error be rejected on the grounds that such large errors are not random, but rather are erroneous. A total of 11 points were rejected on this basis. All those points retained in the attenuation profile are displayed in Figure 7.

Because attenuation is directly proportional to the specific attenuation, Q^{-1}, in relation (1), it is customary to show the variation of this parameter with depth. Figure 7 is a plot of the reciprocals of the Q values for the Sherard F-34 well. Each of the 108 points is plotted at the middepth of the interval over which it was collected. The relative absence of points between 2800 m and 3800 m is due to the large receiver spacing used during the survey between these two depths.

A prominent feature is immediately apparent at 2100 m where the specific attenuation peaks to a maximum for the hole. More detail is found by smoothing the profile with a running average. Each point in Figure 7 was replaced by the weighted arithmetic mean of all points lying within 100 m to either side. Thus, a running average with a window of 200 m was performed (see Figure 8a). The mean value was subsequently plotted at the same depth as the point in question. The following points were then processed in the same manner. Consequently, the number of points in the profile remained 108. The largest source of error in the calculation of specific attenuation comes from the determination of the slope arising from the log plot. The reliability of the interval velocity and trace separation is considerably higher than that of the slope obtained from the least-squares linear regression curve. With this in mind, each point in the profile was weighted by the inverse of the standard deviation of the slope of the respective regression line.

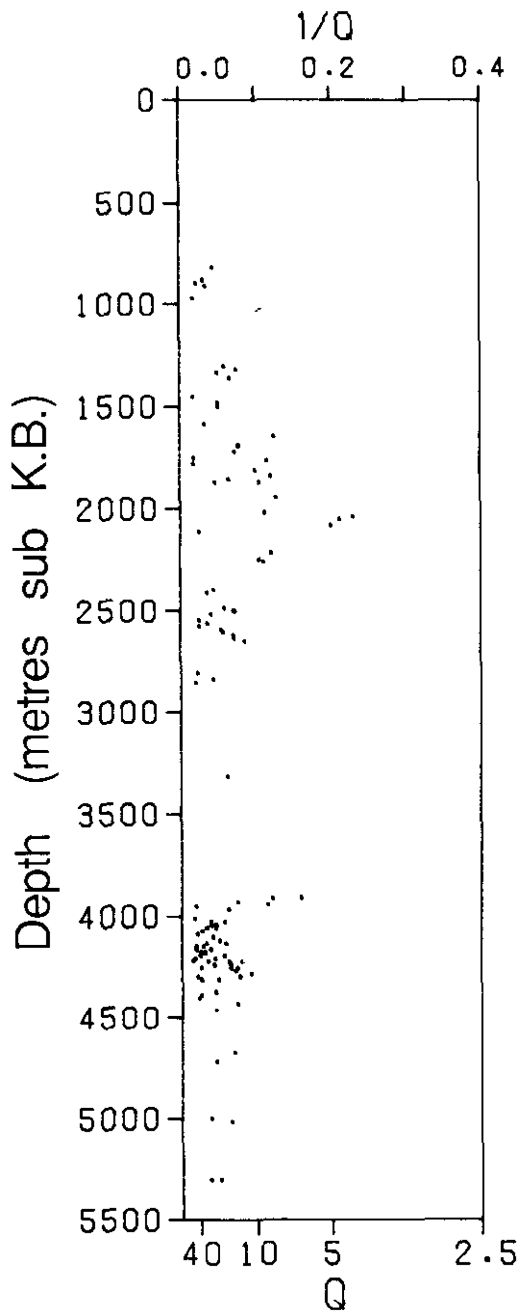

Fig. 7. Plot of specific attenuation versus depth with accompanying Q-axis.

The profile of Figure 8a exhibits an interesting peak between 1900 and 2300 m and an indication of a zone of high attenuation immediately above 4000 m. Consider Figure 8 in its entirety where the interval-velocity, lithologic, specific attenuation, and interval-Q^{-1} profiles can be reviewed simultaneously. The interval-Q^{-1} profile was obtained by finding the weighted arithmetic mean of those Q^{-1} values lying within a specified depth interval in Figure 7. The depth intervals were selected according to distinct Q^{-1} zones as apparent from part (a). The lithologic section was duplicated from Figure 3 and the key remains unaltered. The interval-velocity profile was generated from the sonic log where a 5% change in velocity was required to define the commencement of a new layer.

