Determination of Poisson’s ratio from pre-critical wide-angle seismic reflection data from the BABEL project

Adam Law,* David B. Snyder and Satish C. Singh
BIRPS, Bullard Laboratories, Madingley Rise, Madingley Road, Cambridge CB3 0EZ, UK

SUMMARY
The pre-critical wide-angle seismic reflection data collected during the BABEL survey using multireceiver spreads contain high-amplitude P-wave and mode-converted S-wave reflected arrivals. The two phases can be partially separated and their signal-to-noise ratio emphasized by exploiting the multifold nature and the increased sensitivity of velocity analysis afforded by these data to produce zero-offset-corrected stacked sections. Using these sections, a method has been developed to derive estimates of Poisson’s ratio by optimizing the cross-correlation of P-wave and S-wave arrivals, and hence to constrain petrological models of the crust sampled by these data. Although the apparently high degree of temporal and spatial resolution of reflected arrivals that are visible within the stacked sections must be reduced to achieve a satisfactory degree of correlation robustness, the results shed some light on crustal petrology, and concur with existing estimates of Poisson’s ratio for the Baltic shield.

Key words: Baltic shield, multichannel seismic data, Poisson’s ratio.

1 INTRODUCTION
Determinations of crustal composition using P-wave velocities alone are limited by the non-uniqueness of the correlation between petrological type and P-wave velocity (Holbrook et al. 1992). This can be partly overcome if S-wave information is also available (Christensen & Fountain 1975), as a more limiting set of petrological types can be constrained using the ratio of P-wave velocity to S-wave velocity ($V_p/V_s$), or equivalently Poisson’s ratio. Previous studies of Poisson’s ratio on a crustal scale have involved the use of high-resolution wide-angle reflection/refraction data (Sandemeier & Wenzel 1990 and references therein; Goodwin & McCarthy 1990; Johnson & Hartman 1991; Gohl & Pedersen 1995), and also P-wave and S-wave normal-incidence reflection data (Lüschen, Nolte & Fuchs 1990; Ward & Warner 1991). Models of Poisson’s ratio are typically determined using P-wave and S-wave velocity models produced with forward modelling methods (Gohl & Pedersen 1995), or by calculating the ratio of P-wave and S-wave traveltimes of prominent correlatable reflected phases (Goodwin & McCarthy 1990). If a traveltimes ratio method is employed, it is advisable to separate the P and S phases as much as possible before correlating the reflections. This can be performed in the f–k domain (Dankbar 1985) or t–p (Tatham & Gooolsbee 1984) domains within individual shot or common-midpoint gathers, or by simply stacking the data using differing velocity models (Ward & Warner 1991).

The pre-critical wide-angle data collected during the BABEL survey (Dahl-Jensen et al. 1995) provide a high-resolution image of both P-wave and S-wave arrivals for sections of the Baltic shield. Survey geometries are such that these data can be common-midpoint (CMP) sorted and stacked to produce zero-offset-corrected stacked sections of pre-critically reflected P-wave and S-wave arrivals (Dahl-Jensen et al. 1995). To utilize the high degree of resolution, a method based on P-wave and S-wave event correlation is employed here to estimate Poisson’s ratio from the zero-offset-corrected stacks of these data. In this paper we describe the development of the method using synthetic data, and go on to discuss its application to stacked panels of the BABEL pre-critical wide-angle data from two land stations: station 602 and station 702.

2 THEORY
In this study, a grid search through the data is used to determine Poisson’s ratio by optimizing event correlation of P-wave and S-wave arrivals over a given range of Poisson’s ratio. The relationship of Poisson’s ratio to the ratio of P-wave and S-wave velocities or traveltimes can be explained by considering a simple zero-order reflector upon...
which are normally incident P and converted S wavelets from the same source. The Poisson's ratio above the reflector (\( \sigma \)) is related to the reflected P and S velocities (\( V_P \) and \( V_S \)) and the P and S traveltimes (\( t_P \) and \( t_S \)) by the equation

\[
\frac{t_S}{t_P} = \frac{V_P}{V_S} = \sqrt{\frac{1 - \sigma}{0.5 - \sigma}}.
\]

By applying this method to NMO-corrected stacked data, the value of Poisson's ratio represents the rms Poisson's ratio for the reflected arrival, assuming stacking velocities are rms velocities. If this rms Poisson's ratio or \( V_P/V_S \) is known, then it is possible to determine the equivalent S arrival time for a given P-wave arrival time. The duration of the S wavelet (\( t_{sw} \)) can be computed from the length of the P wavelet (\( t_{pw} \)) in a similar way by assuming that the effect of \( \sigma \) contrast across the reflecting interface is negligible relative to the resolving power of the method. Thus

\[
\frac{t_{sw}}{t_{pw}} = \frac{V_P}{V_S} = \sqrt{\frac{1 - \sigma}{0.5 - \sigma}},
\]

\[
\frac{t_{sw}}{t_{pw}} = V_P/V_S = \sqrt{\frac{1 - \sigma}{0.5 - \sigma}}.
\]

Using eqs (1)–(3), we have developed an algorithm to predict both the S-wave arrival time (\( t_S \)) and duration (\( t_{sw} \)) at each search grid-point (\( t_p \)) over a range of rms Poisson's ratio, generally from 0.2 to 0.3. Instead of identifying individual P-wave arrivals (\( t_{pw} \)), a P-wave correlation window of fixed duration is used to define the P-wave arrival to be correlated, with its start time at \( t_p \). This is used to predict an equivalent S-wave correlation window. For each prediction, a linear compression is applied to the predicted S-wave correlation window to equate it in length and frequency content to the P-wave correlation window. The amplitudes of the P- and compressed S-wave correlation windows are then equated and normalized to the maximum P-wave correlation window amplitude, and subsequently cross-correlated at zero lag, with correlation normalized to the P-wave correlation window length. For coherent P-wave and S-wave arrivals, a correct prediction of rms Poisson's ratio should result in a maximum in this cross-correlation value, as the P-wave and compressed S-wave correlation windows should be of similar form. In the digital domain, the linear compression is approximated by a resampling at a coarser frequency, proportional to the rms Poisson's ratio. Cubic spline interpolation is therefore used to predict the S-wave function during resampling, with a high-cut filter applied to eliminate wraparound of the higher frequencies.

