ACCEPTED ON ANNALS OF GEOPHYSICS, 62, 2019; Doi: 10.4401/ag-7794

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Distribution of Shallow Isochronous Layers in East Antarctica Inferred from Frequency-Modulated Continuous-Wave (FMCW) Radar

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1 Subject classification:
2 02.02.03. Geomorphology;
3 02.02.04. Ice;
4 02.02.10. Instruments and techniques;
5 02.02.99. General or miscellaneous;
6 02.03.05. Paleoclimate

7 ABSTRACT
8 During the 32nd Chinese National Antarctic Research Expedition, the Frequency-Modulated Continuous-Wave (FMCW) radar was used for the first time to obtain the distribution of shallow isochronous layers within the East Antarctic region extending from Zhongshan Station to Kunlun Station. Taking a typical area as a case study, this article describes the complete workflow used in radar data processing, including signal processing and extraction of isochronous layers. The wave velocity model is established according to an empirical formula to calculate the depth of the layer, and the result is compared and corrected with the volcanic record in ice core DT263; the relative error of depth is only approximately 5%. The echograms of the 20 isochronous layers in three regions are presented, including the 21 area around the dome A, the area 100 km from the dome A and 22 the area in the Lambert Glacier. A comparison of the echograms 23 within the three regions shows that the isochronous layers are 24 relatively stable in the Dome A and change more intensely in the 25 Lambert Glacier region, while the folding of the layer occurs in a 26 concentrated area near Dome A. This folding may be due to the 27 local layer mixing and compression caused by the ice flow and 28 wind-driven processes. The analysis of the distribution of the 29 shallow isochronous layers and age-depth information from 30 different regions provides important data that support the 31 calculation of large-scale accumulation rates and flow history in 32 the Antarctic region.

33 1. Introduction
34 The density of ice formed in different periods differs within 35 glaciers [J.G. Paren and G. de. Q. Robin, 1975], and the different 36 characteristics caused by variations in density or conductivity, as 37 well as variations in crystal orientation fabric (COF), can result in 38 the detection of distinct layers (isochronous layers) from 39 electromagnetic-wave reflection signals [J.G. Paren and 40 G. de. Q. Robin, 1975; M J. Siegert, 1999; S Fujita et al., 1999; 41 DG Vaughan et al., 1999]. Based on this feature, various radar 42 systems (AWI, BAS, CRe-SIS, INGV and UTIG) have been 43 developed for the study of Antarctic ice sheets [A Winter et al., 44 2017]. Most radar systems can provide reliable detection at 45 depths of several kilometers [D. J. Drewry, 1975; Y.Y. Macheret 46 and A. B. Zhuravlev, 1982]. Radio-echo sounding (RES) is 47 widely used to detect the thickness of the Antarctic ice sheet, the 48 presence of subglacial lakes and the distributions of ice and 49 bedrock [S. Gogineni et al., 1998].

50 The Prydz Bay-Amery Ice Shelf-Dome A Project Expedition 51 and Research Plan is an important program of the Chinese 52 Antarctic expedition, and completed continuous observations of 53 multiple parameters from Prydz Bay near Zhongshan Station to 54 the highest point on Dome A (Figure 1) [Paul Andrew Mayewski 55 et al., 2005; C. Xiao et al., 2013; Hou ShuGui et al., 2007]. Along 56 this profile, ice sounding radar was used to determine several 57 large-scale glacial parameters (e.g., ice thickness and ice 58 distribution) in the area of interest. The West Antarctic Ice Sheet 59 (WAIS) Divide deep ice core project has also mapped the deep 60 internal layers in West Antarctica [CM Laird et al., 2010]. 61 However, the ability of ice sounding radar to detect the internal 62 features of the top ice sheet is limited, making it difficult to 63 observe the distribution of isochronous layers in shallow ice 64 sheets (within the ~200 m) [H. Conway et al., 2005]. Instead, 65 high-resolution ice sounding radar can be used to detect features 66 within shallow glacier ice. Le Meur et al. [2018] used ground- 67 penetrating radar to detect the East Antarctic Plateau and 68 obtained continuous isochronous layers spatial distributions 69 within ~100 m of depth between Dome Concordia (DC) and 70 Vostok Station.

71 To determine the continuous isochronous features of a shallow 72 ice sheet over longer distances, and to add the available data on 73 East Antarctic ice sheet dynamics and surface mass balance 74 (SMB) [S. Gogineni et al., 2007], the inland Kunlun team of the 75 32nd Chinese National Antarctic Research Expedition 76 (CHINARE 32) used FMCW sounding radar to fully observe the 77 ice sheet along a route extending from the inland base (8 km from 78 Zhongshan Station at the margin of the ice sheet) to Kunlun 79 Station near Dome A (Figure 1), and obtained profile data of 80 shallow isochronous layers in the East Antarctic with a total 81 distance of 1280 km from the coast to the inland area. These data 82 are of great value for the analysis the temporal and spatial 83 variability of SMB and provide a basis for reconstructing 84 Antarctic climates and deducing the effects of future climate 85 change.