From Figure 8, the weighted means of the specific attenuations for each depth interval are given in Table 1 where the standard deviations were calculated with the unweighted means. The number of data points in each interval is given along with the range of Q values. The range was determined from the standard deviation (STD) of the specific attenuation values in each depth interval. The final column in Table 1 gives the most probable quality factor for each interval given by the inverse of the specific attenuation values in column three. One set of values for the entire well is presented in the bottom row of the table.

Interpretation

The strongly attenuating zone between 1930 and 2320 m is associated with a virtually uninterrupted Lower Triassic sandstone layer which extends from 1400 m to 2500 m. The remainder of the sandstone layer seems to coincide with a slightly elevated attenuation level. Hauge (1981) shows clearly that attenuation increases with increasing sand content of rocks. From the results of a VSP survey he found that most of the attenuation took place in sandstone layers while relatively little occurs in the shale which comprised the remainder of the hole. Stainsby and Worthington (1985) give further examples of porous

DEPTH INTERVAL (metres)	# OF POINTS	1/Q	INTERVAL 1/Q STD	Q - RANGE	MOST PROBABLE Q
750 to 1035	5	0.024	0.008	31 - 62	41
1035 to 1550	7	0.046	0.016	16 - 33	22
1550 to 1930	13	0.055	0.035	11 - 50	18
1930 to 2320	9	0.105	0.058	6 - 21	10
2320 to 3813	18	0.04	0.019	17 - 48	25
3813 to 4000	6	0.083	0.042	8 - 24	12
4000 to 5400	50	0.038	0.019	18 - 54	26
750 to 5400	108	0.045	0.04	12 - 217	22

Table 1. Weighted means of the specific attenuations and Q's for distinctive depth intervals.

sand units being associated with high attenuation. It would appear, therefore, that the results obtained in this study further substantiate these findings.

The suggestion of a highly attenuating zone immediately above 4000 m is not related to the presence of sandstone; however, in accordance with the results of Winkler and Nur (1979, 1982) and Murphy III (1982), it could be interpreted as a partially saturated zone (≃95%) in contrast with fully saturated or dry neighbouring rocks. A change in microstructure may, in addition, contribute to the high attenuation, in that pore aspect ratios might be smaller with respect to the overlying and underlying layers. The viscosity of the pore fluid may increase in the low-Q zone, as may the porosity. These variations in rock properties may also be responsible for the varying attenuation within the sandstone interval.

It should be noted that the attenuation observed here will approximate the upper limits of intrinsic attenuation due to the inclusion of frequency-dependent scattering and the effects of intrabed multiples. A Q value of 26 is likely to be too low a value for depths greater than 4 km. Scattering and intrabed multiples are thought to