To apply the grid search to the BABEL pre-critical wide-angle data, two assumptions must be made. Firstly, P to S conversion is assumed to occur at or near to the seabed sediment to basement interface. Here, conditions are favourable, as the S-wave velocity of the basement approaches or is greater than the P-wave velocity of the overlying sediment, and the velocity gradient is large over distances of less than half a wavelength of the P-wave data (White & Stephen 1980; Guest & Thompson 1992). Water depths in the survey area were on average 50 m. Thus, the traveltime of the P wave from shot (at 7.5 m depth) to the seabed has a negligible effect on the total traveltime of the mode-converted S-wave arrivals.

It was also assumed that the vertical and lateral errors in the CMP approximation between P and S phases are small relative to the resolving power of the method, and thus that the divergence between P- and S-wave ray paths is minimal, as was assumed when processing the BABEL pre-critical wide-angle data (Dahl-Jensen et al. 1995). Ray tracing (Zelt & Smith 1992) through a 2-D model built from P-wave velocities and Poisson's ratios determined in this study confirmed the very near coincidence of P- and S-wave ray paths. For efficient P to S conversion as defined above, Snell's law dictates that the deviation in angle of incidence between P- and S-wave ray paths is proportional to the variation of Poisson's ratio across the interface. Where this variation is large, the ray-path deviation error is only appreciable for rays incident at 20° or greater. For these data, this error is only significant for phases reflected from the upper crust at depths above 15 km at the maximum survey offsets. As these arrivals are generally suppressed or muted during the stacking process when applying normal moveout corrections, their influence during the linear search is minimal.

\section{Development}

Initially, the algorithm was tested using P-wave and S-wave synthetic seismograms generated from a single-interference velocity model with a Poisson's ratio above the interface of 0.265. To simulate the BABEL pre-critical wide-angle data, the seismograms were generated using a long, multicycle wavelet with a similar bandwidth. The results were examined as surfaces of normalized correlation strength as a function of Poisson's ratio and traveltime.

The BABEL pre-critical wide-angle data, in common with other deep seismic data, have a long, complex downgoing wavetrain exhibiting a high degree of cyclicity due to their limited bandwidth, low dominant frequency and interference with multiple energy (Juhlin 1990; Dahl-Jensen, Dyrelius & Palm 1991). For a given P-wave arrival, this pronounced cyclicity gives many spurious correlation maxima as the predicted S-wave arrival time (\( t_S \)), and hence the S-wave correlation window (\( t_{sw} \)) move over successive cycles of the S-wave arrival with increasing Poisson's ratio during the grid search. This effect is known as cycle-skip. The low (approximately 10 Hz) dominant frequency of the data makes the change in the bandwidth of the S-wave correlation window (\( t_{sw} \)) during linear compression generally small over the range of Poisson's ratio used to guide the search, so that all correlation maxima will be nearly equivalent in magnitude. As the P-wave arrival is also multicyclic, each time-step will result in this multiple correlation series being repeated.

\subsection{Reduction of cycle-skip}

Uniqueness of correlation, and thus the severity of cycle-skipping problems, is related to the waveform within the P-wave and S-wave correlation windows. For a P-wave correlation window of 100 ms within the P-wave data (\( P-TWT \)), correlation is affected by cycle-skip even for
A. Law, D. B. Snyder and S. C. Singh

Figure 1. Correlation surfaces from results of grid search for synthetic data, showing variation of normalized P-wave to S-wave correlation with Poisson’s ratio and P−TWT. Grey-scale is normalized correlation. (a) P-wave correlation window length = 100 ms P−TWT. (b) P-wave correlation window length = 600 ms P−TWT. (c) and (d) As (a) and (b), but correlating perigram (Shteivelman et al. 1986) of data. Note the single maximum over the Poisson’s ratio search range when the perigram function of the data is used.

synthetic data with a nearly infinite signal-to-noise ratio (Fig. 1a), because the wavelet being correlated is poorly defined. Correlation is therefore more susceptible to short-period noise. Enlarging this to 600 ms P−TWT (Fig. 1b) gives a multicyclic, and therefore better-defined, P-wave arrival, at the expense of vertical resolution of the reflection event, which produces elongated correlation maxima. Although the correlation is strongest at a Poisson’s ratio of around 0.265, there are still peaks of similar magnitude at Poisson’s ratios of approximately 0.255 and 0.275, indicating that cycle-skipping during correlation is still prevalent.

Correlation uniqueness can also be achieved by reducing cyclicity within the data themselves, with an obvious loss in vertical resolution. Other workers (e.g. Ward & Warner 1991) achieve this by correlating the envelope function of the data, but here the perigram function of the data is used. A perigram (Shteivelman, Landa & Gelchinsky 1986) has the effect of moving the amplitude spectrum towards lower frequencies, and thus reducing periodicity, without the introduction of a large base shift, which often perturbs correlation. Shteivelman et al. (1986) derive the perigram of a seismic trace \( g(t) \) from its amplitude function \( a(t) \) by subtracting the low-frequency component of the envelope function \( b(t) \), which is defined as the running average of \( a(t) \) over a given time window \( \Delta T \); thus

\[
g(t) = a(t) - b(t)
\]

where

\[
b(t) = \frac{1}{\Delta T} \int_{t-\Delta T/2}^{t+\Delta T/2} a(\tau) d\tau.
\]

Thus, where \( \Delta T \) is large, \( b(t) \) approaches the constant, or base value, component of \( a(t) \).

Performing the same correlations as in Figs 1(a) and (b) on the perigram of the traces shows a marked increase in correlation robustness, albeit at the expense of resolution of Poisson’s ratio due to the lower frequency content of the data. For a P-wave correlation window duration of 100 ms P−TWT, correlation aliasing is considerably reduced (Fig. 1c). However, the P-wave correlation window is still non-unique, resulting in two correlation maxima at Poisson’s ratios of 0.20 and 0.26. Enlarging the P-wave window length to 600 ms P−TWT results in a more unique arrival for correlation at the expense of vertical resolution, giving one global maximum at a Poisson’s ratio of 0.26 ± 0.01 for
normalized correlation with magnitudes greater than 0.8 (Fig. 1d).

