This is an author produced version of a paper published in:
Journal of Geophysical Research: Earth Surface

Cronfa URL for this paper:
http://cronfa.swan.ac.uk/Record/cronfa33179

Paper:
http://dx.doi.org/10.1002/2016JF004047

This item is brought to you by Swansea University. Any person downloading material is agreeing to abide by the terms of the repository licence. Copies of full text items may be used or reproduced in any format or medium, without prior permission for personal research or study, educational or non-commercial purposes only. The copyright for any work remains with the original author unless otherwise specified. The full-text must not be sold in any format or medium without the formal permission of the copyright holder.

Permission for multiple reproductions should be obtained from the original author.

Authors are personally responsible for adhering to copyright and publisher restrictions when uploading content to the repository.

http://www.swansea.ac.uk/iss/researchsupport/cronfa-support/
Ice and firn heterogeneity within Larsen C Ice Shelf from borehole optical televiewing

David W. Ashmore¹, Bryn Hubbard¹, Adrian Luckman², Bernd Kulessa², Suzanne Bevan², Adam Booth³, Peter Kuipers Munneke⁴, Martin O’Leary², Heidi Sevestre⁵ and Paul R. Holland⁶

¹ – Centre for Glaciology, Department of Geography and Earth Sciences, Aberystwyth University, SY23 3DB, UK;
² – Geography Department, College of Science, Swansea University, SA2 8PP, UK;
³ – School of Earth and Environment, University of Leeds, Leeds, LS2 9JT, UK;
⁴ – IMAU, Utrecht University, P.O. Box 80000, 3508 TA Utrecht, The Netherlands;
⁵ – School of Geography and Geosciences, University of St. Andrews, College Gate, St. Andrews, Fife, KY16 9AJ, UK
⁶ – British Antarctic Survey, High Cross, Madingley Road, Cambridge, CB0ET

Corresponding author: David Ashmore (david.w.ashmore@outlook.com)

Key Points

- Larsen C Ice Shelf is composed of distinct material facies units recording its dynamic and melt history
- Existing firn air estimates commonly neglect the presence of buried low-density continental ice
- >40% of the firn zone of Larsen C is composed of refrozen ice in its central and northern sectors
Abstract

We use borehole optical televiewing (OPTV) to explore the internal structure of Larsen C Ice Shelf (LCIS). We report a suite of five ~90 m long OPTV borehole logs, recording a LED-illuminated geometrically-correct image of the borehole wall, from the northern and central portions of LCIS collected during austral spring 2014 and 2015. We use a thresholding-based technique to estimate the refrozen ice content of the borehole column and exploit a recently calibrated density—luminosity relationship to reveal the structure of each site in detail. All sites are dense and strongly influenced by surface melt, with frequent refrozen ice layers and mean densities 1.87-90 m ranging from 862 to 894 kg m\(^{-3}\). We define five distinct material facies that comprise LCIS and relate these to ice provenance, dynamic history and past melt events. These facies are in situ meteoric ice (F1), ice that has undergone enhanced densification (F2), refrozen melt pond ice (F3), compressed F2/F3 ice (F4), and advected continental ice (F5). We show that the OPTV-derived pattern of firn air content is consistent with previous estimates, but that a significant proportion of firn air is contained within our F5 unit, that we interpret to be deposited inland of the grounding line. The structure of LCIS is strongly influenced by the E-W gradient in föhn-driven melting with sites close to the Antarctic Peninsula being predominantly comprised of refrozen ice. Melting is also substantial towards the ice shelf centre with >40% of the firn zone overall being comprised of refrozen ice.
1. Introduction

Ice shelves fringe much of Antarctica and are particularly sensitive indicators of environmental changes owing to their low elevation, lower latitude and exposure to both oceanic and atmospheric forcing. Ice shelves form when glaciers and ice streams flowing across the grounding line begin to float and spread laterally to fill an embayment. Once established by this core of continental ice, they are able to gain and lose mass at their upper and lower interfaces, as well as calving at the marine margin. Although ice shelves are floating, and therefore experience negligible basal shear stress, lateral stresses and intermittent pinning to the seabed result in an ability to buttress grounded ice upstream. Through this, ice shelves modulate the contribution of the Antarctic Ice Sheet to eustatic sea level rise. Where ice shelves have thinned or collapsed inflowing glaciers are observed to have accelerated and their discharge to have increased accordingly (e.g. Paolo et al., 2015; De Rydt et al., 2015; Scambos et al., 2004).

A critical characteristic of ice shelves is their predisposition to lose mass episodically once their geometry becomes unstable (Kulessa et al., 2014). The end-member of this behaviour is a near-total disintegration of ice shelves, best exemplified by the collapse of Larsen B Ice Shelf in early 2002 (Scambos et al., 2003). The disintegration of Antarctic Peninsula ice shelves has progressed poleward for the last 30 years commensurate with climate warming (Cook and Vaughan, 2010) and a mean annual isotherm of -9 °C has been proposed as a practical limit of viability for ice shelves (Morris and Vaughan., 2003). The largest remaining Antarctic Peninsula ice shelf, Larsen C Ice Shelf (LCIS), is intersected by this isotherm and may be currently experiencing surface melting conditions similar to those that preceded the collapse of its neighbours, Larsen A in 1995 and Larsen B in 2002. The surface of LCIS has lowered during the satellite era, and approximately half of this lowering signal is caused by the loss of air from the ice shelf firn (Holland et al., 2015). This air loss may be related to
decreased surface accumulation or enhanced melting or firn compaction associated with the historical warming of the Antarctic Peninsula (Turner et al., 2016).

Proposed mechanisms of ice shelf disintegration commonly invoke intense surface melting and ponding leading to flexure and (hydro-) fracture (Banwell et al., 2013; Scambos et al., 2009). In order for melt ponds to form, snowpack pore space must be first filled (Kuipers Munneke et al., 2014) or a layer of impermeable refrozen (infiltration) ice of sufficiently large thickness and extent must exist (e.g. Machguth et al., 2016).