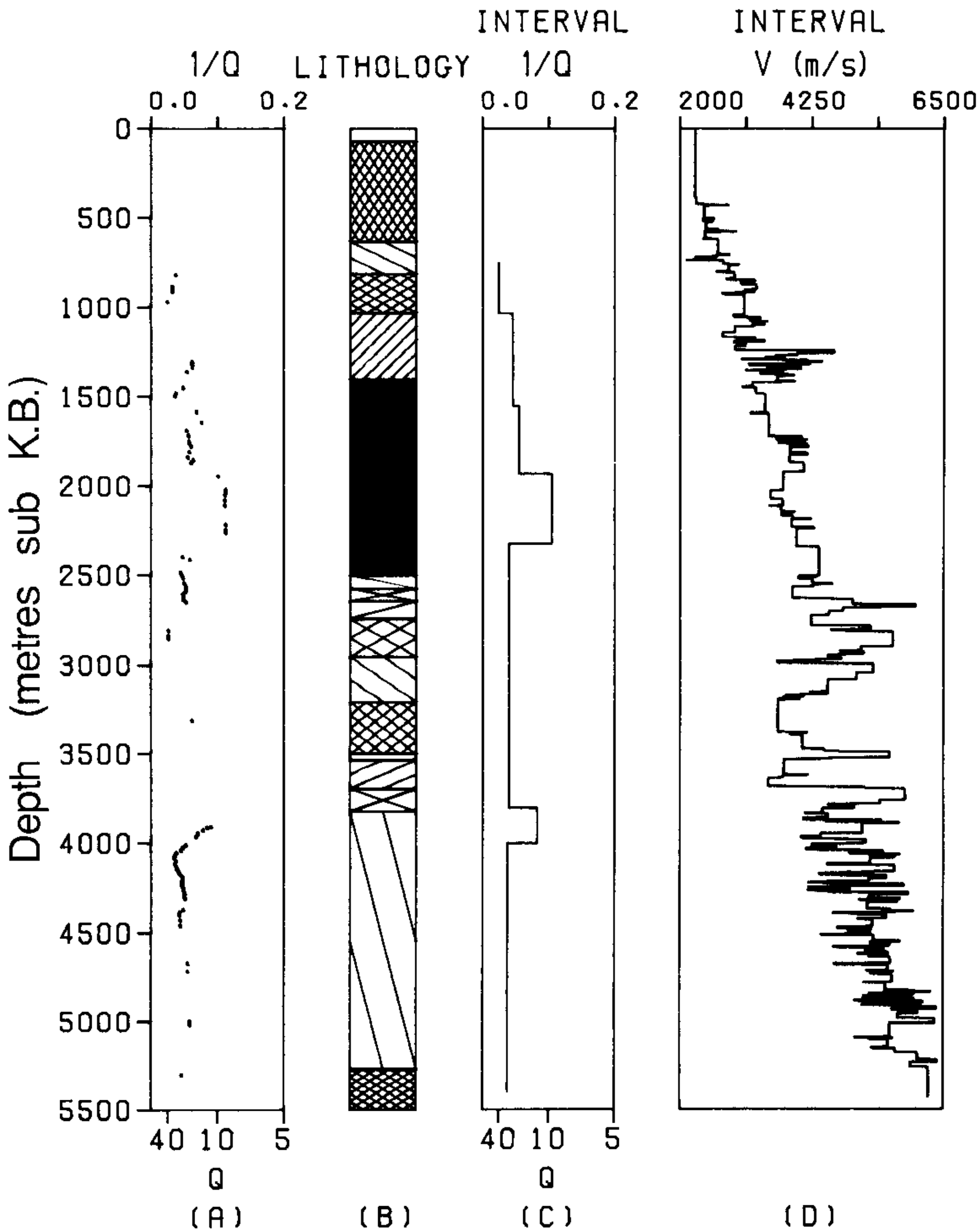

Fig. 8. Profiles: a) weighted $1/Q$ smoothed with a 200 m running average b) lithology c) interval $1/Q$ d) interval velocity.

contribute to this value because of the highly cyclical nature of the velocity log (Spencer *et al.*, 1982). It is possible that some scattering is due to irregular interfaces; we conclude this from the failure of the correction factor discussed earlier. The sandstone layer, however, as can be seen in the velocity and geologic profiles, is relatively homogeneous. Thus, scattering and the number of intrabed multiples may be minimal in this zone. The highly cyclic stratigraphy between 1200 and 1400 m and below 2500 m could be depleting the signal of higher frequencies, hence, inflating the attenuation at these depths. This serves to reduce the contrast in attenuation between the sandstone layer and bordering zones. If the effects of scattering and intrabed multiples could be eliminated, the contrast between the sandstone interval and neighbouring layers would likely be enhanced.

The Effects of Apparent Attenuation

Ganley (1979) obtained phase velocities that agreed well with Futterman's theory of velocity dispersion (1962). By considering Futterman's third dispersion relation, it is possible to approximate the percentage attenuation that is a result of intrabed multiples and scattering versus that which is due to intrinsic attenuation. According to Richards and Menke (1983), the apparent attenuation due to intrabed multiples and intrinsic absorption are approximately additive. At the same time, it will be possible to see whether or not the observed drift between seismic and sonic arrival times can be accounted for by the attenuation observed. In the following discussion it must be noted that attenuation is a necessary and sufficient condition for dispersion (Futterman, 1962) and that dispersion accounts for the vast majority of the drift (Stewart *et al.*, 1984; Jain, 1986).

