Seismic reflection constraints on the glacial dynamics of Johnsons Glacier, Antarctica

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Abstract

During two Antarctic summers (1996–1997 and 1997–1998), five seismic refraction and two reflection profiles were acquired on the Johnsons Glacier (Livingston Island, Antarctica) in order to obtain information about the structure of the ice, characteristics of the ice-bed contact and basement topography. An innovative technique has been used for the acquisition of reflection data to optimise the field survey schedule. Different shallow seismic sources were used during each field season: Seismic Impulse Source System (SISSY) for the first field survey and low-energy explosives (pyrotechnic noisemakers) during the second one. A comparison between these two shallow seismic sources has been performed, showing that the use of the explosives is a better seismic source in this ice environment. This is one of the first studies where this type of source has been used. The analysis of seismic data corresponding to one of the reflection profiles (L3) allows us to delineate sectors with different glacier structure (accumulation and ablation zones) without using glaciological data. Moreover, vertical discontinuities were detected by the presence of back-scattered energy and the abrupt change in frequency content of first arrivals shown in shot records. After the raw data analysis, standard processing led us to a clear seismic image of the underlying bed topography, which can be correlated with the ice flow velocity anomalies. The information obtained from seismic data on the internal structure of the glacier, location of fracture zones and the topography of the ice-bed interface constrains the glacial dynamics of Johnsons Glacier.

Keywords: Seismic interpretation; Seismic sources; Glacier environment; Near-surface seismic; Glacier structure

1. Introduction

The dynamics and evolution of glaciers strongly depend on the bed topography, the conditions of the ice–rock contact and the internal structure of the ice itself. Ice flow is dominated by gravity and the friction at the bed and sides (Whillans, 1987). There are different models to explain the role played by ice-bed contact in the ice motion (Nolan and Echelmeyer, 1999): hard-bed models, formed by rock or impermeable non-deforming sediments, and soft-bed models, which are characterised by deformable subglacial sediment called till (Patterson, 1994).

The problem of obtaining the glacier characteristics can be addressed using several geophysical methods. The most widely used are radar and seis-
mic methods. The first studies on the Antarctic continent were seismic reflection and refraction surveys performed in 1949–1952 (Robin, 1958). Other targets have been ice shelves surrounding the continent, such as the Ross Ice Shelf (Beaudoin et al., 1992) or the Larsen Ice Shelf (Jarvis and King, 1995). These early studies did not address the ice itself but the effect of the ice layer on the wavefield. Other seismic surveys were designed to investigate conditions at the ice-bed interface (Nolan and Echelmeyer, 1999), the nature of sub-ice material (Bentley and Clough, 1972), and whether or not bed deformation is occurring (Blankenship et al., 1986; Smith, 1997).

Radar surveys have also been successfully carried out to get better knowledge of the stratification and thickness of ice in polar regions (Dahl-Jensen et al., 1997), but in temperate glaciers the scattering effect of thin layers of water makes the radar technique less useful (Smith and Evans, 1972; Nicollin and Kofman, 1994). In such cases, the reflection seismic method provides an optimum way of obtaining the bed topography of the glaciers. Although these studies are not very common, a good example can be found in the study of the Lobbia glacier in the Alps (Levato et al., 1999).

The present work is a contribution to seismic investigation on glacier ice. During the 1996–1997

Fig. 1. Location of Livingston Island (South Shetland Islands, Antarctica) and map of Johnsons Glacier showing the location of the seismic reflection (L3) presented in this paper. The location of two refraction profiles (L1, L6) from which two shot gathers were chosen for the comparison of sources (Fig. 2) are indicated in grey. Dashed line shows the glacial flows: (a) Dorotea flow and (b) Johnsons flow. Shaded areas above sea level represent the outcrop zones.
and 1997–1998 field seasons, reflection and refraction seismic surveys were undertaken (Benjumea, 1999). These data constrain the ice thickness and provide an image of the ice-bed contact in the Johnsons Glacier (Livingston Island, Antarctica). For the acquisition two different shallow seismic sources were used; this paper includes a comparison of the data recorded in order to establish which is the optimum source in ice environments. A seismic reflection profile, which follows one of the main ice flows, is analysed developing (i) an analysis of the raw shot gathers and its contribution to the final interpretation, (ii) a discussion of the processing steps in order to obtain a good quality final stack, and (iii) a presentation of a structural interpretation combining the seismic reflection and glaciological data of the structure and flow parameters of Johnsons Glacier.