Distributed estimates of ice shelf thickness, a fundamental input into glaciological models, are typically derived through satellite altimetry by way of an assumption of hydrostatic equilibrium, requiring a correction for the low-density firn layer. Modelling (Pritchard et al., 2012; Chuter and Bamber, 2015) and geophysical (Drews et al., 2016) methods typically prescribe a shape to the depth-density profile. However, recent results show that intense surface melt in one NW inlet on LCIS has led to an anomalously dense ~42 m thick layer of refrozen ice and an irregular density profile over the upper 97 m (Hubbard et al., 2016). Here, the firn zone is ~10 °C warmer and ~170 kg m\(^{-3}\) denser than would be expected without melt pond refreezing. The presence of this rheologically softer ice has implications for understanding and predicting ice shelf flow and fracture in warming regions. The stability of ice shelves also depends therefore on their large-scale three-dimensional structural composition and stress field. Surface mapping (Glasser et al., 2009) and ground-penetrating radar surveys (McGrath et al., 2014) show evidence of suture ice down-flow of coastal promontories and islands. This warmer, softer ice (Dierckx et al., 2012) is thought to form by the basal freeze-on of seawater in the lee of flow obstacles and appears farther down-flow to hinder the large-scale transverse propagation of brittle fractures such as basal crevasses (McGrath et al., 2012) and rifts (Jansen et al., 2015; Kulessa et al., 2014).
The requirement for knowledge of the three-dimensional structure of ice shelves places a reliance on inferences from geophysical data or labour-intensive ice core drilling. Borehole hot-water drilling and optical televiewing (OPTV) is able to make a contribution to this requirement by being relatively rapid and logistically undemanding and, as a direct RGB image of the borehole wall is acquired, highly complementary to surface- or airborne geophysical surveys. The aim of this paper is to investigate LCIS internal structure. Herein we present five borehole OPTV records, including the one presented and discussed briefly by Hubbard et al. (2016), thereby enabling spatially-distributed ice facies fields to be considered in detail across the northern and central portions of the ice shelf.
2. Field site and data

LCIS covers an area ~50,000 km² on the eastern Antarctic Peninsula (Fig. 1). Föhn winds in the lee of the Graham Land mountains (Elvidge et al., 2015) intermittently deliver warm air to the shelf, driving intense surface melt which occasionally forms melt ponds visible in SAR imagery (Holland et al., 2011; Luckman et al., 2014). Analysis of weather stations located in the remnant Larsen B area indicates that, at least that region, has experienced föhn events of increased frequency and temperature over the last ~50 years, likely driven by a positive trend in the Southern Annular Mode (Cape et al., 2015). Airborne radar estimates of bulk firn air content, the equivalent reduction in ice thickness if it were completely comprised of bubble-free glacial ice, indicate a near-complete depletion of firn air in the NW inlets, with firn air increasing eastwards and southwards (Holland et al., 2011). Repeated radar surveys along a transect through the centre of LCIS show a loss of firn air averaging ~4 cm yr⁻¹ during 1998-2012 (Holland et al., 2015).
Figure 1. Location map of LCIS showing the sites of OPTV borehole logs overlain on 2009 MODIS “MOA” image.

Study sites were selected in order to sample the evolution in ice properties along both N-S and E-W gradients. Three drill sites lie on a flowline emanating from Cabinet Inlet (CI) spaced at 0, 22 and 120 km, and two further drill sites lie on a flowline emanating from Whirlwind Inlet (WI) spaced at 0 and 70 m along flow (Fig. 1). Boreholes are named according to their associated inlet and along-flow distance (see Fig. 1). CI-0, CI-22 and WI-0 are considered as “inlet sites” and CI-120 and WI-70 as “shelf sites.” CI-0, presented by Hubbard et al. (2016), was drilled in November 2014 and the remaining four boreholes drilled in November and December 2015. All OPTV logs begin at 1.87 m depth, the length of the OPTV sonde, and reach depths of 90 or 97.5 m (Table 1).

Table 1. Site locations and depths.

<table>
<thead>
<tr>
<th>Site</th>
<th>Start (m)</th>
<th>Depth (m)</th>
<th>Location</th>
</tr>
</thead>
<tbody>
<tr>
<td>CI-0</td>
<td>1.87</td>
<td>97.5</td>
<td>-66.403-63.377</td>
</tr>
<tr>
<td>CI-22</td>
<td>1.87</td>
<td>90</td>
<td>-66.588-63.212</td>
</tr>
<tr>
<td>CI-120</td>
<td>1.87</td>
<td>90</td>
<td>-67.001-61.486</td>
</tr>
<tr>
<td>WI-0</td>
<td>1.87</td>
<td>90</td>
<td>-67.444-64.953</td>
</tr>
<tr>
<td>WI-70</td>
<td>1.87</td>
<td>90</td>
<td>-67.500-63.336</td>
</tr>
</tbody>
</table>
3. **Methods**

3.1. *Hot-water drilling and optical televiewing*

Drilling used pressurised hot water, resulting in boreholes of ~12 cm diameter. The ice column stratigraphy was logged using an optical televiewer (OPTV) instrument. This technique has previously been described in detail by Hubbard et al. (2008). The borehole wall is illuminated by a ring of 72 sonde-mounted LEDs while a downward-looking CCD camera records a reflected image on a hyperboloidal mirror. This results in a geometrically accurate RGB image of the complete borehole wall which can be rolled and visualised as a 3D “virtual core”. As the borehole is solely illuminated by stable sonde-mounted LEDs, image brightness is a measure of borehole wall reflectivity. Elsewhere in Antarctica, OPTV has been used successfully in boreholes to identify facies associated with a rift on Roi Baudouin Ice Shelf, including the identification of marine ice and infiltration ice layers (Hubbard et al., 2012).

The OPTV image resolution is defined during acquisition and a balance must be struck as higher resolution necessitates a reduction in the logging rate. For our images we sample 360 pixels in each row and sample the borehole vertically every 1 mm. The exception to this is CI-22 where a 2 mm vertical sampling rate was used due to inclement weather conditions during acquisition.

