AMPLITUDE-OFFSET RELATIONSHIPS OVER
SHALLOW VELOCITY INVERSIONS

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ABSTRACT

The amplitude and phase of reflected seismic data can vary significantly with the offset distance between source and receiver. Analysis of amplitude-offset relationships in high-resolution seismic data from areas of shallow permafrost in the Beaufort Sea is undertaken in this paper. Reflection attributes produced by energy partitioning at interfaces under non-normal angles of incidence were studied by means of ray-tracing and the Zoeppritz equations. Observed amplitude and phase variations were found to cause CDP stacking techniques to degrade reflection data quality, particularly where the zone of interest features velocity inversions, or is shallow in relation to receiver lengths. As well, it is demonstrated that in these cases velocity analysis based on coherency measures may result in the selection of incorrect stacking velocities.

INTRODUCTION

In recent years, considerable interest has developed in the amplitude and phase information contained in reflection seismic data collected at non-normal angles of incidence. Much of the original work discussing plane-wave reflection and transmission coefficients as functions of angle of incidence can be found in Muskat and Meres (1940), Koefoed (1955, 1962) and Tooley et al. (1965). More recently, Ostrander (1984) discussed the significance of several factors that affect recorded amplitudes and their dependence on source-receiver offset. This paper concentrates on one of these factors: the dependence on energy partitioning as a function of the angle of incidence at a subsurface interface (i.e. the offset dependence of the reflection and transmission coefficients).

Each trace in a common-depth-point (CDP) gather contains amplitude and phase information that will depend on the path associated with that particular source-receiver geometry. It is apparent that variation in recorded amplitude and phase with offset will be particularly pronounced where: 1) total receiver aperture is long with respect to the depths of interest, 2) large velocity contrasts occur in or above the depths of interest, 3) large Poisson's ratio contrasts occur at boundaries, and 4) reflectors are dipping.

In this paper we illustrate the variation in amplitude and phase with source-receiver offset by using data from the shallow sedimentary section of the eastern part of the Beaufort Sea (Arctic Canada). The continental shelf of this area forms a broad, flat, shallow coastal plain with the shelf break occurring at the 80-m isobath some 100 km offshore. Sediment accumulation has been dominated by widespread fluvial deltaic progradation during the glacial periods of the Pleistocene, when sea levels were as much as 100 m below present. Thin, discontinuous shelf deposits mark transgressive phases associated with interglacial periods and rising sea levels (O'Connor, 1980).

During glacial cycles, permafrost aggraded to significant depths within prograding delta plain deposits subaerially exposed for prolonged intervals to arctic climatic conditions. To a lesser extent, the shallow section of these ice-bearing deposits were degraded by transgressions during interglacial cycles. As a result, a thick sequence of ice-bearing sediments has accumulated on the shelf during the Pleistocene (Blasco, 1984).

High-velocity frozen sediments occur at depths from 10 m (below the last transgressive sequence) to more than 600 m below seabed (O'Connor, 1981). The spatial distribution of these high-velocity frozen layers varies significantly both laterally and vertically in response to changes in lithology, porewater salinity, and thermal regime. These changes result from the original variability of sedimentary environments found within the delta plain complex, and the degree of reworking caused by transgressions during cycles of sea-level rise. These spatially discontinuous high-velocity layers can have compressional wave velocities more than twice that of

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the unfrozen sediments. Hunter (1984) reports velocities from less than 2000 m/s to 4000 m/s for ice-bearing sediments. Hence, the near-surface position and high-velocity contrast of the ice-bonded sediments provide an example that is well suited to a study of reflection amplitude and phase characteristics with incident angle.

We have calculated models for the amplitude and phase variations observed in the Beaufort Sea data by using a ray-tracing program we have developed that utilizes the well-known Zoeppritz equations for displacement amplitude partitioning at an interface as coded by Young and Braile (1976). We will demonstrate why conventionally processed seismic data from the shallow parts of this area are often of poor quality, and how amplitude-offset variations can be used as a powerful interpretive tool.