Over the depth interval 750 m to 3695 m the F-34 well exhibits a drift (sonic arrival time minus seismic arrival time per kilometre) of -8 ms/km. Between 3695 m and 4600 m the seismic and sonic velocities are roughly comparable while below 4600 m the seismic velocities are greater than the sonic velocities. The behaviour of the velocities below 3695 m is not explained by Futterman's theory and may result from reduced sonic tool or geophone-ground coupling, or errors in time picking of the first arrivals. Stewart *et al.* (1984) found an average drift of -6.6 ms/km in the literature they reviewed. In 5 wells they themselves studied (4 wells in the Anadarko basin of the southern United States and 1 well in east Texas) intrabed multiples caused a drift of -6.6 ms/km while dispersion (or intrinsic attenuation) resulted in a drift of -23 ms/km.

Futterman's third dispersion relation is given by,

$$v(\omega) = c\left[1 - \frac{1}{\pi Q_0}\ln\left(\beta\frac{\omega}{\omega_0}\right)\right]^{-1} \tag{6}$$

where $v(\omega)$ = phase velocity for a disturbance of angular frequency ω

ω_0 = an angular frequency below which no attenuation takes place (let it be $2\pi \times 10^{-3}$ sec^{-1})

c = phase velocity for frequencies below ω_0 (a constant)

$\beta = \ln\gamma = 1.78107248\ldots$
γ is Euler's constant

Q_0 = reduced quality factor (a constant)

$$= \frac{\omega}{2\,a(\omega)\,c} = Q(\omega) + \frac{1}{\pi}\ln\left(\beta\frac{\omega}{\omega_0}\right) \tag{7}$$

$Q(\omega)$ = frequency-dependent quality factor

$a(\omega)$ = frequency-dependent attenuation coefficient

An estimate of the reduced quality factor, Q_0, can be obtained if the phase velocity at each of two known frequencies is available. Because c is a constant, equation (6) gives

$$v(\omega_1)\left[1 - \frac{1}{\pi Q_0}\ln\left(\beta\frac{\omega_1}{\omega_0}\right)\right] = v(\omega_2)\left[1 - \frac{1}{\pi Q_0}\ln\left(\beta\frac{\omega_2}{\omega_0}\right)\right]\ . \tag{8}$$

Let ω_1 be a seismic frequency and $v(\omega_1)$ be the phase velocity of a seismic wave. Down to the depth of 3695 m the average phase velocity is 3622 m/s as calculated from the traveltime of a pulse generated by a Geogel source.

Futterman points out that it is the highest dominant frequency component of a pulse which arrives first. In view of Figure 6 the velocity of 3622 m/s will be applied to a frequency, ω_1, of 80 Hz. ω_2 can be assigned to the frequency used in the sonic survey which was 20 kHz. By considering the integrated sonic log prior to correction by the VSP survey, this frequency travelled with an average velocity of 3731 m/s over the same depth interval.

Solving equation (8) for Q_0 gives

$$Q_0 = 64$$

for the zone lying between 750 m and 3695 m. By applying equation (7) it is possible to determine Q for a given frequency. The mean Q down to a depth of 3695 m is 23. Because both Aquaflex and Geogel played an equal part in obtaining this value, and because this Q should be associated with the dominant frequency of the spectrum, the average of the two dominant frequencies, 50 Hz for Aquaflex and 40 Hz for Geogel (see Figure 6), was the frequency for which this Q-value was considered to apply. Thus, $Q(45\text{Hz}) = 23$ between 750 and 3695 m. Equation (7), however, renders $Q(45\text{Hz}) = 60$ based on the observed drift (assuming that intrinsic attenuation caused the majority of the drift). It seems likely that the discrepancy in Q values is due almost entirely to intrabed multiples and scattering because the Q obtained from drift measurements will be weakly affected by such processes (Stewart *et al.*, 1984, in conjunction with Futterman, 1962), whereas that arising from the spectral method is strongly affected. This implies that intrabed multiples and scattering contribute about 60% of the observed attenuation.