3.2. **Density derivation**

Borehole image brightness, averaged by row, commonly referred to as luminosity has been shown to be a good proxy for firn and ice density on metre scales (Hubbard et al., 2013). The underlying principle being that borehole reflectance decreases with density under a stable light source. Density ($\rho$) is reconstructed from the equation:

$$\rho = 950 - 40e^{(0.01L)} \quad [1]$$
where $L$ is borehole luminosity (Malone et al., 2013) following Hubbard et al. (2016).

Slight coaxial misalignment of the OPTV sonde and irregularities in the borehole shape, most notably by drill hose incision into soft snow and firn of the upper sections of CI-22 and WI-70, result in occasional anomalously dark regions of some OPTV sections. This effect is caused by the narrow, incised notch being improperly illuminated and appearing as a quasi-vertical “shadow” on the complete OPTV log. To minimise this effect on our estimated density profiles we reject the upper and lower quartile brightness values on each row when calculating luminosity. Across all logs the mean difference between filtered and unfiltered density is $-0.066 \pm 3.30$ $\text{kg m}^{-3}$. In the upper part (1-30 m) of WI-70, where hose incision is most apparent the mean difference is $-1.15 \pm 2.50$ $\text{kg m}^{-3}$. Firn air volume fraction ($\phi$) is calculated for each depth bin from the OPTV derived density ($\rho$) where the density of glacial ice is taken as $\rho_i = 917$ $\text{kg m}^{-3}$ by:

$$\phi = 1 - \frac{\rho}{\rho_i} \quad [2].$$

To estimate firn air content of the upper 1.87 m, which is not televiewed due to the length of the sonde, we use 0.6 m deep snow pits at CI-0, CI-60 (i.e. 60 km along flowline) and CI-120. Here density was measured gravimetrically using a cylindrical cutter (see Proksch et al., 2016) at 0.05 m increments yielding a mean density of $357 \pm 8.8$ $\text{kg m}^{-3}$. We then assume a linear increase in density from this value to the mean OPTV-derived density in the upper 0.1 m of the OPTV log and add this to our OPTV-derived firn air thickness.

3.3. Facies analysis

Ice character is a function of (1) surface conditions at and shortly after deposition and (2) post-burial compaction-metamorphism and deformation. The principle adopted here is that spatio-temporal differences in these processes lead to classifiable ice facies. Our approach is to
describe facies based on layer style, density characteristics and depth without assigning a genetic mechanism. These can then be interpreted subsequently in terms of physical processes and environments. While some environments lead to distinct facies contrasts (e.g. Hubbard et al., 2012) others lie upon a spectrum where the effect of a particular process becomes increasingly dominant. In such cases the division between facies is made in a consistent, reproducible manner. We achieve this by exploiting the situation whereby the principal characteristic of the facies we report is the contrast between the host ice and included layers. To quantify this we define, $\sigma_{\rho}$, the standard deviation of density over a 2 m moving window with depth. This practical definition is adopted as 2 m represents a distance that likely contains several years of accumulation but one which is not strongly influenced by the decrease in brightness associated with densification due to burial. Thresholds of $\sigma_{\rho}$ can then be stipulated in response to potentially ambiguous facies boundaries.

3.4. Image thresholding and infiltration ice estimation

Three principal processes affect the stratigraphy of ice shelves. The accumulation of snow may result quasi-regular surface-parallel layering reflecting depositional differences during accumulation, often referred to as primary stratification in structural glaciology (Hudleston, 2015). Second, the snowpack undergoes depth-related densification by compaction-metamorphism driven by snow temperature and the weight of the overlying snow. Third, surface meltwater that percolates into a cold snowpack refreezes to form bubble-poor, dense infiltration ice, identifiable as dark (low reflectance) layers in OPTV logs. Furthermore, ice inherited from inland may also be overprinted during a variety of brittle and ductile deformation processes relating to dynamic flow, for example, by the incorporation of debris (Glasser et al., 2014) or platelet ice of marine origin (Hubbard et al., 2012).
Primary stratification in our LCIS logs is typically faint and disrupted by melt processes (see Section 4.1), so we can consider OPTV images to chiefly comprise of infiltration ice layers overprinted onto a densifying host ice. The resulting bimodal distribution of pixel brightnesses lends itself well to image thresholding analysis (e.g. Kinnard et al., 2008) which allows the consistent and reproducible identification of infiltration ice layers. It is important, however, to consider the main ice facies when interpreting thresholding results. Where ice is relatively homogenous other features, which may arise from trapped bubbles, incorporated debris or healed fractures may be identified by the algorithm. In this case the threshold-identified layers do not coincide with refrozen ice layers.

Uneven illumination is a well-known issue in the successful application of thresholding algorithms. Here, we prepare our OPTV logs prior to thresholding in order to minimise lateral and vertical biases in image brightnesses not related to glacio logical features and maximise the effectiveness of image thresholding. Areas of anomalous illumination are removed by row (Section 3.2) and images were smoothed with a 1x10 pixel wide median filter to highlight vertical gradients in brightness. Images were detrended with a sum-of-sine function fitted to the luminosity profile in order to remove long-wavelength density variations with depth (Fig. 2). Thresholding was then performed sequentially on 2 m image sections to achieve optimal results as some minor long-wavelength variations in image brightness remain with depth even after detrending. Three common auto-thresholding algorithms were tested by trial and error: those of Ridler and Calvard (1978), Kittler and Illingworth (1986), and Otsu (1979). Otsu’s (1979) method performed most consistently and was adopted for this investigation.

Otsu’s (1979) auto-thresholding technique divides a given distribution’s population into two clusters based on a threshold value and calculates the difference between these clusters. The optimal threshold is arrived at by finding the values at which the maximum difference
between these clusters is attained. The result is a binary image in which each pixel is classified as either low or high brightness. For our application herein, any pixel row classified as over 90% “low brightness” is identified as “dark” (Fig. 2), allowing OPTV images to be characterised in terms of their thresholding output percentage.