**DATA ANALYSIS**

Figure 1 is a section of 24-fold high-resolution seismic data from a permafrost-affected area of the Beaufort Sea. The source used was a 360-cu-in airgun array. The receiver was a 24 trace streamer with a 12.5-m group interval and a near offset of 75 m. The data were processed at The University of Calgary. The processes were: front-end mute, gather, velocity analysis, normal-move-out (NMO) correction, stacking and automatic gain control scaling.

The main event of interest in Figure 1 occurs at a reflection time of 0.25 s and has limited lateral extent. It is interpreted to be a reflection from the top of ice-bonded sediments. Later in the section, most reflections are laterally discontinuous or exhibit severe static shifts. These shifts result from velocity anomalies caused by the irregular distribution of ice-bonded sediments. Multiple reflections between the sea bottom and the top of the frozen material are also visible, and diffraction hyperbolae are evident whose apices are in line with the ends of the reflection from the top of the ice-bonded sediments. Linear coherent noise, dominant in the lower half of the section, has been discussed previously by Poley and Lawton (1985), who attributed it mainly to off-line reflections from floating sea ice.

Figure 2 is an enlarged view of the area of interest in Figure 1, and shows clearly the limited lateral extent of the reflection interpreted to be caused by the ice-bonded sediments. The reflection from the top of these sediments is rather discontinuous. In comparison, Figures 3 and 4 show respectively the near- and far-trace, unmuted, singlefold displays of the same data. The near-trace data (Fig. 3) show a much more continuous reflection from the top of the ice-bonded sediments than is obtained with either the 24-fold section (Fig. 2) or the far-trace display (Fig. 4). A single event with negative polarity at 3 seconds is particularly evident on the near-trace display (Fig. 3), and is interpreted to be the reflection from the base of the ice-bonded layer. The 24-fold data show little improvement over the singlefold data, and the contradictory appearance in continuity of the reflection from the top of the frozen sediments indicates that a reliable interpretation cannot be made from the fullfold stacked section alone.

In order to analyse the differences between the displays in Figures 3 and 4 further, individual shot records from the seismic lines were examined. Figures 5 and 6 show 24-trace field records obtained over areas interpreted to be non-ice-bonded and ice-bonded, respectively. The positions at which these records were taken are shown on Figure 2. The major difference between the two records is the events on Figure 5, which are marked with arrows. Study of the shot and CDP records, the various fold stacks, and the velocity analysis panels suggests that these are reflections from the top and bottom of an ice-bonded layer. The top reflection varies in amplitude and phase across the shot record, while the bottom event is of opposite polarity and indiscernible beyond about trace 10.

**MODELLING**

In order to examine the observed amplitude and phase variations with source-receiver offset, numerical modelling was undertaken. Elastic theory and the Zoeppritz equations (Koefoed, 1962; Young and Braile, 1976), were used in a raytracing program to determine if these variations can be explained by partitioning of energy at interfaces under conditions of non-normal angles of incidence.

With this intent, the model in Table 1 was developed for the ice-bonded area of Figure 2. Layer 1 is the water column, layers 2 and 4 represent non-ice-bonded sediments, and layer 3 is a layer of frozen sediments.

<table>
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<tr>
<th>Layer No.</th>
<th>P Velocity (m/s)</th>
<th>S Velocity (m/s)</th>
<th>Poisson’s Ratio</th>
<th>Density (g/cc)</th>
<th>Thickness (m)</th>
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<td>0</td>
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<td>1.0</td>
<td>30</td>
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<tr>
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<tr>
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<td>2.0</td>
<td>70</td>
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<tr>
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<td>2000</td>
<td>1155</td>
<td>0.25</td>
<td>2.5</td>
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</tbody>
</table>

Table 1. Model developed for the ice-bonded area of Figure 2.