2. Glacial setting

Johnsons Glacier is located in the south of Livingston Island, Antarctica (Fig. 1), at 62°40’S,
60°30'W. The basement consists of sandstones (Johnsons Bay side, WNW) and contact metamorphic rocks (Smellie et al., 1995).

Johnsons Glacier results from the confluence of two glacial flow lines: the Dorotea and Johnsons lines (Fig. 1). The Dorotea line, which flows northward, is 2150 m long and has an average slope of about 6°. The Johnsons line is 980 m long. It flows south-westward and is steeper (around 10°). The flow lines converge on a non-grounded, 50-m high ice cliff, and the maximum elevations are 330 and 270 m for Dorotea and Johnsons, respectively. Annual horizontal velocities increase downstream with minimum values of about 1 ma⁻¹ near the ice divides and maximum values of 40 ma⁻¹ near the ice terminus (Ximenis, personal communication). Due to the two convergent flows, the ice is folded and highly fractured in a perpendicular direction to the flows (Calvet and Santanach, 1992).

Johnsons Glacier is a temperate ice mass. This type of glacier is characterised by a complex material consisting of ice, water, air, salts and carbon dioxide (Paterson, 1994). Temperate glaciers feature two distinct zones: the accumulation zone characterised by the presence of snow, firm (intermediate stage between snow and ice) and ice, and the ablation zone where there is only ice covered by the seasonal snow layer.

The different orientations of the ice flows relative to the prevailing NE wind direction (Villaplana and Pallas, 1993) result in a larger accumulation of snow in the Johnsons flow line than in the Dorotea one. The equilibrium line altitude (ELA), that separates accumulation and ablation zones, is situated at different levels: around 250 m for the Dorotea flow and 150 m for the Johnsons line (Ximenis, personal communication). The measured ice density is approximately 0.8 g/cm³, less than that calculated for Antarctic continental ice or for ice shelf (Furdada et al., in press).

3. Data acquisition

Five seismic refraction (2685 m total length) and two reflection profiles (2980 m total length) were carried out during two field seasons (1996–1997 and 1997–1998). We present in this paper the analysis of one of the reflection profiles. For the acquisition of this profile, single vertical 40 Hz geophone stations were deployed and a 48-channel digital seismograph (BISON 9048) was used. This seismograph uses a 16-bit A/D converter with a digital instantaneous floating point (DIFP) amplifier, which increases the dynamic range to 126 dB. The acquisition parameters were as follows: a sample rate of 0.1 ms, a 32-Hz analogue two pole Butterworth low-cut filter, a 4000-Hz analogue six-pole Butterworth high-cut filter. The record length was 500 ms. The acquisition was performed with only 24 geophones at 5-m spacing due to logistics. The shot interval was 10 m giving 6-fold data. The shots were located in front of the seismic line with a nearest offset of 30 m. This offset was chosen by taking into account the time window between the first arrivals and ground-roll. The recording cable was attached to a climbing rope and dragged, without excessive tensile stress, from shotpoint to shotpoint. Before each shot, the geophones were planted in the snow. This scheme provided a good acquisition rate allowing the field survey to be achieved in a short period of time.

4. Seismic sources

Two different seismic sources were used for data acquisition; Seismic Impulse Source System (SISSY), distributed by Dynamit Nobel (Troisdorf, Germany).

![Fig. 3. Apparent signal-to-noise ratio as a function of offset for first arrivals. The shot gathers were generated by low-energy explosives and SISSY. The data used are shown in Fig. 2.](image-url)
Fig. 4. Comparison of the logarithm of power spectrum of the two adjacent shot gathers shown in Fig. 2. Note the higher frequency content for low-energy explosives, especially between 150 and 300 Hz.

during the first field season, and low-energy explosives (pyrotechnic noisemakers) during the second one. The refraction profile L1 was acquired using the SISSY source whereas the explosive was used as the seismic source for the rest of the profiles. The characteristics of SISSY are similar to the seismic gun (Pullan and MacAulay, 1987), the main difference is that the former uses “Dynergit” cartridges, which are detonated electronically. This type of source has been used successfully in sedimentary environments (Wiederhold et al., 1998). To increase the source coupling, the SISSY was placed in the snow layer to a maximum depth of 0.6 m. The main component of the low-energy explosives used during the 1997–1998 field season was perchlorite. The explosives were buried in the snow, at the depth of the refreeze snow layer (0.5–2 m), where the source-media coupling is expected to be good. As far as we know, this is the first study to use this type of low-energy explosives as a seismic source.