Figure 2. Stages of thresholding as applied to three environments (figure rows) from the CI-120 OPTV borehole log. Columns show: (a) Raw OPTV log; (b) The result of the rejection of upper and lower quartile values by row, vertical detrending and the application of the median filter; and (c) The result of thresholding to reveal the “dark” layer population.
4. Results

4.1. Optical televiewing and density profiles

Optical televiewer logs and the derived density profiles from the northerly Cabinet Inlet flowline are presented in Fig. 3 and those from the southerly Whirlwind Inlet in Fig. 4. Borehole logging reveals the presence of layers and distinct units across all five sites with several logs displaying clear, visually identifiable step changes in style (explored further in Section 4.2). The ice column of LCIS varies in both N-S and E-W directions in terms of layer presence, layer style, contrast between layers and host ice, and the decrease in host ice brightness with depth. Bright host ice is increasingly present in the easterly (seaward) and southerly directions, particularly in the upper part of the logs, while dark, homogenous ice is increasingly present in the westerly (inland) and northerly directions. All the LCIS logs contrast strongly with OPTV logs from other polar ice masses, including Derwael Ice Rise and Roi Baudouin Ice Shelf, East Antarctica (Hubbard et al., 2013) and Summit, Greenland (Hubbard and Malone, 2013). In these logs images darken gradually with depth, discrete dark layers are infrequent and faint regular banding is commonly visible. Such features are entirely absent from the LCIS results presented here.

The mean density from all our sites is 873.1 ± 39.2 kg m\(^{-3}\), denser and less variable than Roi Baudouin Ice Shelf where mean OPTV-derived density 1.87 – 66.1 m is 742.4 ± 86.8 kg m\(^{-3}\) (Hubbard et al., 2013). The spatial variation in firn/ice column identified visually is reflected in the luminosity-derived density results (summarised in Table 2). They demonstrate that CI-22 has the highest mean density, while WI-70 has the lowest mean density, although only separated by 32.6 kg m\(^{-3}\). Density standard deviation indicates that CI-22 has the least variable ice and firm column while WI-70 has the most variable profile. Firn air content estimates affirm this spatial pattern as they are directly calculated from density. With the
exception of CI-0, a picture emerges of the large-scale density characteristics of LCIS, where the densest ice is centred towards the north and inland portions, with the ice shelf becoming progressively less dense towards the south and the calving margin. CI-0 and WI-0 both host a density inversion at ~45 m and ~65 m depth respectively (Table 3). Using a 2 m moving average of density we estimate the depth to the pore closure density of 830 kg m\(^{-3}\) (Table 2), this is important as it represents in the maximum depth percolating meltwater can possibly reach before refreezing. Despite the lower mean density of CI-0 the pore closure depths from all sites conform to the W-E and N-S pattern, due to the thick layer of impermeable pond ice at 2.9 m depth, as noted by Hubbard et al. (2016).

Table 2. Summary of density characteristics from LCIS boreholes. Mean density and firn air are calculated over the common depth range of 1.87-90.00 m for all logs and 1.87–97.50 m for CI-0 (parenthesised). Surface corrected firn air content refers to the firn air content with the firn air contained within the upper 1.87 m added. Pore close off density is taken as 830 kg m\(^{-3}\) and determined with a 2 m moving average of the density profile.

<table>
<thead>
<tr>
<th>Site</th>
<th>Depth mean density (± 1σ) (kg m(^{-3}))</th>
<th>Firn air content (m)</th>
<th>Firn air content (surface corrected, m)</th>
<th>Pore closure depth (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CI-0</td>
<td>872.2 ± 30.4 (868.3 ± 32.0)</td>
<td>4.31 (5.08)</td>
<td>5.18 (5.95)</td>
<td>3.29</td>
</tr>
<tr>
<td>CI-22</td>
<td>894.2 ± 21.8</td>
<td>2.20</td>
<td>2.87</td>
<td>4.18</td>
</tr>
<tr>
<td>CI-120</td>
<td>879.0 ± 27.7</td>
<td>3.65</td>
<td>4.46</td>
<td>6.05</td>
</tr>
<tr>
<td>WI-0</td>
<td>873.2 ± 42.7</td>
<td>4.21</td>
<td>4.99</td>
<td>7.67</td>
</tr>
<tr>
<td>WI-70</td>
<td>861.6 ± 51.8</td>
<td>5.32</td>
<td>6.18</td>
<td>8.43</td>
</tr>
</tbody>
</table>
Figure 3. Borehole OPTV raw log and density profile (left), binary thresholding output (middle) and facies classification (right) results from (a) CI-0, (b) CI-22 and (c) CI-120.
Figure 4. Borehole OPTV raw log and density profile (left), binary thresholding output (middle) and facies classification (right) results from (a) WI-0 and (b) WI-70.
4.2. Facies classification

Five facies are defined on the basis of characteristics revealed by the OPTV logs (Fig. 5-7), named F1-F5 with increasing number broadly referring to increasing depth within the ice column, although not all facies are present at all sites. The distribution of facies with depth is presented in Fig. 3 (Cabinet Inlet flowline) and Fig. 4 (Whirlwind Inlet flowline), and summarised in Table 3.

F1 is present in all logs as the uppermost unit and is characterised by depth-darkening host material over metre scales and as containing frequent dark layers which may be centimetres to decimetres thick (Fig. 5a-e). The brightness of dark layers decreases also with depth, for example, Fig. 5a shows clearly identifiable dark layers, but they are not as prominent as those in Fig. 5b or 5d. Host ice darkening rates are site-specific, such that despite occurring at a shallower depth CI-120, 17.5-19.5 m is visibly darker than WI-70 20 – 22 m (cf. Fig. 5d and 5c). We note that F1 is a dynamic environment in which melt and densification are ongoing at significant levels. With this, we note that the CI-0 log was collected in Nov 2014, whereas all others were acquired in Nov/Dec 2015 and that melt ponds were visible in Cabinet Inlet in March 2015.

