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ICE PROPERTIES AND BASAL CONDITIONS INFERRED FROM SEISMIC DATA ACQUIRED ON TWO FAST-FLOWING ICE STREAMS (WEST ANTARCTICA)

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INTRODUCTION

Ice streams are components of the ice sheet that move faster than the surrounding ice and are responsible for transporting most of the ice leaving the continent to the floating ice shelves and, ultimately, to the ocean. Direct observations of glacier motion and laboratory studies of ice rheology have identified three main mechanisms by which ice masses move. These are by internal deformation of the ice itself, and through two processes which take place at the base of the glacier; basal sliding and the deformation of water-saturated weak sediments. Flow by internal ice deformation takes place in all ice masses, and generally accounts for motion of a few meters per year. However, basal motion occurs only where the bed is at the pressure melting point such that water is present. Where basal motion takes place, glaciers may move at tens to hundreds, and sometimes a few thousand, meters per year.
The seismic method is very useful to investigate ice dynamics. In this work we present some results obtained from seismic data acquired on the Whillans Ice Stream and Thwaites Glacier (see Figure 1).

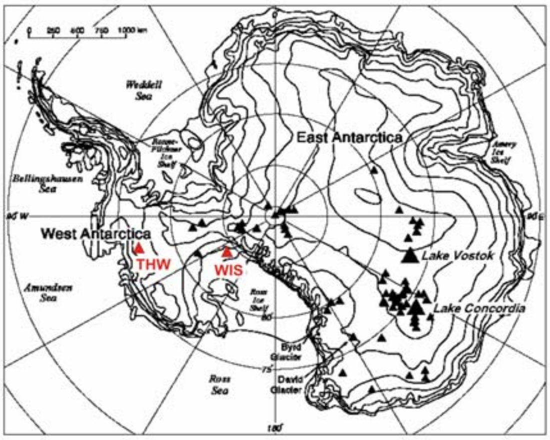


Figure 1 – Location map showing the study locations in West Antarctica (red triangles). WIS and THW indicate the locations of Whillans Ice Stream and Thwaites Glacier, respectively. Black triangles indicate the distribution of main subglacial lakes in Antarctica.

Whillans Ice Stream, previously known as Ice Stream B, is one of the major feeders of ice to the Ross Ice Shelf, Antarctica's largest ice shelf. It is up to 100 kilometres wide, hundreds of kilometres long and 2 kilometres thick at its deepest points. Despite its low driving stress, this ice stream flows at over 300 m/a, achieving this motion in short bursts interspersed with long periods of relative quiescence (Bindschadler et al., 2003; Winberry et al., 2009b). In the past decades this ice stream has been observed to be decelerating, indicating that it may be heading for stagnation (Joughin et al., 2002, 2005). Glaciological drilling and instrumentation in the late 1980s and 1990s concluded that water flow beneath Whillans Ice Stream is likely achieved by regularly spaced channels in the underlying deformable till (Engelhardt and Kamb, 1997; Kamb, 2001). In a recent work, Fricker et al. (2007) provides evidences that subglacial water is stored in a linked system of reservoirs underneath this ice stream and can move quickly into and out of those reservoirs. This draining activity may play a major role in controlling the flow of the ice.

Thwaites Glacier, located in an overdeepened basin that extends far inland, is one of the fastest and largest glaciers draining the West Antarctic Ice Sheet. Together with Pine Island Glacier, it is one of the main candidates for a potential catastrophic collapse of the marine ice sheet along the Amundsen Coast. Recent studies (e.g. Shepherd et al., 2002; Rignot et al., 2008) reported a rapid retreat, thinning, and acceleration near the grounding line of these glaciers. However, little is known on the subglacial environment of Thwaites Glacier far into the interior of the ice sheet.

In this work we show that active single and three-component seismic data can be effectively used to infer the fabric properties of the ice, to image the internal and subglacial structures of the ice sheets and to determine the bed properties of the subglacial environments. These observations will allow modelers to more accurately conceptualize the subglacial environments and ice-sheets dynamics.

WHILLANS ICE STREAM

Main purposes

Widespread anisotropic blue ice in West Antarctica likely results from crystal orientation fabrics (COF) occurring within the ice sheets. Seismic methods can be effectively used to infer COFs (Blankenship and Bentley, 1987). Typically, ice-divide seismic observations show the presence of transversely isotropic ice with vertical axis of symmetry (VTI). This may not be valid far from ice-divides, where the balance of longitudinal deviatoric stresses may play an important role in determining the axis of symmetry for fabrics. Moreover, the presence of fractures oriented along preferential directions may induce other anisotropic effects in the shallow blue-ice layers, that can mask intrinsic anisotropy. Active three-component seismic surveys provides additional insight into the anisotropic properties of ice sheets. In this work we show that three-component seismic data can be effectively used to determine if the symmetry axis is vertical and, therefore, if the medium is VTI. These information are very important in order to understand if a basal sliding regime or a subglacial deformation regime dominates.