We used two shot records from adjacent geophone spreads to compare the two seismic sources

Fig. 5. Seismic data (line L3) acquired on accumulation (left) and ablation zone (central and right). The main arrivals are: (a) air wave, (b) critically refracted waves from ice, (c) reverberations of (b), (d) shear and surface waves, (e) back-scattered energy of shear and surface waves from ice crevasses and (f) reflection from the glacier bed.
Fig. 6. (a) Shot gather and frequency–velocity analysis in the accumulation area, (b) same as (a) but in ablation zone (line L3).
(Fig. 2), one corresponding to SISSY (L1) and the other one to explosives (L6). These two shot gathers represent the different characteristics observed in the 1996/1997 and 1997/1998 data sets. Data were acquired in different field seasons but the homogeneous character of this zone allows this comparison. In this zone, there is a lack of crevasses because it is not influenced by the flow convergence. The thickness of the surface snow layer (between 0.5 and 1 m) as well as the characteristics of snow density, wetness and temperature history are factors that can considerably affect the signal character. The differences in these properties from one to other field season are not considerable. Therefore, it is reasonable to assume that the differences in energy and frequency content between these two shot gathers are mainly due to the character of seismic source and its coupling with the media.

Following Staples et al. (1999), we have made a quantitative analysis of the variation of signal-to-noise ratio with offset for these two seismic sources. The apparent signal-to-noise ratio is given by the equation:

$$ Q_{S/N} = \frac{A_{\text{RMS}}^{t_{0}+20 \text{ ms}}}{\rho_{\text{RMS}}^{t_{0}+14 \text{ ms}}} $$

where $A_{\text{RMS}}$ is the RMS amplitude measured in 20 ms beginning at the first break, $\rho_{\text{RMS}}$ is the RMS noise measured in a 14-ms window preceding the first break, and $t_{0}$ is the first break time. A typical wavelet length constrained the size for the window including the first arrival, whereas the noise window is established to include all the samples before first arrivals in the trace corresponding to the nearest geophone.

The $Q_{S/N}$ of SISSY is lower than that of the low-energy explosive data for the entire offset range (Fig. 3). For example, for offsets larger than 190 m, the signal-to-noise ratio for the first arrivals generated by the SISSY is close to 1, and this makes the first break picking difficult. For the other source, the
signal-to-noise ratio is above 1 for offsets up to 300 m.

The frequency content of the recorded signal is another key factor that needs to be considered when comparing different types of sources. SISSY presents a strong amplitude decay above approximately 150 Hz, whereas, the low-energy explosives feature higher values especially between 150 and 300 Hz (Fig. 4). The signal generated by the low-energy explosives has higher resolution than the signal generated by the SISSY. Furthermore, the ejection of the source during the explosion made field operations awkward and dangerous. Explosives, instead, allowed a faster shooting rate. Although SISSY has been used successfully in other geologic settings, this comparison indicates that low-energy explosives are a better source in glacier environments. These facts encouraged us to use only explosives for the second field survey.

5. Data analysis

The analysis of the raw data from a seismic reflection profile is carried out with the aim of locating anomalies within the ice and delimiting different areas of the glacier. The shot gathers along the reflection line L3 show a strong variability because of the high degree of heterogeneity of Johnsons Glacier. Nevertheless, the signal displays similar patterns in all the shot gathers (Fig. 5). The main features are described below.

The shallow depth of shot positions produce a strong air wave in some shot gathers (Fig. 5a). The

![Fig. 8. (a) Shot gather (line L3) showing the two reflection and (b) detail of these events.](image-url)
first arrivals corresponding to critically refracted waves (Fig. 5b) from the ice surface show apparent velocities of approximately 3000–3500 m/s, with some anomalous zones of 3800 m/s. In some shot gathers, these signals show a reverberation character (Fig. 5c) which is likely due to the velocity gradient in the snow layer (Crary, 1963).