F2 contains bright host ice and dark layers some centimetre or decimetres thick (Figs. f-j), but the density contrast between host ice and layers is much reduced relative to F1 (Figs. 3 and 4). F2 host material is variable but not darkening on the metre scale with depth. Fig. 5e demonstrates the difficulty in qualitatively defining facies boundaries and thus we define the F1-F2 transition as the shallowest point $\rho$ (Section 3.3) decreases below a threshold of 5 kg m$^{-3}$. This value represents the start of a new regime although $\rho$ may exceed this value deeper in the log due to occasional bright firm layers (Fig. 5g). The F1/F2 boundary may be visually striking, as in WI-70, or gradual, as in CI-120.
Figure 5. Rolled virtual core images of LCIS borehole OPTV logs arranged by depth (not to scale) of F1 and F2 with actual depth given in metres.

F3 is dark without defined layers as present in F1/F2 (Fig 6a-g). However, structure is visible by bright speckles that form clouds, layers and, in one case, vertical streaking structures (Fig. 6a and 6g). Layers are typically diffuse and faint (Fig. 6b and 6e) but can be well defined (Fig. 6c). F3 is present only in CI-0 and CI-22, forming units of 41.57 and 9.74 m thick respectively. The upper transition into F3 is abrupt and the base of F3 units is moderately abrupt and followed by a reappearance of dark layers (Fig 6f and 7a).
F4 occurs exclusively beneath F2 and is characterised by a decrease in contrast between host material and layers due to darkening of the normally bright host material. Structure remains visible in some F4 sections (Figs 6h, 6j and 6k) as relatively dark layers overprinted on a lighter host material similar to those in F1 and F2. However, structure within F4 is less clear deeper in CI-22 (Fig. 6i) where it appears near-uniformly dark and homogenous. These sections without visible structure are characterised by occasional bright speckles similar to those in F3 (Fig. 6i). The F2/F4 transition is also variable and may be relatively abrupt as in CI-120 (Fig. 3c), moderately abrupt as in CI-22 (Fig 3b), or gradual as in WI-70 (Fig. 4b).

Therefore we use $\sigma_\rho$ in an effort to define this facies boundary consistently, adopting the working definition that the F2/F4 transition occurs where $\sigma_\rho$ drops below 5 kg m$^{-3}$ and does not exceed it for the remainder of the log.
Figure 6. Rolled virtual core images of LCIS borehole OPTV logs arranged by depth (not to scale) of F3 and F4 with actual depth given in metres.

F5 is characterised by an increase in brightness with depth, yielding an equivalent decrease in luminosity-derived density in this facies (Table 3). F5 is present in only in the two landward sites (CI-0 and WI-0), located closest to the mouths of the two inlets (Fig. 1). F5 contains steeply dipping layers (Fig. 7), sporadic decimetre scale thick dark layers (Fig. 7a and 7b) and metre thick sections of undisrupted ice (e.g. Fig 7d). Layers of different styles cut across one
another, as shown in Fig. 7a and 7f. These dipping layers may have diffuse edges (Fig. 7b and 7e) or be well defined (Fig. 7d). Layers and features are not exclusively (quasi-) planar, for example in CI-0 (Fig. 7b) a dipping bright layer beginning at 58 m depth is associated with a lighter region between 57.2 – 57.5 m. In WI-0 (Fig. 7e) a brighter layer at ~63.9 m appears to branch resulting in a spur centred at 63.4 m. In both CI-0 and WI-0 F5 extends beyond the base of the log and total thickness is unknown (Table 3).

Figure 7. Rolled virtual core images of LCIS borehole OPTV logs arranged by depth (not to scale) of F5 with actual depth given in metres

4.3. Image thresholding

Image thresholding results are displayed in Fig. 3 for Cabinet Inlet and Fig. 4 for Whirlwind Inlet, and summarised in Table 3. In those sections where the brighter host material is overlain by darker, low reflectance layers, our thresholding algorithm performs well with identified layers corresponding closely to those identifiable by eye (Fig. 2). However, results become more uncertain in cases of reduced density contrast between melt layers and host ice.
Where ice is comprised of a less dense host ice overprinted by dense refrozen ice layers the proportion of ice column comprised by refrozen material is equal to the calculated thresholding output percentage (Table 2). F1 has a thresholding output percentage ranging from 6.5% at CI-0 to 48.6% at CI-120. F2 has a thresholding output percentage ranging from 41.0% at CI-120 to 56.4% at CI-22. In F3 the typical host/layer configuration is not present and the thresholding algorithm identifies the absence of speckle clouds and layers, rather than the presence of dark layers, resulting in a thresholding output percentage of 47.7% at CI-0 to 66.0% at CI-22. In F4 at CI-120 and WI-70, layers correspond well to visible layers, yielding thresholding output percentages of 41.0% and 56.0% respectively. However, in F4 in the lower part of CI-22 has a thresholding output of 52.2% but its effectiveness is limited by an ice column composed almost exclusively of ice in this zone. F5 has thresholding output percentages of 17.3% at CI-0 and 27.5% at WI-70.
Table 3. Summary of facies, facies density and thresholding output percentage (% Dark) results from LCIS borehole OPTV logs. * denotes F3 where assumption of bright background and dark layering is not valid. ** denotes that the base of the unit is unknown. Parenthesised thickness and firn air contents represent values with the upper 1.87 m (not sampled by OPTV log) included.