The use of a correlation window of finite duration imposes resolution limits both in the TWT of a reflection giving strong correlation, and in the ability of the method to resolve between individual reflections. For an isolated reflection, the error in the TWT position of its associated correlation maxima is roughly equal to half the duration of the P-wave correlation window used. Therefore, the minimum temporal separation between two reflected events will also be equal to half the correlation window length. Variations in Poisson’s ratio over temporal periods less than this will not be successfully resolved.

3.2 Uncertainty in Poisson’s ratio

Our analysis employs the ratio of zero-offset traveltimes estimated using stacking velocities that were determined from maximizing the coherency of reflections on stacks, and is therefore subject to the usual errors due to approximation of rms velocities by stacking velocities and the non-physical basis of stacking velocities (Al-Chalabi 1974). Similarly, interval velocities derived from the P- and S-wave traveltimes are subject to local errors within each layer that are typical of interval velocities (Al-Chalabi 1974; Landa et al. 1991), but those used here were additionally constrained with velocities from coincident refraction surveys and nearby well logs (BABEL Working Group 1993).

Specifically, errors in the estimated Poisson’s ratio for an individual reflector arise from two related approximations in the method: uncertainties in the zero-offset traveltime due to the incomplete moveout hyperbola available, and uncertainties in correlating the traveltimes of individual P- and S-wave reflections. Correlation uncertainty arises from the limited signal-to-noise ratio and bandwidth of the data. Even after deconvolution, signal contamination by spurious noise within the BABEL pre-critical wide-angle data produces variable normalized correlation values for equivalent P-wave and S-wave arrivals below some threshold value. This value of normalized correlation is here called the correlation cut-off. Uncertainty in the Poisson’s ratio will therefore result from this correlation cut-off defining a finite width to the correlation maxima in Poisson’s ratio and traveltime space. This width will also be affected by the bandwidth of the data, with resolution decreasing in proportion to the dominant frequency.

Correlation uncertainty in Poisson’s ratio (\( \delta \sigma_n \)) can be determined (Landa et al. 1991) as

\[
\delta \sigma_n = \sqrt{\frac{2\rho}{(\delta E/\delta \sigma_n)^2}},
\]

where \( \rho \) is the correlation cut-off value, \( E \) is the correlation function and \( \sigma_n \) is Poisson’s ratio for the \( n \)th layer. This uncertainty is estimated directly from Poisson’s ratio versus traveltime correlation surfaces using the width of the correlation contour in Poisson’s ratio for a particular correlation cut-off. Using synthetic data with bandwidths comparable to those of the perigam of the pre-critical wide-angle data from BABEL stations 602 and 702 (e.g. Fig. 1d), the minimum value of uncertainty can be estimated (Table 1). For example, if all correlation above a normalized correlation cut-off of 0.7 is assumed to result from the correlation of coherent P-wave and S-wave arrivals, then uncertainty in the estimated Poisson’s ratio is around \( \pm 0.02 \).

A further error in Poisson’s ratio is imposed by the uncertainty in the zero-offset-corrected traveltime for both P-wave and S-wave arrivals after stacking. Large offsets violate the basic assumptions of normal-incidence stacking techniques, but these assumptions were deemed valid for offsets less than the target reflector depth (Dahl-Jensen et al. 1995). Better sensitivity in optimizing stacking velocities is counterbalanced by decreased resolution in the zero-offset traveltime due to the lack of near-zero offsets and the limited differential moveout available. The uncertainty in arrival time at zero offset \((\Delta t_0)\) of a reflection with traveltime \( t_0 \) at offset \( X \) that is primarily related to the uncertainty in stacking velocity \((\Delta V_{st})\) can be estimated using the equation

\[
(t_0 \pm \Delta t_0)^2 = (t_0 \pm \Delta t_0)^2 - \left(\frac{X}{V_{st} \pm \Delta V_{st}}\right)^2.
\]

Although insignificant for near-normal-incidence reflection data, the uncertainty in zero-offset-corrected traveltime of reflections \((\Delta t_0)\) becomes large with increasing minimum shot–receiver offset with common-midpoint gathers of the BABEL data. Distortions of the moveout hyperbola that degrade \( V_{st} \) determinations and uncertainty inherent in normal moveout corrections \((\Delta t_0)\) at non-zero offset (Fig. 2) are mapped into uncertainty of \( t_0 \) in these data. At shallow depths and far offsets this uncertainty dominates the error in Poisson’s ratio estimates. At other points, the correlation cut-off described above exceeds and independently encompasses the uncertainty in P- and S-wave traveltimes (Table 1).

4 THE BABEL PRE-CRITICAL WIDE-ANGLE DATA

The BABEL survey was designed to study the Precambrian rocks of the Baltic shield and their transition to Phanerozoic crust using deep seismic reflection and refraction methods.
Figure 2. Estimated error in the normal moveout correction ($\Delta_{t}$) as a function of minimum source to receiver offset ($x_i$) and zero-offset two-way traveltimes ($t_0$) of 5, 10 and 15 s. The correction was calculated for reflections using the stacking velocities for BABEL station 602.

(BABEL Working Group 1990). As part of the data acquisition, land-sited multichannel geophone spreads were used to record the airgun shots for four of the near-vertical reflection lines (Fig. 3). Each spread was composed of 35 groups of six 10 Hz vertical-component geophones, and an associated 2 Hz three-component geophone. Data collected in this form with a static receiver spread makes it possible to record reflected and refracted arrivals from the crust at increasingly wider angles of incidence while maintaining a dense trace spacing and the potential for multifold coverage. Because the maximum source to receiver offset of these data is less than 80 km, most arrivals are pre-critical reflections.

These data can be sorted into CMP gathers and stacked to provide a zero-offset-corrected section (Dahl-Jensen et al. 1995). Although the differential offset within each CMP gather remains roughly constant, the geometry of a fixed receiver spread and moving shots results in the minimum offset within each CMP gather increasing monotonically along the chosen processing line. The hyperbolic approximation of normal moveout used by most velocity analysis packages (Taner & Koehler 1969) implies a corresponding increase in stacking velocity sensitivity for a given reflected arrival with increasing minimum offset if the fold remains constant. In practice, this results in the resolution of stacking velocity being as good as $\pm 200$ m s$^{-1}$ for arrivals with zero-offset traveltimes of 10 to 15 s TWT at offsets where the normal moveout approximation is still valid (McBride et al. 1993; Dahl-Jensen et al. 1995), although with decreasing fold and signal-to-noise ratio this may drop to $\pm 400$ m s$^{-1}$ (Law 1993).