86 This paper first describes the process of data acquisition and 87 processing during the CHINARE 32. To better demonstrate the 88 process by which information on the isochronous layers and their 89 depths were extracted, we employed Profile 1, as shown in 90 Figure 1, to illustrate the long-distance profile generation process. 91 We established a velocity model of the electromagnetic waves in 92 the ice sheet and calculated the depth of the layer according to the 93 two-way travel time (TWT). The results were calibrated with 94 reference to the volcanic records of the ice core (code: DT263; 95 GPS coordinates: S76°32'28", E77°01'32") drilled by CHINARE 96 19 [Cui X et al., 2009]. The age-depth information of the layer 97 was calibrated by the ice core [Jean-Robert Petit et al., 1981], 98 which can be used for large-area extraction isochronous layers to 99 calculate the accumulation rate. In addition, portions of the Dome 100 A region, a region 100 km from Dome A, and the Lambert 101 Glacier were selected to map the isochronous layers. These 102 regions are illustrated in Figure 1 as Profiles 2, 3, and 4, 103 respectively. By analyzing the echograms of the long-distance 104 isochronous layers in these regions, they exhibit completely 105 different characteristics, which are closely related to the 106 accumulation rate between regions. The data presented in this 107 paper facilitate calculations of large-scale accumulation rate and 108 can be used to analyze the temporal and spatial distribution of 109 SMB in the 1280 km range (from Zhongshan Station to Dome A). 110 These data also provide the foundation for reconstructing past 111 climates in the Antarctic region and analyzing the flow history.
Figure 1. The location of the radar detection trajectory extending from the base to Kunlun Station. The locations of four typical areas (Profile 1 to Profile 4), the ice core and nearby stations are shown.

2. Materials and methods

Considering the need for precise detection of shallow isochronous layers and the lack of a large amount of convenient energy, FMCW sounding radar was used in this work. Compared to sounding radars with a depth of several kilometers, the FMCW sounding radar sweep bandwidth used in this study was 1.5 GHz, and the period was 4 ms, resulting in a high vertical resolution of 6.16 cm in firn/ice. The trace repetition rate was 100 traces/s; each trace contained 24,576 points at a sampling rate of 6.25 MS/s and was able to measure the isochronal layer within 10-100 m. The system power was 2 W, but due to power limitations, only the rising component of the radar signal was used [A. G. Stove, 1992; C Leuschen et al., 2014; S Urbini et al., 2017].

The sounding radar system was mounted on the trunk of a 130 PistenBully 300 Polar snowmobile and collected data at a speed of 14 km/h. The main control box was installed inside the trunk, and the antennas were affixed to the top of the snowmobile; its small size provided portability. The distance between the transmitting and receiving antennas exceeded 2 m [N. Galin et al., 2012]. Because a metal rod would interfere with the signal and could cause accidents when violent bumps occur, a bamboo pole was used to secure the antenna (Figure 2). The location of the radar traces was obtained by a high-precision global positioning system (GPS) mounted on other vehicles and was sampled synchronously with the FMCW sounding radar. The inland Kunlun team of CHINARE 32 departed from the inland base on December 15, 2015, and arrived at Kunlun Station on December 30.

3. Signal processing

The main purpose of this stage is to optimize the quality of the data and prepare for the subsequent layer mapping. The specific process is shown in Figure 3.
The raw data contain considerable coherent noise when they are collected. The presence of this noise will affect the extraction and analysis of ice structure, making the echogram blurry and unable to effectively show the isochronal layer. Taking into account the signal-to-noise ratio, only the last of every 8 traces received by the radar was preserved to suppress coherent noise [S Urbini et al., 2010].

The original data were stored in the time domain; a fast Fourier transformation (FFT) was applied to the original data to obtain a spectrum plot, and a Hanning window was applied to the data to reduce the range sidelobes. Because there is a 1/e^2 reduction in power with range, we applied a lowpass filter to the spectral data, and a filter with a very low cutoff frequency was used. The original spectra were then multiplied by the inverse of the lowpass filter response to correct for the gain. The median filter approach filters out multiplicative noise and maintains the sharpness and smoothness of the edge regions of the layer [KM Bakwad et al., 2009].

On the surface of the detection area, the maximum interference in the radar signal is caused by direct waves [QZ Bao et al., 2014]. A direct wave is generated when the transmitting antenna propagates a signal through the wire to the receiving antenna, and electromagnetic waves are reflected when passing through the surface of the ice sheet. According to the FMCW ranging principle (the distance is proportional to the frequency), the frequency of a direct wave is lower than the normal radar reflected signal; a highpass filter was constructed to filter out the short direct waves so that the fundamental harmonic of the direct waves was completely filtered.

After the above processing, the curvelet transform must be applied to attenuate background and random noise [E Uslu et al., 2014]. According to previous studies [S Lang, Liu, X. et al., 2015; S Lang, Zhao, B. et al., 2015], the curvelet transform was established to extract the features of the layer. This step allowed us to effectively distinguish isochronous layers. Figure 4 shows that the distribution of the layers is more obvious after this intensive processing.