<table>
<thead>
<tr>
<th>Site</th>
<th>Facies</th>
<th>Depth range (m)</th>
<th>% Dark</th>
<th>Mean density ($\pm 1\sigma$) (kg m$^{-3}$)</th>
<th>Firn air (m)</th>
<th>Unit thickness (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CI-0</td>
<td>F1</td>
<td>1.87 – 2.90</td>
<td>6.5</td>
<td>696.3 ± 67.0</td>
<td>0.25 (1.12)</td>
<td>1.03 (2.90)</td>
</tr>
<tr>
<td></td>
<td>F3</td>
<td>2.90 – 44.87</td>
<td>47.7*</td>
<td>892.5 ± 7.1</td>
<td>1.12</td>
<td>41.97</td>
</tr>
<tr>
<td></td>
<td>F5</td>
<td>44.87 – 97.50</td>
<td>17.3</td>
<td>852.4 ± 21.0</td>
<td>3.71</td>
<td>52.63**</td>
</tr>
<tr>
<td>CI-22</td>
<td>F1</td>
<td>1.87 – 5.90</td>
<td>29.9</td>
<td>822.9 ± 37.4</td>
<td>0.41 (1.08)</td>
<td>4.03 (5.90)</td>
</tr>
<tr>
<td></td>
<td>F3</td>
<td>5.90 – 15.64</td>
<td>66.0*</td>
<td>896.7 ± 5.4</td>
<td>0.22</td>
<td>9.74</td>
</tr>
<tr>
<td></td>
<td>F2</td>
<td>15.64 – 47.61</td>
<td>56.4</td>
<td>889.6 ± 18.3</td>
<td>0.95</td>
<td>31.97</td>
</tr>
<tr>
<td></td>
<td>F4</td>
<td>47.61 – 90.00</td>
<td>52.2</td>
<td>903.8 ± 2.3</td>
<td>0.61</td>
<td>42.39**</td>
</tr>
<tr>
<td>CI-120</td>
<td>F1</td>
<td>1.87 – 29.54</td>
<td>48.6</td>
<td>856.6 ± 38.3</td>
<td>1.82 (2.63)</td>
<td>27.67 (29.54)</td>
</tr>
<tr>
<td></td>
<td>F2</td>
<td>29.54 – 68.56</td>
<td>36.5</td>
<td>883.9 ± 9.5</td>
<td>1.41</td>
<td>39.02</td>
</tr>
<tr>
<td></td>
<td>F4</td>
<td>68.56 – 90.00</td>
<td>41.0</td>
<td>899.0 ± 2.1</td>
<td>0.42</td>
<td>21.44**</td>
</tr>
<tr>
<td>WI-0</td>
<td>F1</td>
<td>1.87 – 17.34</td>
<td>47.5</td>
<td>825.1 ± 75.2</td>
<td>1.74 (2.52)</td>
<td>15.47 (17.34)</td>
</tr>
<tr>
<td></td>
<td>F2</td>
<td>17.34 – 64.95</td>
<td>55.4</td>
<td>889.5 ± 13.5</td>
<td>1.37</td>
<td>47.61</td>
</tr>
<tr>
<td></td>
<td>F5</td>
<td>64.95 – 90.00</td>
<td>27.5</td>
<td>876.8 ± 7.6</td>
<td>1.10</td>
<td>25.05**</td>
</tr>
<tr>
<td>WI-70</td>
<td>F1</td>
<td>1.87 – 32.52</td>
<td>38.8</td>
<td>818.0 ± 67.1</td>
<td>3.31 (4.17)</td>
<td>30.65 (32.52)</td>
</tr>
<tr>
<td></td>
<td>F2</td>
<td>32.52 – 72.48</td>
<td>43.0</td>
<td>880.0 ± 12.2</td>
<td>1.61</td>
<td>39.96</td>
</tr>
<tr>
<td></td>
<td>F4</td>
<td>72.48 – 90.00</td>
<td>56.0</td>
<td>896.1 ± 4.0</td>
<td>0.40</td>
<td>17.52**</td>
</tr>
</tbody>
</table>
5. Discussion

5.1. Facies interpretation

The framework for the interpretation of OPTV logs in terms of physical conditions and processes is guided by research indicating LCIS is strongly influenced by föhn winds (Elvidge et al., 2015), which drive intense surface melting and ponding (Holland et al., 2011; Luckman et al., 2014). Föhn-driven melting is focused on the inland portion of the ice shelf, although some melt occurs across almost all LCIS during summer (Luckman et al., 2014; van Wessem et al., 2016). Therefore our interpretative framework is designed to accommodate the OPTV signature of (a) inland-focused, intense, episodic föhn-driven melt and (b) spatially-widespread (but of variable intensity) seasonal melting.

F1 represents snow and firm accumulated on the ice shelf that is progressively undergoing compaction-metamorphism and moderate melt infiltration and refreezing. These processes lead to a facies in which the host ice is densifying and interrupted sporadically by layers of refrozen melt. This observation confirms that substantial melting, meltwater percolation and refreezing on LCIS has been occurring for several decades. The upper 1.87 m is not sampled by the OPTV log and can be safely assumed to also comprise of F1. Thus, F1 has a thickness of 29.54 m at CI-120 and 32.52 m at WI-70. It is relatively thin at CI inland sites (2.90 m at CI-0, 5.90 m at CI-22) as föhn-driven melting and percolation events change the ice character to form either F2 or F3 and prevent the accumulation of a thick F1 layer. F1 at WI-0 is 17.34 m thick, which is thicker than expected given the site’s proximity to the grounding line; however, our method based on 2 m $\sigma_p$ may overestimate the true depth of the F1-F2 transition here due to the colder mean annual temperature resulting in less föhn-driven melting. Melt layers are present throughout F1 at all sites, including shelf sites (i.e. CI-120 and WI-70, see Section 2), which, beneath the upper few metres, must be some years or decades old,
indicating that the facies is formed by spatially-widespread summer melting, rather than being driven by föhn events restricted to inlet locations.

F2 is material that has experienced enhanced firnification due to either föhn-driven melting, not sufficiently intense to form melt ponds (e.g. F2 in CI-22 and see discussion below) or due to climate gradients not directly attributable to föhn events along the flowline. At CI-120 the F1-F2 boundary is unlikely to be directly related to föhn melting as it would have been at the surface downstream of the region thought to be affected by föhn events (Luckman et al., 2014). Van Wessem et al., (2016) discuss the importance of local topography on the meteorology on the eastern Antarctic Peninsula. Therefore the transition to F2 ice may partially reflect the movement of ice away from the influence of topographic features (e.g. Cole Peninsula and Francis Island) and towards open ice shelf conditions. Regardless of the precise mechanism, this warming drives increased compaction rates and meltwater production, which is then available for percolation and refreezing. Despite the dense host ice, refrozen layers are visible throughout F2 and appear to be successfully identified by our thresholding algorithm. There is a potential for equifinality as F1 may densify towards material with an F2-like appearance and there is unlikely to be a clear spatial limit of föhn melting or on the effect of local topographic features. Nevertheless, the step-change in host/layer density contrast at the F1/F2 transition at CI-120 and WI-70 increases our confidence that these are indeed distinct ice facies.