Seismic data

Four active source seismic profiles were acquired on the Whillans Ice Stream (West Antarctica) during the austral summer 2010-2011 (Horgan et al., 2012) , with the main purpose of image a shallow subglacial lake at about 700m depth (Fricker et al., 2007). One profile was acquired sub parallel to the ice stream flow direction, and three transverse profiles were acquired across the lake. Seismic profiles were surveyed using kinematic GPS, traditional surveying and electronic distance measuring. Shot charges, consisted of 0.4 kg of an explosive composed mainly of PETN (pentaerythritol tetranitrate), were placed at a nominal depth of 27 m using a hot-water drill. During the main acquisition, three-component (3C) seismic data were also acquired, in order to determine the anisotropy within the ice sheet. To this aim, 10 passive 3C stations were used along the longitudinal and the transverse profiles. passive 3C stations are comprised of a Guralp 40-T broadband seismometer with a 40 second long period corner coupled to a Reftek RT-130 data acquisition system with GPS timing.
The detonation of an explosive charge in an isotropic medium should generate only compressional waves. However the generation of both P and S waves from explosive sources is well documented in the literature and, in particular, SH waves from buried explosive charges are commonplace (Geyer and Martner, 1969). This is due to asymmetry in the physical properties around the shot, which in our case is very common along all the profiles for the presence of many buried crevasses. This makes fractured polar firn an ideal medium to produce and study S waves.

S-wave velocity profiling from surface waves

Surface waves can represent a valuable tool for assessing qSV- and SH-wave velocities. Rayleigh and Love wave velocities are found to be mutually incompatible with a single isotropic model when the medium is effectively anisotropic. Moreover, azimuthal seismic anisotropy within the stream, caused by aligned *c*-axes in the direction of flow or aligned fractures potentially developed by the ice flow, could be evaluated by comparing Rayleigh and Love waves observed on orthogonal seismic surveys.
A 2-D transform can be applied to the recorded seismic wave field to enhance the detection and separation of the surface wave phase velocities. The algorithm was developed by Herrmann (2005) and follows the formalism described by McMechan and Yedlin (1981). The phase velocity dispersion curves are obtained from a slant stack (τ - p) of the seismic traces followed by a transformation into the ω - p domain, where p is the ray parameter. The interpretation of modal identity, often difficult for the higher modes domination and contamination in complex media, is here simpler and driven by seismic velocity models for ice. The observed phase velocities can be used to determine the structure of the ice by comparison with synthetic phase velocity in a model. The phase velocity depends primarily on the shear wave velocity and is scarcely

sensitive to realistic variations in density and compressional wave velocity with depth (Nolet, 1981). The Rayleigh and Love phase velocity curves are used to determine the shear wave velocity structure of the ice by applying a linearized inversion scheme (Herrmann and Ammon 2002). A linearization around a reference model makes it possible to find a reliable model that fits the data, and facilitates the computation of the model error (Gabriels et al, 1987).

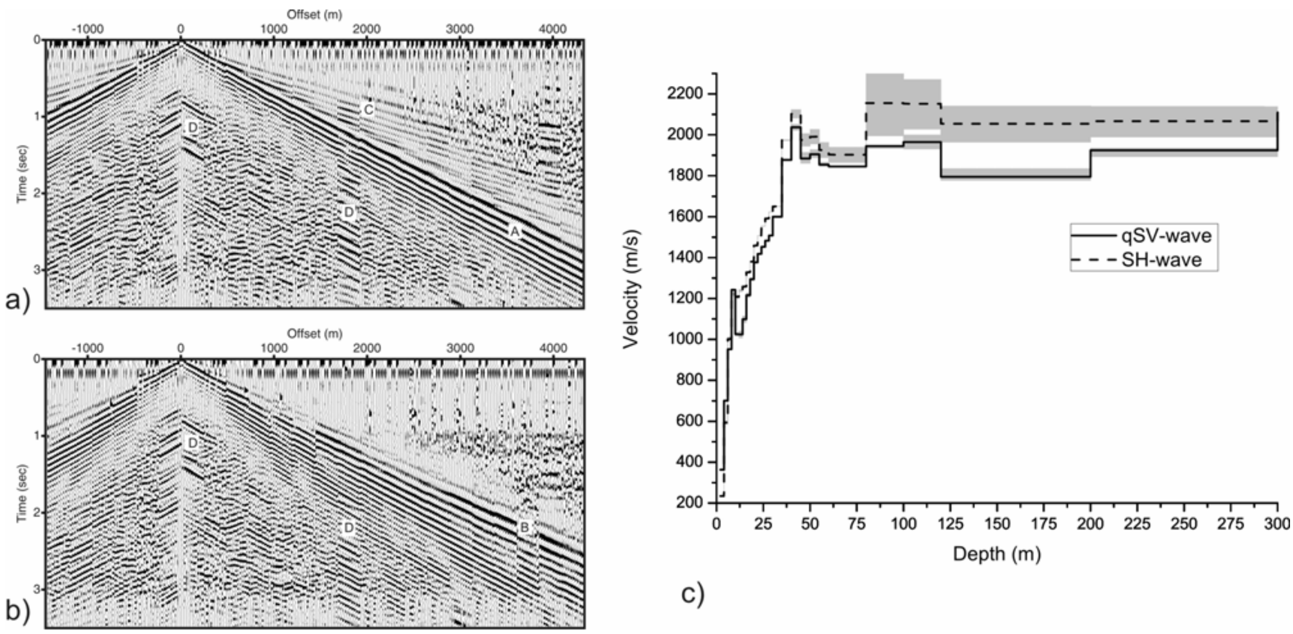


Figure 2 – Horizontal longitudinal (a) and transverse (b) components recorded by the passive 3C stations along the longitudinal profile. A 2-20Hz bandpass filter is applied to enhance the surface waves. The following main events are indicated: Rayleigh waves (A), Love waves (B), mode converted in the firn (C) and seismic events due to ice falls in the crevasses after the shots (D). Resulting qSV- and SH-wave velocities profiles obtained inverting Rayleigh and Love wave fundamental and first higher modes (c), where grey bands represent error ranges.