Using this stacking velocity sensitivity, the pre-critical wide-angle data from BABEL land stations 602 and 702 have been CMP-sorted and -stacked to produce separate stacked panels of coherent P-wave and S-wave reflected arrivals. The shot-ordered data are first corrected for spherical divergence, deconvolved to suppress strong water-bottom reverberation common to most of the data from the BABEL survey, bandpass filtered using filters designed from the bandwidths of the P-wave and S-wave phases, and sorted into the CMP domain. Separate stacked panels of the P-wave and S-wave reflected arrivals can then be produced with independent stacking velocity models. During normal moveout correction, the increasing initial offset within the CMP gathers implies that NMO stretch may adversely distort the data at small traveltimes. Thus, slanted muting functions were applied to the CMP-gathered data before NMO correction, with the maximum TWT of the muting function increasing as a function of offset.

4.1 Station 602

Here, survey geometry was such that the final fold of coverage of the CMP-sorted data was much greater than that of the other multichannel stations. Data were also collected over a much larger range of offsets (6 to 80 km). The resulting stacked sections (Figs 4a and b) therefore cover a greater distance and angle of incidence and also have a higher signal-to-noise ratio than the corresponding sections from station 702 (Figs 5a and b). There are several similarities between the P- and S-wave stacks from station 602 (Fig. 4), including the upper-crustal refracted and reflected arrivals (A), several northward-dipping reflections (B), an increase in reflectivity to the south, and coherent arrivals from the P-wave and possibly S-wave reflection Moho (C and ?C).

4.2 Station 702

The data from BABEL station 702 have a lower average fold (approximately 15) than station 602 (approximately 25) when CMP-sorted using equivalent processing strategies. They also have a lower signal-to-noise ratio for individual traces as the geophone spread was laid out over very variable near-surface geology. Thus signal-to-noise ratio varies considerably from geophone group to geophone group across the spread. Data were also recorded over a much more limited range of offsets (27 to 72 km). Most importantly, the amplitude of recorded S waves was, on the whole, less than that at station 602. The P- and S-wave stacks for 702 are therefore more noisy than those of station...
Determination of Poisson's ratio from BABEL project data

602, with S-arrivals only stacking coherently above 10 s TWT (Figs 5a and b). These data are therefore used here to test the sensitivity of the grid search to noise.

5 APPLICATION OF THE GRID SEARCH TO THE BABEL PRE-CRITICAL WIDE-ANGLE DATA

The grid search algorithm described in Section 3 was applied to the perigram function of stacked traces from BABEL stations 602 and 702. The data were partially summed over 10 traces to enhance persistent reflections, and correlated with a P-wave correlation window duration of 1400 ms P-TWT, giving a minimum TWT resolution of ±700 ms. This is the minimum window that resulted in repeatability when applying the algorithm to P-wave and S-wave perigram pairs from the stacked sections of both station 602 and 702. For each perigram pair, the algorithm was programmed to search from 0 to 16 s P-TWT with a search grid interval of 8 ms, the sampling interval of the data. At each grid-point, the maximum P-wave to S-wave correlation and associated Poisson's ratio value between 0.2 and 0.3 were recorded, rather than the range of cross-correlation values (cf. Fig. 6), as correlation was found to be sufficiently robust if the perigram function of the data was used. At the end of each trace, the correlation magnitude was normalized to its maximum value.

The results from this form of grid search can be used to produce two surfaces: one showing the magnitude of each normalized correlation maximum and the other the rms Poisson's ratio, both as a function of CMP or trace number and two-way traveltime position in the P-wave stack (P-TWT). The algorithm returns a value of Poisson's ratio at each search grid-point, irrespective of whether it is associated with strong correlation of coherent P-wave and S-wave events. The results from the grid search are therefore analysed by first studying a surface of normalized correlation magnitude to determine where coherent arrivals have been correlated.
Figure 4. Zero-offset-corrected stacks for station 602 (after Dahl-Jensen et al. 1995). Note that the vertical scale for the S-wave stack (b) has been compressed by a factor of $1/\sqrt{3}$ for visual comparison of P-wave and S-wave reflected phases. Several visual correlations can be made between the P-wave (a) and S-wave (b) sections. The more important of these are the strong upper-crustal arrivals (A—where not destroyed by NMO stretch), a north-dipping series of reflections (B), an increase in reflectivity within the southern third of both panels, and strong P-wave and possible S-wave arrivals from the reflection Moho (C).

5.1 Correlation of P-wave and S-wave arrivals

The most noticeable feature of these surfaces of normalized correlation magnitude is the variability of correlation maxima with CMP number (Fig. 6). Despite this, several major phases can be identified, including strong upper-crustal correlation for both stations at CMPs where NMO stretch has not destroyed arrivals, for instance from 0 to approximately 3.5 s $P$–$TWT$ for station 602 (A in Figs 4a and b). The correlation surface from the station 602 zero-offset stacks also shows strong correlation of a mid-crustal dipping event (8 s $P$–$TWT$ at CMP 600 to 12 s $P$–$TWT$ at CMP 200: B in Figs 4a and b), and some correlation of lower-crustal events at the southern end of the profile (CMP 680–700). The increase in reflectivity within the southern third of the stacked sections of the station 602 data has resulted in a corresponding increase in the density of correlation maxima. However, there is little strong
Figure 5. Zero-offset-corrected stacks for station 702 (after Law 1993). Display parameters are as in Fig. 4. Note that the signal-to-noise ratio is lower than in the corresponding data from station 602, and little S-wave energy appears to stack coherently below 10 s TWT.
Figure 6. Correlation surfaces from results of grid search for partially summed $P$- and $S$-wave perigrams. Blank areas indicate zones of data excluded in the subsequent interpolation of the data. (a) BABEL station 602. The main visual correlations between Figs 4(a) and (b) appear here, including upper-crustal correlation (A) and the north-dipping reflectors (B). The approximate position of the reflection Moho is also indicated (C), although no strong correlation has occurred at these traveltimes. Note also that the increase in reflectivity in the southern third of both stacks is mirrored by an increase in correlation maxima density within the correlation surface. (b) BABEL station 702. Note the lack of strong correlation below 8 s $P$-TWT.
correlation where no visual correlation of P-wave and S-wave arrivals can be made. For instance, few correlation maxima with magnitudes greater than 0.5 exist below around 8s P–TWT within the correlation surface of the station 702 stacks (Fig. 6b).