Our analysis of the OPTV properties of F3 is consistent with its interpretation as “pond ice” (Hubbard et al., 2016) formed proximal to the location of maximum föhn warming. Here meltwater is abundant enough to refreeze into near-continuous units of massive ice and the meltwater may manifest as surface melt ponds prior to refreezing. The exact mechanism of F3 formation must allow water in the snowpack to become sufficiently mobile to percolate to the previous F1/F3 interface. In this way F3 is built up episodically after periods of extreme
melt. Our interpretation of F3 being formed by a significant volume of mobile liquid water is bolstered by features visible in the log. For example, at CI-0 at 4.6 m (Fig. 6a and 6g) a quasi-vertical bright streaking feature is observed. This is interpreted as being formed by an upper and lower freezing-front rejecting dissolved gases and forming two coalescing vertical bubble-trains. The bright speckles that are present throughout F3 are interpreted as bubble clouds. Typically these bubbles form into layers and vary in prominence, as seen in CI-0 22-24m (Fig. 6c), this contrasts within firn layers within F4, as visible in Fig. 6c at the base of F4 in CI-22.

F4 is comprised of ice which has undergone enhanced densification in the past but the density contrast between host ice and melt layers is now noticeably diminished. This may be due to in-situ compression of F2 ice as it is buried or represent a change in surface conditions, the manifestation of which is subsequently advected and progressively buried. Consequently, this transition may be abrupt where a change in surface conditions is rapid in either time or space, or gradual where surface conditions change slowly spatially or in cases where the transition reflects the vertical compaction-metamorphism of F2. Furthermore, F4 can be comprised of metamorphosed F2 or F3, such that the transition is more likely to be abrupt where it underlies F3 or less abrupt where it underlies the more variable but generally less dense F2. The transition into F4 is the most subjective of our facies boundaries, and we have therefore attempted to identify it quantitatively as the point at which $\sigma_{\rho}$ drops below 5 kg m$^{-3}$ and does not exceed it for the remainder of the log (Section 4.2). In CI-120 this corresponds to a clear visible boundary, which we are confident in attributing to a real facies boundary. In CI-22 and WI-70, however, we are less confident of the precise boundary.

F5 is ice formed upstream of the region of föhn-driven melting at an inland location of sufficient elevation to experience reduced surface melt intensity. Since F3 ice is interpreted to form though melt pond refreezing, the surface outcrop of the boundary between F3 and F5
likely corresponds broadly with the grounding line. This interpretation is supported by the
presence of steeply dipping layers up to several metres thick and of variable brightnesses,
which we interpret as healed crevasses and other layers having experienced ductile
deformation as ice passes over the grounding line. These features are more abundant in CI-0
than WI-0 due to the configuration of feeder glaciers in CI leading to a more compressive
regime than in WI (Fig. 1). F5 is potentially comparable to the “continental ice” reported by
Craven et al. (2005) within Amery Ice shelf. Similar to Amery, this unit, discounting its
contorted layers, is bubble-rich and less dense relative to the overlying facies.

5.2. Spatial variability of firn air content and refrozen ice

The pattern of firn air distribution estimated from our OPTV logging (Table 2) is broadly
consistent with the pattern identified by Holland et al. (2011). Their results indicate an
increase in firn air from 3.0 m at CI-0, to 6.5 m at CI-22, to 10.2 m at WI-0, to 10.6 at CI-120
and 12.0 m at WI-70. One notable exception to this spatial correspondence is CI-0, where F5
leads to an estimated firn air content in the upper 90 m of 5.18 m. Our results indicate a
significant proportion of firn air content is contained within continental ice (F5), an effect
that is neglected by commonly applied models of ice shelf firn air content in which firn air
thickness is contained within the uppermost densifying layers. Firn air within F5 is an
important component of firn air at inlet sites at least. However, the extent and along-flow
geometrical development of F5 is unknown. The fact that our firn air estimates (Table 2) are
generally lower than those reported by Holland et al. (2011) may indicate that significant firn
air is contained at other sites within F5 ice but is not sampled by our borehole logs. Craven et
al. (2005) reported on a 205 m thick layer of “white bubbly” continental ice located < 100 km
from the ice front at 70 m depth on Amery Ice Shelf directly above a marine ice unit.
Independent estimates of full-depth firn air (Nicholls et al., 2012) in LCIS are in agreement
with those of Holland et al. (2011). Interpreting OPTV-derived density estimates in terms of
firn air places reliance on the luminosity–density relationship of Hubbard et al. (2016). At high densities the RMSE of this relationship is 21.7 kg m$^{-3}$, an uncertainty which can contribute several metres of firn air over the ice thickness. Despite the uncertainties relating to the absolute density our results indicate that a substantial proportion of firn air is contained at depth, at least at CI-0, within “continental ice” and trapped between refrozen layers at other sites.

At each drill site we measured the ice thickness at using a 50 MHz radar, a propagation velocity of 0.166 m ns$^{-1}$ and a firn-velocity correction from a coincident common-offset survey. If the mean densities of F5 at CI-0 and WI-0 extended throughout the remainder of the ice column then 29.2 m and 14.2 m of firn air could be added to the CI-0 and WI-0 values reported in Table 3. At the remaining sites, where F5 is not observed (but may be present at depth in some form) the mean density in the bottom metre (89 – 90 m) is 901 ± 3 kg m$^{-3}$. If we assume this density extends to the base of the ice shelf at each site then a further firn air content of 6.9 m to CI-22, 3.1 m to CI-120 and 3.5 m to WI-70 would be added. However, we acknowledge that these are maximum estimates as further compression will occur and that the ice-ocean interface will likely be poorly consolidated in areas of basal accretion where some of the “firn air” is taken up by the penetration of relatively dense sea water, although we do not believe any of our sites overlie accreting areas.