Results

Surface wave dispersion analysis of multichannel seismic data acquired on the Whillans Ice Stream (West Antarctica) indicate that the shear-wave splitting is the same in the two directions parallel and orthogonal to the ice sheet flow direction. More specifically, the ice at the base of the firn (about 60 m depth) is nearly isotropic, because the shear-wave splitting is negligible. With increasing depth, shear-wave splitting increases and stabilizes, at about 75 m depth, to an average value of about 130 m/s (see Figure 2c). The average qSV-wave and SH-wave velocities between 75 m and 300 m obtained from inversion are, respectively, V_{qSV}= 1967±40 m/s and V_{SH}= 2098±80 m/s. Therefore, these results indicate that the blue ice in the study area is a VTI medium, with the Thomsen anisotropic gamma (γ) parameter equal to 0.07.
We don't report evidences of englacial reflectivity, which may indicate the presence of high basal shear stress (i.e. non-deforming sediments at the bed). This observation suggests that probably all the ice sheet can be considered a VTI mass of ice. Further traveltime analysis of the compressional P waves and converted P-SV waves reflected from the bed can be used to verify if this VTI model is valid as average anisotropic model for the whole ice sheet.
This technique is useful to determine whether the axis of symmetry is vertical or not. This is important when trying to extract COF properties from seismic anisotropy, because the possible presence of longitudinal deviatoric stresses and fractures aligned along preferential directions may modify the axis of symmetry. These information are also important in order to understand if a basal sliding regime or a subglacial deformation regime dominates. In this case, the ice structure is typical of an ice sheet where basal shear stress and normal stress are very low and sliding is dominant over internal deformation.

THWAITES GLACIER

Main purposes

Recent studies (e.g. Shepherd and others, 2002; Rignot and others, 2008) indicate that the glaciers along the Amundsen Coast are thinning rapidly. In particular, the most dramatic changes have occurred on Pine Island Glacier and Thwaites Glacier, where the speed near the grounding line increased more than 25% between 1974 and 2008 (Rignot and others, 2002; Joughin and others, 2003b; Rabus and Lang, 2003; Rignot and others, 2008). These studies evidence, however, some differences between the two glaciers, suggesting that Thwaites Glacier is more stable than Pine Island glacier. Joughin and others (2009) used models constrained by remotely sensed data from Pine Island and Thwaites Glaciers to infer basal properties. The results indicate strong basal melting in areas upstream of the grounding lines of both glaciers, where the ice flow is fast and the basal shear stress is large. Farther inland, they found mixed bed conditions for both glaciers, alternating from regions of low drag (i.e. deforming sediments), to regions providing greater basal resistance (i.e. non-deforming sediments or even crystalline bedrock). In particular, for Thwaites Glacier they reported that the areas characterized by strong bed are more extensive than the weak regions, explaining the higher degree of stability with respect to Pine Island Glacier. The main purpose of this work is to verify these hypothesis using the seismic method.

Seismic data

During the 2008-2009 Antarctic field season, 60 km of reflection seismic data were collected ~200 km inland of the current grounding line of Thwaites Glacier, West Antarctica, consisting of one 40-km profile along flow and two 10-km transverse profiles (Horgan et al., 2011). In this work only the longitudinal profile is considered. The receivers were single-component vertical geophones with natural frequencies of 28 or 40 Hz. The geophones were buried at shallow depths of up to 0.5m to reduce wind noise. Synchronization between the seismic digitizer and the shot triggering was achieved by GPS time signals. Data sampling rate was 0.25ms. The survey was designed to target a change in inferred basal drag (Joughin and others, 2009), with the longitudinal profile spanning the transition between the two zones. The acquisition geometry was designed to have a maximum fold of 5: the number of channels was 48, with a receiver spacing of 20m and a source spacing of 480 m.

Processing of the seismic data

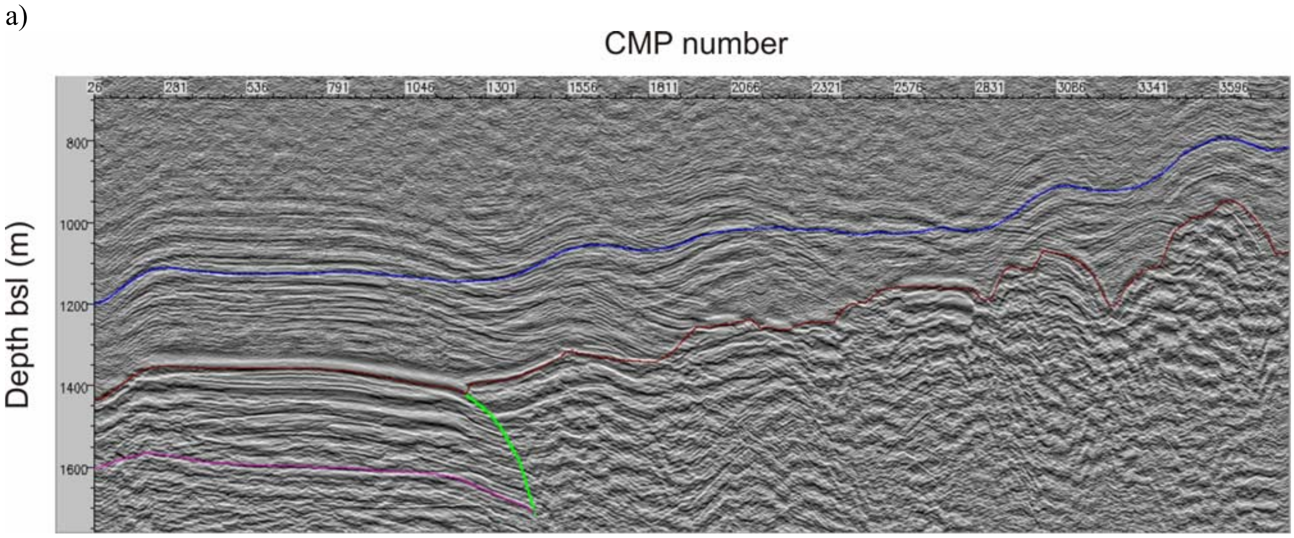
The seismic data processing is aimed at increasing the signal/noise ratio by reducing the coherent noise (i.e. ice-bottom ghosts and refracted waves travelling in the ice layers, etc.) and to increasing the lateral coherency of the reflectors of interest. Since there may be occurrence sediments between the ice and the bedrock, the seismic data must be processed in order to facilitate the detection and picking of the reflected events to be used for the travel-time inversion. The processing adopted the 'true-amplitude' approach which preserves the real amplitudes of the reflected signals (Yilmaz, 2001). This last point is very important because we performed a successive AVO (Amplitude Variations with Offset) analysis to characterise the petrophysical properties of the sediments, as described in detail in the following. The processing included:

- reconstruction of the firn vertical velocity profile and refraction statics using diving waves;
- surface-consistent predictive deconvolution for the ghost elimination and wavelet compression;
- surface-consistent residual statics using the cross-correlation method.

Tomography and Imaging refining

An earth model in depth is usually described by two sets of parameters: seismic velocities and interfaces geometry. The estimation of these parameters with a required level of accuracy make the earth model definition a challenging task for the seismic exploration. Subsurface seismic imaging is finally realised by pre-stack depth migration, which converts the seismic travel-time data into a depth section using the tomographic velocity field. The more reliable is the velocity field supplied to the migration, the more realistic will be the imaging. The modelling technique adopted in this work consists in an iterative updating procedure for refining and improving an initial model in depth, involving pre-stack depth migration, residual move-out analysis and seismic reflection tomography (Yilmaz, 2001). At each iteration, both velocity and reflector geometries are updated, until the two set of parameters reach a good degree of stability. The initial model generally consists in a horizontally layered model with laterally invariant velocities, or in a partially refined model obtained from other techniques and information. Residual move-out analysis is a velocity analysis performed after applying an initial velocity function to the data, in order to find the residual errors in the velocity field. If the initial velocity field has been estimated with sufficient accuracy, then the common image gathers (CIGs) derived from pre-stack depth migration using this model should exhibit a flat sequence of events. Any error in layer velocities and/or reflector geometries, on the other hand, should give rise to a residual move-out along those distribution of events on the CIGs, which are no-flat. In other words, the degree of no-flatness of the reflection events on the CIGs is a measurement of the error in the model, and residual move-out analysis identifies the correction required to flatten the reflections (Yilmaz, 2001).

a)



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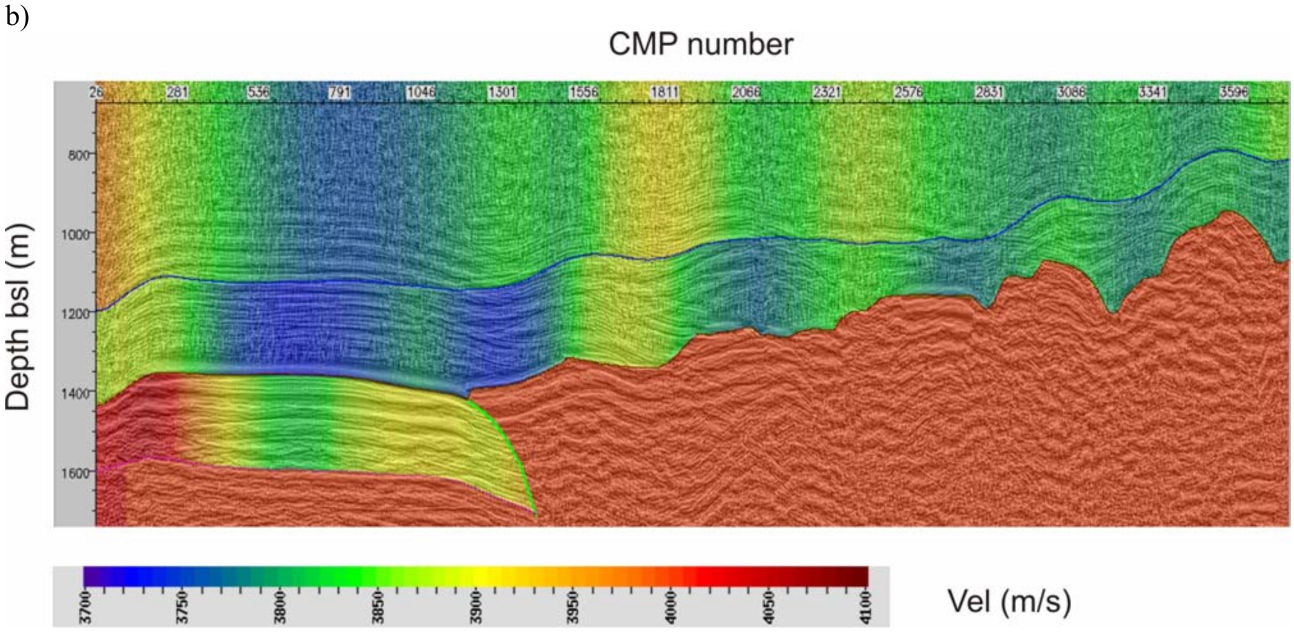

Figure 3 – Final pre-stack Kirchhoff depth migration (a) and the final tomographic interval velocity model (b), where the vertical axis indicates the depth below sea level (bls). The average glacier surface elevation is about 1300m and the CMP interval is 10m. Automatic Gain Control (AGC) is applied to enhance the weak englacial reflections.