The lack of lateral coherency of correlation maxima reflects the lack of lateral coherency and amplitude within the S-wave stack relative to the P-wave stack. This could be due to a number of factors, including variation of CMP fold or problems in the optimization of S-wave statics, but is most likely to be due to variation in P to S conversion efficiency, as sea-bottom geology is laterally variable here (Winterhalter et al. 1975). White & Stephen (1980) have attempted to quantify the transfer of energy from P wave to S wave at interfaces in oceanic environments, and their conclusions imply that in this case conditions at the seabed of the Baltic may result in good mode conversion. Any lateral variability in seabed sediment type and thickness will result in large variations in Poisson’s ratio across the sediment to basement interface, which will affect the amplitude of the vertically polarized mode-converted S wave (Guest & Thompson 1992), the dominant component recorded here.

5.2 Determination of Poisson’s ratio

Two criteria are used to discriminate rms Poisson’s ratio estimates that have resulted from the correlation of coherent arrivals. Firstly, all correlation maxima and associated Poisson’s ratio values that occur at traveltimes and trace numbers where no energy is visible within the S-wave stacked section are discarded (blank areas in Fig. 6). Secondly, a normalized correlation cut-off is defined, above which correlation maxima are assumed to result from the correlation of ‘coherent’ P-wave and S-wave energy. We refer to these time-, offset- and correlation-dependent definitions of ‘coherent’ correlation as T–X criteria. Values of Poisson’s ratio with associated normalized correlation values that pass these T–X criteria define an irregular grid in P–TWT and CMP number space. This Poisson’s ratio grid can then be interpolated for display. As the stacking velocity functions used to stack the BABEL pre-critical wide-angle data are generally not true rms velocities, due to the significant offsets, estimates of Poisson’s ratio derived from them are described here as apparent rms Poisson’s ratio.

Here, we use a correlation cut-off of 0.7 for both BABEL pre-critical wide-angle data sets. Before an rms Poisson’s ratio grid is built, the data are smoothed temporally using a running average with a box-car function length equal to the minimum temporal resolution of the algorithm, in this case 700 ms P–TWT, and laterally using a similar filter with a box-car length equal to the diameter of the first Fresnel zone computed for these data using the P-wave stacking velocity models. Smoothing is necessary to suppress spurious variations in correlation magnitude and Poisson’s ratio over lateral and temporal periods less than the minimum resolving power of both the algorithm and the data themselves.

Using a correlation cut-off of 0.7 results in an uncertainty in the rms Poisson’s ratio as a result of the algorithm of ±0.02. For station 602, the error in Poisson’s ratio due to errors in reflection zero-offset-corrected traveltimes only exceeds this uncertainty towards the southern end of the stacked sections where absolute offset within CMPs is large, and at P–TWT less than 5s, resulting in an average error of ±0.02 for Poisson’s ratio estimates from within this data window. For station 702, the rapid decrease in stacking velocity resolution with TWT coupled with the concentration of strong correlation above 8s P–TWT results in a cumulative error in rms Poisson’s ratio estimates of around ±0.03.

The apparent rms Poisson’s ratio surfaces produced in this way for both stations are, as is indicated by the correlation surfaces (Fig. 6), somewhat contaminated by poor correlation. To develop a clearer picture of the bulk variation of Poisson’s ratio, the irregular grid of apparent rms Poisson’s ratio values defined by the T–X criteria described above is first averaged over all CMPs after temporal smoothing. We examine the variation of apparent rms Poisson’s ratio with P–TWT for both stations using this averaged curve, and study lateral variation from this function with reference to the interpolated surface after temporal and lateral smoothing.

5.3 Apparent rms Poisson’s ratio—station 602

Figure 7 shows the averaged apparent rms Poisson’s ratio curve for station 602. The statistical validity of each value of apparent rms Poisson’s ratio (Fig. 7c) is given by the overall averaged normalized correlation (Fig. 7b) above the cut-off of 0.7 and the number of traces involved in the summation, or population (Fig. 7a) at each P–TWT grid point. Some short-period variations in correlation are still present even after smoothing the data. Although strong correlation has been obtained to around 15.0 s P–TWT, values of apparent rms Poisson’s ratio below around 14.2 s P–TWT are constrained by few traces (Fig. 7a).

Relatively high values of apparent rms Poisson’s ratio have been estimated from upper-crustal arrivals (1 to 2 s P–TWT) of around 0.26 ±0.02 to 0.28 ±0.02, where reflection zero-offset-corrected traveltime errors dominate, decreasing to 0.24 ±0.02 at around 4 s P–TWT. From 4.5 s to 7.5 s P–TWT there is a similar distribution of apparent rms Poisson’s ratio, with a slow decrease from 0.26 ±0.02 to 0.23 ±0.02. The apparent rms Poisson’s ratio curve then gradually increases to 0.26 ±0.02 at 14.2 s P–TWT.

The smoothed surface of apparent rms Poisson’s ratio for station 602 (Fig. 8) shows a similar variation as a function of P–TWT, with high values within the first few seconds at near CMPs. Some lateral variation is visible at mid-crustal traveltimes, with a background value of around 0.24 ±0.02 perturbed by an isolated area with apparent rms Poisson’s ratio of around 0.27 ±0.02, although this is poorly constrained by correlation maxima. This background value increases to around 0.26 ±0.02 towards the Moho (15 s P–TWT).

5.4 Apparent rms Poisson’s ratio—station 702

The lack of strong correlation below around 8 s P–TWT is the most obvious feature of the averaged apparent rms Poisson’s ratio curve for station 702 (Fig. 9c). Where values of apparent rms Poisson’s ratio have been calculated below this P–TWT, little usefulness can be placed in the results as
A. Law, D. B. Snyder and S. C. Singh

Figure 7. Horizontally averaged apparent rms Poisson’s ratio curve for station 602 as a function of P-TWT, calculated using the T-X criteria discussed in text. (a) Population of traces included in summation. (b) Average correlation magnitude passing correlation cut-off. (c) Apparent rms Poisson’s ratio. Poisson’s ratio has been reset to 0.20 where no information was available to be averaged after T-X constraints were applied.