To estimate the percentage of our OPTV logs composed of refrozen ice from our thresholding output we scale and sum our percentages from each facies according to their depth range. We assume that F3 (“pond ice”) and F4 ice in CI-22 both have a thresholding output percentage of 100%. This is justified because F3 pond ice is effectively a specific type of refrozen ice (Hubbard et al., 2016) and we have previously interpreted CI-22 F4 as advected, massive F3 ice due to its lack of visible structure. F4 ice in CI-120 and WI-70 maintains its structure and therefore we interpret thresholding output area as the reflecting refrozen ice percentage. A
further consideration is that, as the ice column densifies, the vertical velocity will be preferentially accounted for by the compaction of bubble-rich host ice, not the bubble-poor refrozen ice layers. Consequently, it is expected that refrozen ice will constitute a progressively larger proportion deeper in the ice column given constant levels of melt through time. In the absence of further data, we assign the upper 1.87 m, not sampled by OPTV, a thresholding output percentage equal to that of the underlying F1 facies. For 0–90 m depth this results in refrozen ice estimates of 56.5% for CI-0, 79.9% for CI-22, 41.5% for CI-120, 46.1% for WI-0 and 44.0% for CI-70.

Those inlet sites most affected by föhn warming (CI-0 and CI-22) have notably higher refrozen ice contents than those less affected by intense föhn warming (WI-0, WI-70, and CI-120). The similar refrozen ice contents of the latter three suggest that these reflect the effect of ice shelf wide seasonal warming, melting and percolation, rather than föhn-driven melting. The relatively low refrozen content of WI-0 may contrast with the conceptual framework of föhn-driven and seasonal melting combining to overprint on the ice column of LCIS. However, we note that the WI-0 refrozen ice content is inclusive of F5 which has a relative paucity of refrozen ice layers (17.3 %, Table 3). F1 and F2 ice at this site therefore accounts for a larger proportion of the total refrozen ice content here.

6. Conclusion

Herein, we have analysed a suite of five optical televiewer (OPTV) logs from Larsen C Ice Shelf, Antarctica of its northern and central portions. We have reconstructed ice shelf density based on OPTV image luminosity and the refrozen melt portion with a thresholding-based technique. The thresholding technique performs well when the ice column is composed of densifying, reflective host ice which has been overprinted by layers of refrozen melt, but less well where ice is homogenous due to compression or the refreezing of melt ponds. Refrozen
ice and remnant refrozen ice makes up >40% of the top 90 m of LCIS and thus is a major contributor to the ice shelf mass redistribution. At inland sites, close to the location of maximum föhn-warming, refrozen ice comprises up to 80% of the top 90 m of the ice column.

We bring these observations together to devise a five-part facies scheme defining: F1 as densifying ice shelf accumulation; F2 as föhn-affected densified ice; F3 as ice formed by the refreezing of föhn-driven surface melt ponds; F4 as buried and compressed föhn-affected ice and F5 as lower density continental ice formed upstream of the föhn warming region. While the boundaries between some of these facies are subjective we have taken care to define them in a reproducible, quantitative manner. Where a firn air correction is guided by a firn densification model outputs it neglects the effect of lower density, buried ice from continental sources which may form a significant proportion of the total firn air content at some sites. For example, in Cabinet Inlet (CI-0), the lower unit of continental F5 ice contains 62% of the firn air between 0 and 97.5 m depth, despite occupying the lower 46% of this depth range. Away from CI-0 the refreezing of melt water effectively traps firn air at depth and this effect should be integrated into firn densifications models used in correcting altimetric data which otherwise risk underestimating the amount of firn air within the ice column. However, if lateral discontinuities in these melt layers can be exploited then this pore space becomes available to be filled by percolating meltwater.

This work demonstrates the efficacy of hot-water drilling and OPTV borehole logging in acquiring high-resolution structural and melt information from ice shelves. OPTV logging has successfully imaged bubble structures formed during melt water refreezing, generations of layers which cut across one another formed during ice deformation and fracture close to the grounding line and refrozen ice layers preserved at depths of up to 90 m. The facies scheme outlined here could be extended in the future to include other ice shelf ice types such
as dry, cold deposited snow, marine ice or rift ice. Importantly, a growing library of ice
OPTV images exists (e.g. Hubbard et al., 2012; Hubbard and Malone, 2013) and could be
used to build a unified and widely-applicable facies scheme.

Acknowledgements and Data

Research was funded by the UK Natural Environmental Research Council grants
NE/L006707/1 and NE/L005409/1. Thanks to British Antarctic Survey logistics and field
guides for assistance during data collection. Data will be available via the project website
(www.projectmidas.org) and the relevant public research council repository. Authors declare
no conflicts of interest.

References

   Ice Shelf triggered by chain reaction drainage of supraglacial lakes. Geophysical
2. Cape, M. R., Vernet, M., Skvarca, P., Marinsek, S., Scambos, T. and Domack, E.
   (2015) Foehn winds link climate-driven warming to ice shelf evolution in Antarctica.
   Journal of Geophysical research: Atmospheres, 120, 11037-11057. doi:
   radar altimetry. Geophysical Research Letters, 42, 10,721–10,729,
   the Antarctic Peninsula over the past 50 years. The Cryosphere.4:77-98.
   doi:10.5194/tc-4-77-2010
5. Craven, M., Carsey, F., Behar, A., Matthews, J., Brand, R., Elcheikh, A., Hall, S. and
   Treverrow, A. (2005) Borehole imagery of meteoric and marine ice layers in the
   Amery Ice Shelf, East Antarctica. Journal of Glaciology, 51 (172), 75-84.
   doi:10.3189/172756505781829511