This procedure goes on until the quality of pre-stack depth migration is not sufficient. Generally, this point is reached when the events on the CIGs become flat (Yilmaz, 2001) and the seismic energy is well focused. The final tomographic interval velocity model and pre-stack depth migration are shown in Figure 3, where the picked and inverted interfaces are indicated: the ice-bed interface (brown line), the base of a sedimentary layer (orange line), a fault (green line) and a continuous englacial reflected event (blue) throughout all the profile. The vertical axis indicates the depth below sea level (bls), while the abscissa indicates the Common Mid Point (CMP) number, where the CMP interval is 10m. The average glacier surface elevation is about 1300m. From the results of the migration, we can identify the correct morphology of the analysed structures and evaluate the reliability of the tomographic model. In this case the seismic energy in the migration is very well focused and the error in the tomographic velocity model is about 50 m/s. The migrated section and interval velocity model (Figure 3a,b) evidence the presence of a sedimentary basin in the upstream (left) part of the survey, where low basal shear stress was estimated (Joughin and others, 2009). The depositional structures of the basin, including its bottom, are clearly identifiable. In the downstream (right) part of the migrated section the tectonic features are clearly identifiable and the presence of faults is also evident. The imaged complex bed tomography seems to be in according with the estimated high basal shear stress (Joughin and others, 2009). The imaging evidences a clear continuous bed-conformable englacial reflection throughout the whole profile. In the upstream end of the profile, the englacial event is almost flat-lying. As the basal topography becomes more pronounced, the englacial horizon becomes more complicated with cross-cutting structures and greater topography. There is also a correspondence between pronounced bed topography and englacial structure. The englacial reflectors dip from upstream to downstream with their downstream ends terminating at the bed features. The tomographic interval velocity model shows no appreciable difference in velocity across the picked englacial horizon (less than the experimental error of about 50 m/s), while it shows a clear horizontal velocity variation in the ice sheet throughout all the profile.

Attenuation tomography and data preconditioning

Before performing AVO inversion, the input seismic data must be conditioned for the analysis. In fact, AVO inversion assumes that:

• input gathers are NMO corrected and flattened,

• amplitudes of the data represents the reflection coefficients.

Therefore, residual moveouts have to be applied to the data, which should be preprocessed with amplitude preserved procedures. Amplitude effects other than reflection process should be removed from the data. This includes geometrical spreading and attenuation compensation, surface-consistent amplitude recovery (to correct for instrument responses), etc.

A significant limitation of the seismic method in subglacial exploration is the large uncertainty associated with the unknown (and highly variable) firn and ice attenuation and source strength. Attenuation is a function of temperature, which is generally poorly known in glaciers. These uncertainties lead to difficulty in determining the bed properties. Uncertainties in the source strength computation can be significantly reduced using a modified version of the techniques described in Holland and Anandakrishnan (2009), while uncertainties in the attenuation computation can be reduced using attenuation tomography (Rossi et. al, 2007; Picotti and Carcione, 2006). The attenuation tomographic algorithm adopted in this work is based on the frequency-shift approach (Quan and Harris, 1997), which is based to the fact that, as the wavelet propagates within the medium, the high frequency part of the spectrum decreases faster than its low frequency part. As a result, the centroid of the signal spectrum is downshifted from f_S to a lower frequency f_R after the propagation from the source to the receiver. Under the assumption of a constant-Q model, this downshift $\Delta f = f_S - f_R$ is proportional to a linear integral of the attenuation along the ray path (Quan and Harris, 1997). In our case, the estimated average quality factors of the firn layer (Q_f) and of the blue-ice (Q_i) are Q_f=220±50 and Q_i=550±50.

Attenuation is generally approximated as an exponential decay which depends primarily on the quality (Q) factor and the travelled raypath length. Assuming that the frequency dependent nature of the problem can be exchanged with a centre frequency approximation, a single correction for each (t,x) point can be calculated. Given the effective Q-factor model and the velocity model we can obtain the correction for all the samples on the moveout corrected CMP gathers by ray tracing through the background velocity model. The amplitude correction for each sample can be approximated as follows

$$C \cong \exp\left(\pi \sum_{ray} \frac{S_i f_i}{Q_i v_i} \right), \tag{1}$$

where S_i is the length of the segment of the ray in each layer having quality factor Q_i and interval velocity v_i, while f_i is the associated average dominant frequency.

Surface-consistent scaling may be used to compensate for variations occurring at or near the surface, which affect trace amplitudes. Differences in source strength and receiver coupling as well as lateral changes in near surface attenuation contribute to undesirable variations in trace amplitudes. Processes such as AVO analysis or trace inversion, which assume relative amplitude data, produce more reliable results when the input data have been scaled in a surface-consistent manner.

AVO inversion

AVO technique is the analysis of the variation of amplitude as a function of offset. The amplitude varies with offset because the reflection coefficient varies when the angle of incidence of the wave at the interface varies. Thus, AVO analysis should actually be called AVA (amplitude versus angle) analysis, and inversion should be carried out converting the offset to angle of incidence using ray tracing. The basic phenomena is described as follows: When a P wave arrives at an interface between two layers, some of the energy reflects back to the surface and some is transmitted. The amount reflected and the amount transmitted depend on the contrast in parameters of the two layers. The relevant parameters are the P-wave velocity Vp, the S-wave velocity Vs, and the density ρ of the two layers. When the wave arrives at the interface at non-normal incidence, some of the P-wave energy is converted to shear, while at normal incidence the reflection coefficients depend only on Vp and ρ. Zoeppritz equations relate the reflection coefficients to the angle of incidence θ, the change in P-wave velocity ΔVp, S-wave velocity ΔVs, and density $\Delta\rho$ at the interface. These equations are quite complex, and this may be the reason why AVO is a relatively new technique. AVO theory evolved only after approximations to the Zoeppritz equations were developed.