Few traces have been included in the average (Fig. 9a). Within the first 6 s P-TWT there is some variation of apparent rms Poisson’s ratio, from 0.23 at 2 s P-TWT to around 0.27 at 6.5 s P-TWT. However, the increased errors in the stacking velocity function used to NMO-correct these data, coupled with the concentration of strong correlation at small P-TWT, results in an increase in apparent rms Poisson’s ratio uncertainty to ±0.03. The small correlation

Figure 8. Smoothed apparent rms Poisson’s ratio surface from results of grid search for station 602. Surface is produced by interpolating an irregular grid of Poisson’s ratio values that have associated normalized correlation magnitudes that passed the T-X criteria discussed in text. Grey-scale is now Poisson’s ratio from 0.2 to 0.3. Note how Poisson’s ratio is partially contaminated by short-period anomalously high or low values associated with the spurious correlation that remains after smoothing.
peak at around 8 s P-TWT appears to be fairly well constrained by the population curve (Fig. 9a), giving a value of apparent rms Poisson’s ratio of around 0.26 ± 0.03. The very high values of apparent rms Poisson’s ratio at around 8 s P-TWT are poorly constrained by both correlation and population curves.

Some lateral variation is again seen within the smoothed apparent rms Poisson’s ratio surface for station 702 (Fig.

Figure 9. Averaged apparent rms Poisson’s ratio curve for station 702 as a function of P-TWT, calculated using the \( T-X \) criteria discussed in text. (a) Population of traces included in summation. (b) Average correlation magnitude. (c) Apparent rms Poisson’s ratio. Poisson’s ratio has been reset to 0.20 where no information was available to be averaged after \( T-X \) constraints were applied. Values of Poisson’s ratio below 10 s P-TWT are poorly constrained, as is indicated by the population curve (a).

Figure 10. Smoothed apparent rms Poisson’s ratio surface from results of grid search of station 702. Surface produced by interpolating irregular grid of Poisson’s ratio values with associated normalized correlation maxima magnitudes that passed the \( T-X \) criteria discussed in text. Grey-scale is now Poisson’s ratio from 0.2 to 0.3.
6 INTERVAL POISSON'S RATIO

The method used here to determine Poisson's ratio, in common with many others, relies on the ratio of $P$- and $S$-wave traveltimes for a particular reflection. The resulting Poisson's ratio is therefore a bulk or apparent rms Poisson's ratio for the crust above the reflector. Thus, variations in Poisson's ratio close to the reflector may be masked by the averaging effect of the crust above. If we are interested in a better determination of Poisson's ratio for a particular crustal interval, this averaging effect may need to be removed from the calculated value of Poisson's ratio, particularly for deeper intervals. For wide-angle reflection/refraction surveys (e.g. Assumpção & Bamford 1978), this effect is small and reduces with increasing source-receiver offset, as more and more of the ray path is within the interval of interest. Determining Poisson's ratio with offset for such surveys results in a function that becomes asymptotic to the value of Poisson's ratio for that reflector or interval. However, for data such as the BABEL pre-critical wide-angle data used here, the ray paths are near-vertical, and are corrected to the vertical during the stacking process. Therefore, for deeper targets, the ray path will have a significant component within the crust above. To glean a more realistic picture of the variation of Poisson's ratio with depth, this averaging effect must be progressively removed from the data with increasing $P$-TWT, to determine the interval Poisson's ratio.

The method used to determine the interval Poisson's ratio ($\sigma_{\text{INT}}$) from the results of the grid search involves calculating the differential traveltimes for both $P$ and $S$ phases within a defined interval of interest, using the equations

$$\sigma_{\text{INT}} = \frac{1 - 0.5p}{1 - p}$$

(8)

where

$$p = \left(\frac{t_{s2} - t_{s1}}{t_{p2} - t_{p1}}\right)^2$$

(9)

where $t_{s1}$ is the $S$-wave traveltimes to the top of the interval; $t_{p1}$ is the $S$-wave traveltimes to the base of the interval; $t_{p2}$ is the $P$-wave traveltimes to the top of the interval; $t_{p1}$ is the $P$-wave traveltimes to the base of the interval and is almost identical to the calculation of interval velocities from stacking velocity functions (Dix 1955). As the $P$-wave traveltimes to the top and bottom of the interval are already known, the corresponding $S$-wave traveltimes can be determined from the value of apparent rms Poisson's ratio for the crust for each of the $P$-wave traveltimes. The interval $P$- and $S$-wave traveltimes can then be calculated, and the interval Poisson's ratio determined from the ratio of these traveltimes.

Theoretically, it should be possible to determine an interval Poisson's ratio directly from rms Poisson's ratio by calculating the difference between two successive search grid-points. In practice, errors in the apparent rms Poisson's ratio determined from the zero-offset-corrected stacks of the BABEL pre-critical wide-angle data can lead to strong gradient changes over such short $P$-TWT intervals. Thus, the error in the corresponding $S$-wave traveltimes will dominate the calculation, and the resulting values of the interval Poisson's ratio may vary considerably over short lateral and temporal periods.

To minimize this error, the interval Poisson's ratio is determined over the traveltimes between strong correlation maxima as defined by the $T$-X criteria discussed above, with the top and the base of each interval fixed at the centre of its defining correlation maxima. Where one correlation maximum has been found but a second cannot be identified within the normalized correlation information for a perigram pair, the method used here searches laterally within the data at traveltimes below the correlation maximum. Values of apparent rms Poisson's ratio for each maximum are calculated using a box-car averaging function, as the maxima are elongate in $P$-TWT. The calculation of interval Poisson's ratio is constrained by placing a minimum limit on the size of each correlation maximum in $P$-TWT, to reduce the inclusion of short-period correlation spikes within the calculation, and by constraining the maximum gradient change between the two maxima to within the Poisson's ratio range used by the algorithm.

A further error in Poisson's ratio is due to the error in the zero-offset-corrected traveltime to both $P$- and $S$-wave arrivals. Because uncertainty in the interval Poisson's ratio in a given layer depends only on uncertainties in the velocities from the top and the bottom of the layer, no error accumulates with time (Landa et al. 1991).