High amplitude and spatially aliased surface waves are seen on all shot gathers preceded by an S-wave arrival (Fig. 5d). These waves show different seismic signatures depending on the glacier area where the data were acquired. In the ablation zone, a wavetrain composed of S and surface waves is observed with velocities corresponding to ice (1800 m/s for S-wave). The surface waves are dispersive in the accumulation zone in marked contrast to their character in the ablation zone. The distinct signal characteristics result from the different internal structure and variability of the thickness of the layer covering ice.

In the accumulation zone, this shallow layer, composed of snow and firn, is thick enough (> 10m) to provide a propagation channel for surface waves. In the ablation zone, the thickness of the snow layer is much smaller than the characteristic wavelength of the surface waves and therefore the surface waves travel with a velocity characteristic of the ice. Two shot gathers from both zones were transformed into the phase velocity–frequency domain (Fig. 6). More complexity can be observed in the accumulation zone where two peaks corresponding to two wavetrains can be identified. The lower velocity arrival is the dispersion of the surface wave caused by a shear wave velocity variation with depth (snow-firn layer). The wavetrain with higher velocity is the S wave and the surface waves, which travel along the ice with a high frequency content limited to 70 Hz. On the other hand, the S and surface waves acquired in the ablation area show a high energy content in the

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**Fig. 9. Data processing flow for data acquired in line L3.**
range 32–90 Hz, which indicates lower attenuation of seismic energy than in the accumulation zone. The non-dispersive character of the surface waves in the ablation zone is interpreted to indicate little variation with depth of the shear wave velocity in the near surface.

Vertical discontinuities in ice cause backscattered energy from shear and surface waves (Fig. 5e) and variations in the lateral continuity of events, for example, delay in arrivals of refracted waves and abrupt change in frequency content of these signals (Fig. 7). These wavefield effects were used to identify the surface location of these vertical fractures.

In some places, between the first arrivals and the surface waves, a complex hyperbolic reflection can be observed (Fig. 8). The comparison of this event with the first arrival character, as an estimate of the effective source wavelet, shows that the reflection is not a simple one. The limited depth of the shots indicates that the second hyperbolic event that forms this arrival, is not a ghost of the first one. For example, for a shot depth of 2 m and a snow layer
velocity of 1500 m/s, the ghost is expected at 2.6 ms, which is smaller than the travel time difference between both events. Thus, this complex event is likely to be the interference of reflections from a complex ice-bed.

6. Data processing

The seismic reflection data processing flow, summarised in Fig. 9, consists of standard steps designed to reveal the prominent reflection along the L3 profile. The main objective was to eliminate the high-amplitude surface waves, which interfere with the reflection. These surface waves are spatially aliased, which makes 2-D pre-stack filtering useless. Therefore, a two step processing was carried out: first, a pre-stack band-pass filter was applied to attenuate the low frequency content and second, post-stack F-K filtering was used to eliminate the remaining noise due to the surface waves with a high frequency content around 90 Hz (Fig. 10).

7. Results and discussions

The final stacked section features a variable double event reflection, which could be identified in both the raw data (Fig. 8) and the processed image (Fig. 11). As discussed above, this double event is

Fig. 11. Stack section of L3 profile. The continuous line corresponds to surface topography. (a) Sectors where the shallow layer is thick enough to provide a low velocity channel for surface waves. (b) Sector where surface waves travel with a higher velocity. (c) Area with surface waves of different character with a broad range of velocities. (d) Sector where surface and shear waves travel with velocity corresponding to ice. Areas with vertical discontinuities (crevasses) detected by back scattered energy are indicated by (e).
considered to be the interference of two reflections from the top and the bottom of a thin layer located at the base of the glacier. A minimum thickness for this layer can be calculated from the $\lambda/4$ criterion (Sheriff and Geldart, 1995). Considering a predominant frequency of these events around 140 Hz, and a range of velocities between 1400 and 2200 m/s corresponding to sub-ice material (Smith, 1997), we can estimate a range for the minimum thickness of this layer between 1 and 4 m. The 3-D bed topography effects cannot be ruled out as a cause for the complexity of the reflection wavelet. However, this explanation is not consistent with the results of a refraction profile perpendicular to the line L3 (Benjumea, 1999) showing slight lateral variation in the bed depth.