a)

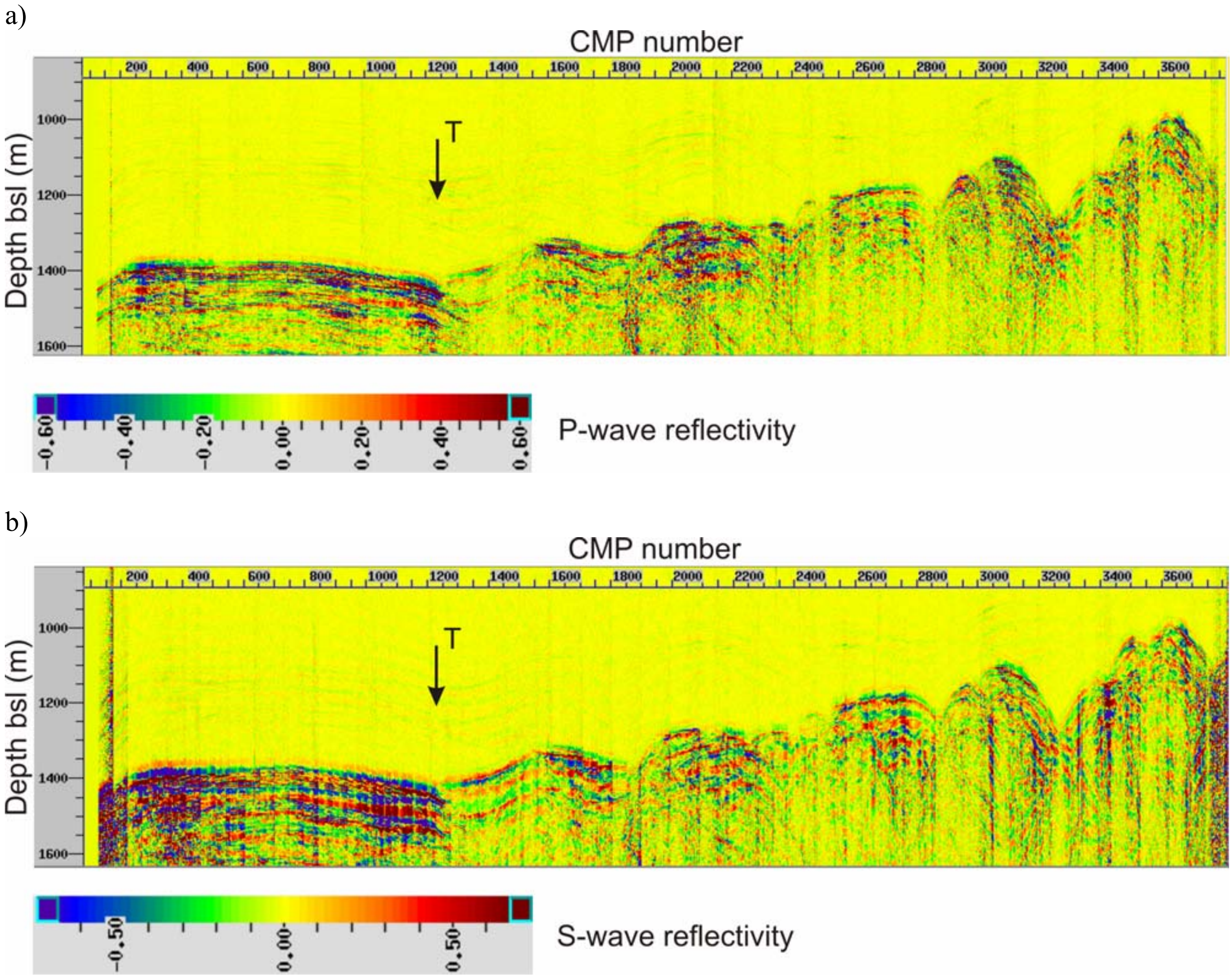

Figure 4 – P-wave reflectivity (a) and S-wave reflectivity (b) sections, final results of the AVO inversion procedure, where the vertical axis indicates the depth below sea level (bls). The average glacier surface elevation is about 1300m and the CMP interval is 10m. Indicated with T is a change in bed topography and a transition between two subglacial regimes.

In AVO inversion we take the seismic processed and balanced pre-stack data (which represents the reflection coefficients) and, using an approximation of the Zoeppritz equations, we try to convert these data to reflectivities, which have clear physical meanings, i.e. relative change in a rock parameters. The basic approximation was given by Aki & Richards (Aki and Richards, 1980). It represents the variations in reflection coefficient $R(\theta)$ with angle of incidence θ:

$$R(\theta) \cong \frac{1}{2}\left(1 - 4\frac{V_S^2}{V_P^2} sen^2\theta\right)R_\rho + \frac{1}{2\cos^2\theta} R_{VP} + 4R_{VS}\frac{V_S^2}{V_P^2} sen^2\theta, \tag{2}$$

where the three reflectivities R_{VP}, R_{VS} and R_ρ are, respectively:

- P-wave reflectivity $R_{VP}=\Delta Vp/Vp$ - relative change in *P*-wave velocity;
- S-wave reflectivity $R_{VS}= \Delta Vs/Vs$ - relative change in *S*-wave velocity;
- density reflectivity $R_\rho=\Delta\rho/\rho$ - relative change in density.