The model evolved from analysis of the shot record in Figure 5 as well as velocity analysis of CDP gathers from the same area. Because no data on shear wave velocities for this area were available, values of Poisson’s ratio for the four layers were estimated. The 150-m thick layer below sea bottom includes shallow marine sediments as well as the underlying unfrozen, coarser, delta plain deposits. An average Poisson’s ratio of 0.4 was therefore used for this layer, which is slightly lower than the range noted by Hamilton (1976) for shallow marine sediments (0.45-0.5). A Poisson’s ratio of 0.25 was used for the frozen layer and the sediments of the half space (layers 3 and 4). To examine the effect of incident angle on reflection and transmission coeffi-
Fig. 1. 24-fold seismic data from a permafrost affected area of the Beaufort Sea. The arrows delineate the reflection from the top of the ice-bonded sediments.
Fig. 2. Enlarged view of the area of interest in Figure 1. 24-fold data. Arrows (a) and (b) are located over and off the ice-bonded layer.


ear trace display

Fig. 3. Near-trace singlefold section from the data in Figure 2. Arrow indicates reflection from the base of the permafrost layer.
FAR TRACE DISPLAY

Fig. 4. Far-trace singlefold section from the data in Figure 2.

Fig. 5. Shot record over ice-bonded sediments from the location marked by arrow (a) in Figure 2.

Fig. 6. Shot record over non-ice-bonded sediments from the location marked by arrow (b) in Figure 2.
cients for a P wave incident on the top of the frozen layer, calculations of displacement amplitude for angles of incidence 0 to 90 degrees are plotted in Figure 7. The extreme drop of transmitted P wave amplitude at 28 degrees indicates the compressional critical angle.

In order to better relate the incident angles in Figure 7 to the model and acquisition geometry used for the data, ray-tracing was used to generate the raypaths shown in Figure 8. These paths are for plane compressional wave reflections from the sea bottom as well as the top and bottom of the ice-bonded layer. The source-receiver geometry is the same as that used for the data in Figures 1 to 6. From the raypaths, the incident angles for each interface-receiver pair can be determined. Figure 9 shows the variation of incident angle with offset for each raypath and interface in Figure 8. Interface 1 is the sea-bed (between layers 1 and 2), interface 2 is the top of the frozen layer, and interface 3 is the base of the frozen layer. Note that at the base of the high-velocity layer in this model (layer 3) the angles of incidence are higher and cover a broader range than those for the top. This effect is caused by the severe refraction that rays undergo at the large velocity contrast associated with the top of the frozen layer (Fig. 8).

With the incident angles and raypaths defined, the modelling programs were used to calculate amplitudes of the reflections from each interface for each receiver. Displacement amplitude of the reflected compressional wave is plotted for each interface in the model as a function of source-receiver offset on the surface in Figure 10. These amplitude values account for transmission loss at all interfaces crossed before and after reflection at the target interface. Comparison of Figures 10 and 5 confirms that the Zoeppritz displacement amplitude equations alone describe the relationship between observed pressure amplitude and source-receiver offset rather well.

In order to understand the effect of non-normal incidence on the wavelet shape, phase changes on reflection and transmission must also be considered. Phase of the reflected compressional wave is plotted as a function of source-receiver offset for each interface in the model in Figure 11. Any phase changes on transmission to and from the target interface are incorporated. Figure 11
illustrates that there are large and sudden changes in phase with offset for the reflections from the sea bed and the top of the frozen layer. The rapid phase change for both interfaces occurs at the offset corresponding to critical incidence (Fig. 8). This is consistent with the phase rotation observed in the corresponding reflection event on the shot record of Figure 5 (marked with a solid arrow).

DISCUSSION

To facilitate easy comparison of the graphs in Figures 10 and 11 with the shot record of Figure 5, displacement amplitude and phase, as functions of offset, have been incorporated directly in a synthetic seismogram (Fig. 12). The arrival times were determined from ray-tracing. A zero phase Ricker wavelet with a dominant frequency of 70 Hz was used to be consistent with the data. The synthetic was computed for compressional plane wave reflections only, and does not consider forms of attenuation other than transmission losses. Note that the amplitude of the reflection from the base of the high-velocity layer is a minimum (almost zero) at an offset of about 225 m. Beyond this offset the arrival time of this reflection begins to merge with that of the reflection from the top of the high-velocity layer, making comparison of the amplitudes on the synthetic with the real data difficult. Also evident on the synthetic of Figure 12 is the severe change in phase for the reflection off the top of this layer. The major phase rotation occurs approximately half-way across the record, between 180 m and 250 m offset. Obviously the variations in phase and amplitude with offset are substantial where shallow high-velocity layers occur, suggesting that caution must be exercised when incorporating these traces into CDP stacks.