The lack of information in the first part of the seismic section (Fig. 11), between CDP 105 and 207, labelled b, is due to the lack of the near source–receiver offsets required to resolve a shallow ice-bed reflector. The maximum ice thickness in this area can be interpolated from the basement morphology obtained for the rest of the profile. This area has the shallowest ice–rock contact along the studied transect. It is interesting to note that this zone of the bed is not directly below the highest part of the profile. This fact explains some of the glaciological data anomalies measured by means of markers in the ice since 1994 (Ximenis, personal communication). The marker 1 (Fig. 12) located at the beginning of the Dorotea flow (highest part of the profile) moves in a different direction than marker 3 due to this basement knoll. In addition, the high ice flow velocity measured at marker 3 (Fig. 12) can be explained by the relatively thin ice thickness corresponding to this zone (Ximenis, personal communication). The short arched events (CDPs 430–580 in Fig. 11) that overlap the strong ice–rock reflection indicate the existence of heterogeneities at the base of the ice. Reflections cannot be observed beneath the basement reflection due to the insufficient energy penetration and the high impedance at the ice–bedrock interface.

Different sectors of crevasses, or vertical discontinuities (panel e, Fig. 11) can be identified from the back-scattered energy visible in the shot gathers (Fig. 7). Some of the discontinuities detected by the scattered energy from shear and surface waves cannot be observed from the surface due to the snow bridge.

![Sketch of the information provided by seismic section (Fig. 11). The zones of crevasses detected by reflected energy from surface waves are illustrated by grey arrows. Glaciological data (Ximenis, personal communication) show flow velocity anomalies for stakes 1 and 3.](image-url)
over crevasses. Back-scattered energy could originate from either open of closed-up crevasses, since both represent a horizontal change in the elastic parameters. On the other hand, there is a limitation in the detection of these crevasses by seismic prospecting depending on their size and orientation.

Fig. 11 shows the zones characterised by differences in the surface and shear wave character. Thus, the first part of the profile, CDP 29–126 and CDP 210–291, (a) is characterised by surface waves travelling along a low velocity channel which is correlated to a snow-firn layer. The shallow bed inferred in the first sector of the profile (b) partially coincides with the thinning of this layer (CDP 127–209). Between CDP 291 and 405, sector (c), the character of surface waves varies strongly. This area can be considered as the transition zone between the ablation and the accumulation zones. In the second part of the line, CDP 405–829, sector (d), surface waves feature a strong non-dispersive signal travelling at 1600 m/s. This zone is interpreted as the ablation zone with only a very thin snow layer (0.5 m–1 m) corresponding to the seasonal layer in summer time. An analysis of the velocity of the S-wave travelling in this zone indicates a shear wave velocity in ice of 1800 m/s. The structure of the glacier and the topography of the ice–bedrock constrained by the seismic data (Fig. 12) provided the physical explanation for the anomalies of the flow lines derived from the glaciological data.

The seismic data analysed here imaged the key structures of this temperate glacier which, in combination with the glaciological measurements, provide a model for the dynamics of Johnsons Glacier.

8. Conclusions

The experimental acquisition setup, simulating a marine streamer, provides a fast shooting rate and allows the acquisition of long profiles in a reasonable period of time, within the short Antarctic field season. In this ice environment, low-energy explosives yield a better signal-to-noise ratio, higher amplitude signals and broader frequency content than the SISSY source. The raw data analysis of the reflection profile delineates zones with large vertical discontinuities and sectors with different glacier structure (ablation and accumulation zones). The main problem during the processing of seismic reflection data acquired on the glacier was to remove strong surface waves characterised by spatial aliasing. A well-resolved seismic image of the ice–rock interface was obtained featuring a prominent but complex reflection event. This event was interpreted as an interference of reflections from the top and the base of the sediment and moraine layer between glacial ice and bed. Geophysical information obtained with this study provides the geometrical key and structural knowledge that coupled with the glaciological field observations constrain the models of ice movement of Johnsons Glacier.

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