The AVO procedure used in this work can be resumed as follows:

1. Given all the source and receiver locations, we computed the angle of incidence for all gathers, offsets, and time (or depth) samples performing a ray tracing in the background tomographic velocity model shown in Fogure 3b.
2. Amplitude picking for all offsets and time samples.
3. Fit of the picked amplitudes with a Zoeppritz approximation curve (the Aki & Richards approximation) as a function of angle of incidence θ, using a least squares formulation.

Once the Zoeppritz approximation curve is derived, its parameters R_{VP}, R_{VS} and R_ρ are known. These are the inverted reflectivity values for that sample. This process is repeated for all CMPs and for all samples. The results are reflectivity data sections, which are also called AVO attributes (see Figure 4 a,b). In theory, a straight forward solution of all three parameters is possible. In practice however, the density affects the very far offsets beyond 30°. Unfortunately in most cases far offset data which contains the density information does not exist. Other factors such as noise, low coverage and imperfect amplitude recovery makes inversion for density instable and causes instability in the velocity reflectivity estimation as well. To stabilize the inversion we usually invert for two-terms rather than the three-terms that are involved in the Zoeppritz equation. In practice, in the Aki & Richards formula we substitute $R_\rho=\gamma R_{VP}$ and solve only for R_{VP} and R_{VS}..

Results and discussion

Overall, our seismic results are in agreement with the conclusions of Joughin et al. (2009). The horizontal variations observed in the interval velocity model (Figure 3b) show a clear picture of the internal deformation of the ice mass due to the high variability of the basal shear stress reported by Joughin et al. (2009). Simple shear induces a rotation of the *c*-axes, leading to an evolution of crystal fabric.

Regarding the englacial reflection, we try to give a new explanation based on fractured media theory. In the late 1950s and early 1960s, some researchers started to note the presence of these weak englacial seismic reflections (e.g. Bentley and Ostenso, 1961) a few hundred meters above the bedrock, with an amplitude of approximately one tenth that of the bedrock reflection (Bentley, 1971b). All the subsequent seismic studies have attributed englacial reflectivity to either entrained basal material or abrupt changes in COF (Smith, 1996; Anandakrishnan, 1996; Horgan et al., 2008; Horgan et al., 2011). In this case, both these two explanations does not seem plausible because we cannot see an appreciable vertical velocity variation across the picked englacial horizon, and bedrock obstacles cannot serve as a source for morainal material in ice that is well up-glacier of the obstacles. Another interpretation of this phenomenon can be the presence of fractures favoured by the proximity of a bedrock showing, as in our case, pronounced basal topography and large shear stress. Let us consider a planar fracture (*x-z*) separating two elastic transverse isotropic (TI) media. The non-ideal characteristics of the fracture are modelled through boundary conditions. The model proposed here, introduced by Carcione and Picotti (2011), is based on the discontinuity of the displacement and particle-velocity fields across the fracture interface. The boundary conditions for a wave impinging on a fracture ($z=0$) can be written as follows

$$k_x[u_x]+\eta_x[v_x]=\sigma_{xz}, \quad k_z[u_z]+\eta_z[v_z]=\sigma_{zz}, \quad [\sigma_{xz}]=0, \quad [\sigma_{zz}]=0 \tag{2}$$

(Carcione and Picotti, 2011), where u_x and u_z are the displacement components, v_x and v_z are the particle-velocity components, σ_{xz} and σ_{zz} are the stress components, k_x and k_z are specific stiffnesses and η_x and η_z are specific viscosities. They have dimensions of stiffness and viscosity per unit length, respectively. Moreover, the brackets denote discontinuities across the fracture interface. The model simulates the fracture by a zero width layer of distributed spring-dashpots. It can be shown that relaxation-like functions of Maxwell or Kelvin-Voigt type govern the tangential and normal coupling properties of the crack. The fracture exhibits time dependent mechanical properties through the relaxation functions and, as in a viscoelastic material, this implies energy dissipation. A displacement discontinuity ($k_z \neq 0$) yields a change of phase, while a discontinuity in the particle velocity ($\eta_z \neq 0$) implies an energy loss at the fracture.

A complete description of the fracture model is beyond the scope of this work (for a comprehensive description see Carcione and Picotti, 2011). Here we limit the discussion only to some considerations. At normal incidence and similar upper and lower media, the model reduces to the following relation for the P-wave reflection and transmission coefficient

$$R_{PP} = \left(i\frac{\omega_P}{\omega} - 2\frac{\eta_z}{I_P} - 1 \right)^{-1}, \quad T_{PP} = 1 - R_{PP}, \tag{3}$$

where $I_P = 2k_z / \omega_P$ and ω_P is the characteristic frequency that defines the transition from an apparently perfect interface to the apparently decoupled one. If $k_z = 0$, it is $\omega_P = 0$ and the particle-velocity discontinuity model is obtained. In this case, the coefficients are frequency independent and there are no phase changes. On the other hand, when $\eta_z = 0$, the theory gives the displacement discontinuity model. A discontinuity in the particle velocity implies energy dissipation at the fracture (Carcione and Picotti, 2011). Moreover, if $\eta_z \to 0 \;\; and \;\; k_z \to 0, \;\; R_{PP} \to 1 \;\; and \;\; T_{PP} \to 0$, and the free surface condition is obtained; when $\eta_z \to \infty \;\; and \;\; k_z \to \infty, \;\; R_{PP} \to 0 \;\; and \;\; T_{PP} \to 1$, giving the solution for a perfect welded contact.

a) b)