Here, we study the interval Poisson's ratio in an identical manner to apparent rms Poisson's ratio, by first describing the bulk variation with $P$-TWT using an averaged interval Poisson's ratio curve, and then examining any lateral variation using a smoothed surface. For station 602, the apparent rms (Fig. 11a) and interval (Fig. 11b) Poisson's ratio curves show a considerable degree of similarity at upper- to mid-crustal traveltimes. However, the effect of crustal averaging on apparent rms Poisson's ratio becomes significant with increasing $P$-TWT, supressing the Poisson's ratio for lower-crustal arrivals, and resulting in a difference between the rms and interval Poisson's ratio curves of about 0.02 over this time interval (12.5 s to 14.5 s $P$-TWT). For station 702, averaged apparent rms (Fig. 12a) and interval (Fig. 12b) Poisson's ratio curves show similar values within the first few seconds $P$-TWT, where strong correlation occurs. However, the interval Poisson's ratio curve shows a more uniform gradient, from 0.24 ± 0.03 at 2 s $P$-TWT to 0.28 ± 0.03 at 6 s $P$-TWT, and diverges from the apparent rms Poisson's ratio curve with increasing $P$-TWT. As with the apparent rms Poisson's ratio curve, values below 8 s $P$-TWT and above 2 s $P$-TWT are poorly constrained.

Lateral variation in the interval Poisson's ratio is difficult
Determination of Poisson's ratio from BABEL project data

Figure 11. Apparent rms (a) and interval (b) Poisson's ratio averaged curves and (c) smoothed interval Poisson's ratio surface for BABEL station 602 stacks. Note how the crustal averaging of the rms Poisson's ratio values minimizes this curve relative to the interval Poisson's ratio curve with increasing P-TWT, as discussed in the text.

to determine from these data, due to the perturbation of the calculation of the interval Poisson's ratio by the effects of spurious correlation. However, the interval Poisson's ratio surface for station 602 (Fig. 11c) shows values of interval Poisson's ratio at upper-crustal traveltimes (1 to 3 s P-TWT) of 0.28 ± 0.02, decreasing slowly at mid-crustal traveltimes (∼10 s P-TWT) to around 0.24 ± 0.018, with the isolated area of higher Poisson's ratio seen in the equivalent apparent rms Poisson's ratio surface (Fig. 8) giving an interval Poisson's ratio of 0.28 ± 0.018. The interval Poisson's ratio then increases from this background value of 0.24 ± 0.018 to values of 0.28 ± 0.018 towards the Moho at 15 s P-TWT, where correlation is strong, with isolated areas of around 0.24 ± 0.018.

Figure 12. Apparent rms (a) and interval (b) Poisson's ratio averaged curves and (c) smoothed interval Poisson's ratio surface for BABEL station 702 stacks. Again the rms Poisson's ratio curve is minimized relative to the interval Poisson's ratio curve with increasing P-TWT.
There is considerable lateral and temporal variation within the interval Poisson’s ratio surface for station 702 (Fig. 12c), even after lateral smoothing, reflecting the increase in the effect of spurious correlation in the equivalent apparent rms Poisson’s ratio surface (Fig. 10), although the broad gradient shown by the averaged curves for station 702 is still present within the interval Poisson’s ratio surface. There is also a large area with interval Poisson’s ratios of 0.28 ± 0.03 at around 5 s P-TWT, extending approximately from CMP 150 to CMP 300, which has led to the high values at these P-TWTs when averaged (Fig. 12b). However, this appears to be an artefact of the calculation of the interval Poisson’s ratio, as it is not constrained by any strong correlation. Below 6 s P-TWT, the surfaces are poorly constrained as there is little coherent correlation, which has resulted in the large zones of interval Poisson’s ratio below 0.20.

7 POISSON’S RATIO OF THE EASTERN SWEDISH SHIELD

The averaged interval Poisson’s ratio curves (Fig. 11b and Fig. 12b) are used here to study the bulk variation of Poisson’s ratio for the crust sampled by stations 602 and 702. Any lateral variation can be sought using the interval Poisson’s ratio surfaces (Fig. 11c and Fig. 12c). Values of Poisson’s ratio for igneous and metamorphic rocks from various global locations compiled by Chrostosn & Brooks (1989), Chrostosn & Simmons (1989), Goodwin & McCarthy (1990) and Holbrook, Mooney & Christensen (1992) are used to interpret these results in the context of crustal petrology, supplemented by values of Poisson’s ratio calculated from laboratory measurements of the P- and S-wave velocities for samples of varying petrological type taken from the vicinity of station 702 (Table 2).

7.1 Station 602

In the locality of station 602, the top 10 km of upper crust is thought to be composed primarily of metasupracrustals and granitoid gneiss intruded by basic igneous bodies (Gorbatschev, Solyom & Johansson 1979) that are interpreted as being thick enough (100 to 300 m) to be seen on nearby deep-seismic reflection profiles (BABEL Working Group 1993). Such a petrological model would result in alternating Poisson’s ratios from 0.28–0.30 (basic igneous material) to 0.22–0.24 (metapelite and granitoid). Within errors, the interval Poisson’s ratio curve for station 602 (Fig. 11b) shows a broad variation of Poisson’s ratio from 0.28 ± 0.02 to 0.245 ± 0.02 from 0 s to around 5.5 s P-TWT. This appears to be in general agreement with the petrological model.

There is a gradual increase in Poisson’s ratio from 5.5 to 12 s P-TWT (Fig. 11b), indicating a crust composed of granitoid (σ = 0.24) to granodioritic (σ = 0.255 to 0.26) rocks. The smoothed interval Poisson’s ratio surface (Fig. 11c) shows a background interval Poisson’s ratio of 0.24 ± 0.018. One isolated area of 0.28 ± 0.018 may result from a local, more basic body within this granitoid crust. At traveltimes greater than 12 s P-TWT, there is an increase in interval Poisson’s ratio with P-TWT to values around 0.28 ± 0.018, indicating an increase in the percentage of basic igneous material towards the Moho. No values of interval Poisson’s ratio have been determined for the Moho to upper-mantle transition itself, due to the lack of coherent S-wave arrivals from the Moho over the survey offsets.

7.2 Station 702

The results from station 702 (Fig. 12b) show a gradual increase in interval Poisson’s ratio from 0.23 ± 0.03 at 2 s P-TWT to almost 0.28 ± 0.03 at around 5 s P-TWT, although values greater than 5 s P-TWT may be an artefact of the interval Poisson’s ratio calculation. Where constrained by strong correlation, this increase may imply a broad variation from rather acidic granitoid rock within the upper crust to rocks of more basic composition with increasing traveltime. A possible decrease in Poisson’s ratio to around 0.26 ± 0.03 is indicated by isolated strong correlation at around 8 s P-TWT, suggesting a granodioritic composition. Very little lateral variation can be determined from the interval Poisson’s ratio surface (Fig. 12c), due to the effect of spurious correlation on the calculation of the interval Poisson’s ratio.