The variation in phase with offset for the reflection from the top of the ice-bonded layer is more clearly seen in the data if a CDP gather is corrected for normal move-out (NMO). Figure 13 shows the NMO-corrected gather from the same area as the shot record in Figure 5, using velocities that best 'flatten' each event. It is apparent that stacking these gathered data will produce a cancellation effect for the reflection marked with the

![Fig. 10. Zoeppritz displacement amplitude versus source-receiver offset for the raypaths shown in Figure 8.](image)

![Fig. 11. Zoeppritz P wave reflection coefficients](image)

![Fig. 12. Synthetic seismogram for the model in Table 1 and the source-receiver geometry of the data in Figures 1 to 6.](image)
NMO-CORRECTED CDP GATHER
OVER ICE-BONDED SEDIMENTS

Fig. 13. Normal-move-out corrected common-depth-point gather from the ice-bonded area marked by position (a) in Figure 2. The arrows mark the reflections from the bottom of the ice-bonded layer. Note the phase rotation of the upper reflection, midway across the record (solid arrow). This explains why the reflection from the top of the ice-bonded layer is poorer in the stacked section (Fig. 2) than in the near-trace display (Fig. 3). Not only is useful information lost by the stacking process, but this effect will also be detrimental to velocity analysis techniques using a coherence measure. A CDP gather is shown in Figure 14 with 9 different constant velocity move-out corrections applied. The vertical arrow marks the gather that has been corrected at the proper velocity for the reflection from the top of the frozen layer (marked with a horizontal arrow). The variation in phase and amplitude with offset for this reflection event is apparent.

To demonstrate the significance of the cancellation effect, the data were stacked. Each panel in Figure 15 corresponds to 13 CDP gathers corrected for normal move-out at a constant velocity and stacked. The stacking velocity used in each panel corresponds with that used for each panel in Figure 14. The vertical arrow marks the correct stacking velocity as determined from Figure 14. At this correct velocity, the reflection marked with the horizontal arrow has poor coherence and low amplitude. In fact, the best coherence and the highest amplitude for this reflection are found at a significantly higher but incorrect stacking velocity. If semblance were used to pick the velocity producing the highest amplitude on this event, an incorrectly high stacking velocity would be chosen.

VELOCITY ANALYSIS PANELS -- UNSTACKED GATHER

Fig. 14. Velocity Analysis Panels - Unstacked gather from the ice-bonded area. The horizontal arrow indicates the reflection from the top of the ice-bonded layer and the vertical arrow marks the correct stacking velocity for this event.
**CONCLUSIONS**

This approach considered a simplified model and examined the variation in reflection attributes produced only by energy partitioning at non-normal angles of incidence. It has been effective, however, in demonstrating that:

1) Large changes in reflection amplitude and phase with source-receiver offset occur in high-resolution seismic data from the Beaufort Sea, where shallow, high-velocity layers occur and streamer lengths are 300 to 600 m.

2) Multifold, stacked reflection seismic sections do not necessarily provide higher-quality data than singlefold near-trace displays. Standard CDP stacking techniques will degrade reflections from the shallow part of the section.

3) Far offset information does not necessarily improve velocity determination in these data. Velocity analysis based on a coherency measure may result in artificially high stacking velocities being determined.

4) Consideration of amplitude and phase attributes on shot and CDP records can greatly aid the interpreter. It is important that an interpretation not be based only on the stacked section, but that shot and CDP gathers, common offset stacks, and velocity panels be considered as well.

5) A recording geometry that provides high fold at near offsets (i.e., small group interval and/or small shot interval) should be used in areas such as those discussed in this paper.

**REFERENCES**