It must be kept in mind that there is strong evidence of an attenuation peak in the mid-kilohertz frequencies (Murphy III, 1985) which is capable of increasing drift. Futterman's relation assumes a linear function of attenuation with frequency up to an arbitrarily high frequency. The intrinsic attenuation for the seismic frequencies calculated strictly from Futterman's relation should, therefore, be lower in order to leave room for the attenuation peak to increase the drift to the observed level. This places the lower limit of attenuation due to intrabed multiples and scattering at 60%. In a study of two wells, Schoenberger and Levin (1978) found that intrabed multiples alone were responsible for 33% to 50% of the observed attenuation. Summarizing the results of 4 additional wells from basins worldwide, they found that the apparent attenuation due to intrabed multiples accounted for 14% to 77% of the observed attenuation. Kan *et al.* (1982) found a negligible contribution to attenuation due to intrabed multiples in an analysis of 2 wells.

The standard method by which the contribution of intrabed multiples to attenuation is estimated involves generating synthetic pulses from a horizontally layered model that do not include the effects of intrinsic attenuation. The pulses are then analysed via the spectral ratio method in the same fashion as in the case of field data. The effect of scattering is therefore ignored, unlike in our approach, and the resulting attenuation due to intrabed multiples alone is assumed to be the same for the field data, which in general may or may not be the case.

Conclusion

The variation of intrinsic attenuation with depth plays an important role in the distortion of seismic wavetrains. Since seismic energy is a convenient probe for studying the structure of the earth, knowledge of the attenuation profile increases the power of the seismic method of geophysical exploration. The attenuation profile is diagnostic in its own right as it is capable of discerning zones of varying fluid saturation, lithology, rock microstructure, porosity, and temperature. The quality of synthetic seismograms would be improved with the use of attenuation profiles in their generation. Seismic inversion would, likewise, be improved, as would the monitoring of steam injection zones, and possibly, earthquake prediction. Synthetic seismogram generation and seismic inversion both require the Q values themselves while the latter two require the level of fluid saturation obtained from the attenuation profile.

As with any experiment, direct sampling of a parameter usually yields the best estimates for the parameter. Laboratory experiments are useful in determining dependencies on parameters but are unable to study rocks in their undisturbed state. The vertical seismic profiling (VSP) survey is the most direct approach for studying in-situ intrinsic attenuation. VSP data from a borehole on Melville Island, in the Canadian Arctic, was employed in this study and subsequently analysed by the spectral ratio method. 108 depth intervals were studied. The depth intervals ranged in thickness from 200 m to 400 m and spanned more than 5 km down the well.

It is proposed here that the oscillations observed on the log graphs likely result from source to source variations and/or irregular scattering and intrabed multiples arising from contorted interfaces. The combined effects of intrabed multiples and frequency-dependent scattering of seismic energy appears to contribute at least 60% of the observed attenuation when considering the entire well. This value was obtained through an analysis of the observed seismic velocity dispersion employing Futterman's third relation (1962) which states that attenuation is a necessary and sufficient condition for dispersion.

The majority of Q values obtained were reasonable, lying between 5 and 130. The data shows the existence of a highly attenuating zone immediately above 4000 m for which the specific attenuation is 0.083 ($Q \simeq 12$) with a standard deviation of 0.042, in contrast with a regional specific attenuation of 0.045 ($Q \simeq 22$) with a standard deviation of 0.040. A strongly attenuating zone, where the specific attenuation is 0.105 ($Q \simeq 10$) with a standard deviation of 0.058, lies between 1930 and 2320 m and is associated with a sandstone horizon of the Lower Triassic. If the effects of intrabed multiples and scattering could be reduced, the contrast in attenuation between the two low-Q zones and the remainder of the borehole would likely increase. This is indicated because the 60% contribution of apparent attenuation could possibly be provided by the more cyclic velocity structure above and below the attenuating zones.

It was the intention of this work to obtain an attenuation profile for the Sherard Bay F-34 well, to relate the profile to the local lithology, and ultimately, to infer properties of the neighbouring rocks. The data were less than ideal due to the use of two types of sources, to the varying source location, and to the fact that the data were collected on three separate occasions. Nonetheless, excellent correlation of large attenuation with a sandstone layer provided confidence that intrinsic attenuation was in fact being observed and, thus, allowed predictions to be made regarding rock properties for the entire well. Given a higher-quality data set, it is felt that further conclusions could be drawn in addition to those presented here. The existence of the two low-Q zones may be attributed to one or more of the following factors which contribute to high attenuation: the presence of fluids with higher viscosity than those in the remainder of the hole, pores with smaller aspect ratios, more pore space, and a higher degree of saturation ($\simeq$95% saturation).

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