7.3 Geological discussion

These results compare well to earlier estimates of Poisson’s ratio for Precambrian shield. Tarkov et al. (1981), Grad & Luosto (1987), Luosto et al. (1990) and Gohl & Pedersen (1995) model values of Poisson’s ratio from 0.24 within the upper crust to 0.27 or 0.28 within the mid to lower crust. They also model a slow increase in Poisson’s ratio through the mid to lower crust to the Moho, indicating a probable increase in the percentage of mafic material with depth. The results from station 602 indicate a broadly similar picture for the mid to lower crust.

However, these values of Poisson’s ratio differ markedly from those determined for Phanerozoic crust and active continental crust, which indicate a higher value of interval Poisson’s ratio for the lower crust, from 0.27–0.28 (Holbrook et al. 1988) to 0.31 (Ward & Warner 1991), perhaps indicating a higher proportion of mafic material within Phanerozoic lower crust than within Proterozoic lower crust. The gradient of Poisson’s ratio with depth reported by these studies also appears to be different, with a marked increase in Poisson’s ratio from mid to lower crust, with mid-crustal values being, on the whole, lower than those of Precambrian shield. This latter difference may.

Table 2. Poisson’s ratio for rock samples taken in the vicinity of BABEL station 702. The Poisson’s ratio was calculated using the ratio of the average of P-wave and S-wave velocities measured at a confining pressure of 0.3 GPa for three orthogonal cores taken from each sample.

<table>
<thead>
<tr>
<th>Description</th>
<th>σ</th>
<th>Error</th>
</tr>
</thead>
<tbody>
<tr>
<td>Granitoid - Gneissose Fabric</td>
<td>0.251</td>
<td>0.0043</td>
</tr>
<tr>
<td>Mylonitized Granitoid</td>
<td>0.255</td>
<td>0.0082</td>
</tr>
<tr>
<td>Mylonite: Granitoid &amp; Amphibolite Screens</td>
<td>0.258</td>
<td>0.0080</td>
</tr>
<tr>
<td>Strained Amphibolite Dyke</td>
<td>0.267</td>
<td>0.0075</td>
</tr>
<tr>
<td>Regional Unstrained Granitoid</td>
<td>0.241</td>
<td>0.0089</td>
</tr>
<tr>
<td>Hornblendeite</td>
<td>0.247</td>
<td>0.0044</td>
</tr>
<tr>
<td>Mylonitized Metagabbro</td>
<td>0.292</td>
<td>0.0063</td>
</tr>
<tr>
<td>Metagabbro</td>
<td>0.303</td>
<td>0.0057</td>
</tr>
</tbody>
</table>
however, be due to the differences in reflectivity between Phanerozoic and Proterozoic crust. Where it has been sampled by the BABEL survey, the entire crust of the Baltic shield is reflective. There is therefore the potential for recording strong P- and S-wave arrivals throughout the crust, and thus good control on the variation of Poisson’s ratio with depth. However, large sections of Phanerozoic upper and mid crust are unreflective on deep-seismic reflection and refraction profiles (e.g. Barazangi & Brown 1986a, b), leading to intervals where Poisson’s ratio must be extrapolated due to lack of control.

8 DISCUSSION AND CONCLUSIONS

Here, a grid search has been developed to derive estimates of Poisson’s ratio from the high-amplitude P-wave and S-wave arrivals of the BABEL pre-critical wide-angle data. The main challenge encountered throughout the development and application of this method was achieving correlation uniqueness whilst maintaining resolving power. For data such as the BABEL pre-critical wide-angle data, trace by trace correlation of P-wave and S-wave arrivals can become dominated by cycle-skip, as the data exhibit a high degree of cyclicity due to their limited bandwidth, low dominant frequency, and to the presence of residual short-period multiple energy.

Correlation uniqueness is achieved here by reducing data periodicity with the perigram (Shteivelman et al. 1986) function of the data, and by correlating this with a long P-wave correlation window. This reduces the travelt ime resolution of reflected arrivals and therefore Poisson’s ratio. Lateral resolution is also reduced by partially summing the data to accentuate laterally persistent arrivals and suppress incoherent noise that may perturb estimates of Poisson’s ratio. Although both summation and reduction of vertical periodicity result in robust correlation, there is still some contamination of the results of the grid search by short-period noise. The calculation of interval Poisson’s ratio from these results may become unstable in the presence of such short-period variations.

The Poisson’s ratio determined using this method is subject to errors imposed by the finite signal-to-noise ratio of the data, plus, in this case, errors in the zero-offset-corrected traveltimes of correlatable P-wave and S-wave arrivals due to uncertainties in the stacking velocity function used to NMO-correct the data. The latter error may dominate the uncertainty in Poisson’s ratio unless careful attention is paid to stacking velocity errors during data processing. Although the dominant error in values of Poisson’s ratio calculated using the station 602 data is imposed by the method of estimation, the poorer velocity control achieved when stacking the data from station 702 has resulted in an increase in Poisson’s ratio uncertainty to ±0.03.

Despite these errors, final calculated interval Poisson’s ratios still yield valid results, which concur with regional Poisson’s ratio models, and are consistent with interpretations of the geology near to one of the BABEL stations used. Although much of the potential resolution of the BABEL pre-critical wide-angle data was sacrificed to achieve robustness of results, the estimates of Poisson’s ratio still shed light on the composition of the crust sampled by these data, and show some lateral resolution of Poisson’s ratio over the small range of offsets of the data where signal-to-noise ratio permits.

ACKNOWLEDGMENTS

The authors would like to thank everyone at the Section for Solid Earth Geophysics, University of Uppsala, Sweden who were involved in the acquisition of the BABEL pre-critical wide-angle data used here. We would also like to thank the British Antarctic Survey Geophysics Department for the generous loan of time on their Micromax processing computer during data demultiplexing, and also Prof. E.R. Flüh of Geomar, Kiel, for his constructive advice during the preparation of this paper. University of Cambridge contribution 4122.

REFERENCES