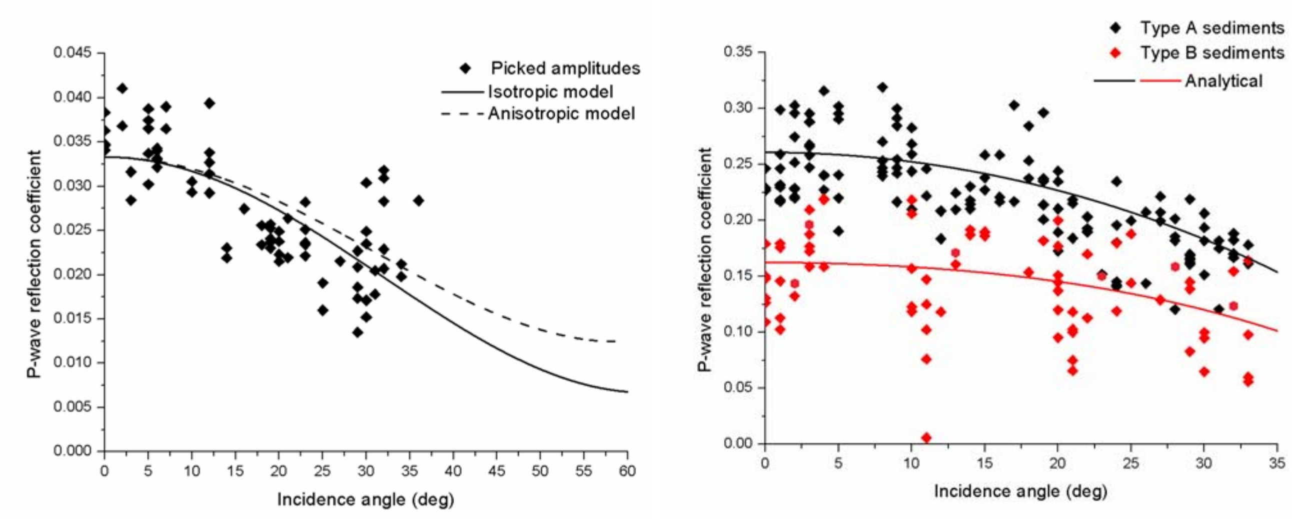

Figure 5 – Best fit of the englacial reflection coefficients using the fracture model (a). Best fit of the bed P-wave reflection coefficients corresponding to type A sediments (V_P=2850 m/s, V_S=1500 m/s, ρ=2150 kg/m^3) and type B sediments (V_P=2400 m/s, V_S=1100 m/s, ρ=2100 kg/m^3) using the full Zoeppritz equations (b).

We picked the reflection coefficients along the englacial horizon on the preconditioned CMP gathers and fit them versus angle of incidence using the fracture model (see Figure 5a). The adopted ice parameters are the following: V_P=3800 m/s, V_{SH}=2200 m/s, $|\varepsilon|$=0.015, γ=0.07 (see previous section on Whilans Ice Stream) and δ=-0.01, where V_P is the vertical P-wave velocity, V_{SH} is the SH-wave velocity and ε, γ and δ are the anisotropic Thomsen parameters. As shown in Figure 5a, the fit is very good (the error on the picked amplitudes is about 0.01), and it is not necessary to assume any difference between the upper and the lower mediums. Moreover, the plot shows that, having incidence angles lower than 40 degrees, the contribution of anisotropy is very small.

The results of AVO inversion (Figure 4 a,b) highlight moderate variability in the spatial coverage of basal sediments. We evidence alternation of low deformable sediments (type A sediments, with average P- and S-wave velocity reflectivity R_P=-0.27 and R_S=-0.29, respectively) and moderate deformable sediments (type B sediments, with average P- and S-wave velocity reflectivity R_P=-0.44 and R_S=-0.58), and some variability between these two types of sediments. The estimated reflectivities correspond to the following parameters: P-wave velocity V_P=2850 m/s, S-wave velocity V_S=1500 m/s and density ρ=2150 kg/m^3 for type A sediments, and V_P=2400 m/s, V_S=1100 m/s and ρ=2100 kg/m^3 for type B sediments. Inversion also indicates, accordingly to Joughin et al. (2009), a prevalence of type B sediments in the upstream (left) part of the survey, and a prevalence of type A sediments in the downstream (right) part of the survey. The transition between the two subglacial regimes is indicated in Figure 4 a,b, which coincides with a change in bed topography and, approximately, with a lateral change in the P-wave velocity (see Figure 3 a,b). We verified the results of inversion trying to fit, using the full Zoeppritz equations (Carcione and Gei, 2003), some reflection coefficients picked in patches of the bed with type A and type B sediments (with an error on the picked amplitudes of about 0.05). As shown in Figure 5b, the fit is satisfactory for both type of sediments.

Outputs and Budget Justification

Picotti S., Vuan A., Carcione J. M., Horgan H. J. And Anandakrishnan S., 2012. The crystalline fabric of Whillans Ice Stream (West Antarctica) inferred from multicomponent seismic data. To be submitted to Journal of Geophysical Research.

Picotti S. and others, 2012. Seismic analysis of subglacial structures and englacial reflectivity: Thwaites Glacier, West Antarctica. To be submitted to Journal of Geophysical Research.

Carcione, J. M. and Picotti, S., 2012. Reflection and transmission coefficients of a fracture in transversely isotropic media, Stud. Geophys. Geod., 56, (doi: 10.1007/s11200-011-9034-4)

A total of 5610 Euros ($8000) was allotted for this study. About 1300 Euros were used for travelling between Italy and US, 1200 Euros were used for housing in State College and about 1500 euros were spend for living there one month. A new external hard disk was brought for storing the work and data. The remaining money (about 1500 euros) will be spent for the above publications and to present the final work at PSU and conferences during 2012.

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