After a clear image was obtained, the distance and depth of this echogram could be calculated. When calculating the distance of 191 the profile, according to the principle of sounding radar and referring to the radar parameters in Section 2, the distance $S$ of the profile can be accurately calculated from the number of traces $N$:

$$S = v \times \frac{T_{\text{swp}}}{N}$$

(1)

where $v$ is the horizontal movement speed of the sounding radar; $T_{\text{swp}}$ a snowmobile can carry the radar at a constant speed of 14 km/h. $T_{\text{swp}}$ is the sweep period of 10 ms. When calculating the depth, to complete the transition between 200 travel time and depth, it is necessary to refer to the speed of 201 electromagnetic waves in the ice sheet, which is related to the 202 dielectric permittivity and density of firm/ice. Most of the current 203 research uses a constant speed for calculations [J A Uribe et al., 204 2014]. This paper establishes a wave velocity profile model by 205 estimating the firm/ice density.

According to the basic formula:

$$c = \frac{c_0}{\sqrt{\varepsilon}}$$

(2)

where $c_0$ is the velocity of electromagnetic waves in a vacuum; $\varepsilon$ thus, the wave velocity is closely related to the relative actual 209 dielectric constant of the ice. A Kovacs et al. [1995] presented an empirical formula for the relative true dielectric constant $\varepsilon$:

$$\varepsilon = (1 + 0.845\rho)^2$$

(3)

where $\rho$ is the density value of pure ice (0.917 g/cm³) at -20 °C and $\rho_s$ is the snow density on the surface of the ice sheet. The 223 surface density of ice sheets at ice cores LGB69 and DT263 is 224 0.39 g/cm³ [J Ren et al., 2001]. The area in Figure 4 is located 225 near DT263, so this value is chosen as the value of $\rho_s$ in this 226 calculation. D Minghu et al. [2011] characterized the snow 227 surface density of the East Antarctic from the coast to Dome A. 228 The present study area is consistent with this previous paper, so 229 this value can be corrected according to the geographical position.

$C$ is a constant, which is defined as $C = 1.9/f_2$, where $f_2$ is the 231 critical depth at which snow is converted to ice. J Ren et al. [2001] 232 selected four different locations in the area from Zhongshan 233 Station to Dome A spanning approximately 1100 km and drilled 234 five 50-100 m firn/ice cores at these locations, similar to the area 235 of the present study. According to the density of the five ice cores, 236 the depth at which firm is converted to ice is approximately 60 m; 237 therefore, the value of $f_2$ in this paper is 60 m. The parameters in 238 the formula are adjusted according to the local environment, and 239 the accuracy of the calculation result can be improved [M 240 Fahnestock et al., 2001]. The correlation between the empirical 241 formula and ice core DT263 regarding the depth-density 242 relationship is 0.94.

According to Equations (2)-(4), we can derive the functional 244 relationship between the wave velocity and the depth and use it to 245 perform the conversion between travel time and depth. After
post-processing of the raw data, the profile shown in Figure 4 was obtained, which shows the lateral variation characteristics within the range of 5 km.

4. Mapping of isochronous layers

After signal processing, the information of different layers can be distinguished in the echogram. In this step, we mapped an isochronous layer with a length of 5.8 km and arbitrarily selected several clearly defined isochronous layers for tracking and extraction.

In the subsequent accumulation rate study, the accumulation rate model can be used to analyze the effect of depth on the SMB. [1 Das et al., 2015]. In Figure 4, the continuity of the layers is better, and the depth of the layers is characterized by greater depth in the middle of the echogram and relatively shallower on both sides. According to the standard SMB equation,

\[
SMB = zp \alpha^{-1}
\]  

where \( z \) is the depth of the layer, the density \( p \) of the layer can be calculated according to Equation (4), and \( \alpha \) is the age of the layer. The depth of the isochronous layer is proportional to the accumulation rate, so the accumulation rate in the middle of this area is higher. Figure 5 corresponds to the elevation in the echogram. Since the length of this area is only 5.8 km and the surface altitude change is not severe, the influence of the terrain can be observed. We need to analyze the distribution of the layers on a larger scale to determine whether the difference in the rate of accumulation is ubiquitous. And this finding proves that the FMCW sounding radar can help us study the differences in the spatial and temporal distributions of accumulation rate.

By connecting all of the cross sections, as shown in Figure 5, the distribution of shallow isochronous layers from Zhongshan Station to Kunlun Station in East Antarctica can be obtained. We chose an echogram of the isochronous layers over a distance of 30 km in the Lambert Glacier area as an example and selected 5 isochronous layers for tracking and extraction (Figure 6). However, when we added the elevation in an area of 70 km (Figure 7), the results were not satisfactory because the terrain changes usually exceed 200 m in a large area (more than 100 km), whereas the shallow ice sounding radar can measure isochronous layers only within a range of 100 m, and the elevation change far exceeded the depth that FMCW radar can detect. To clearly observe the distribution characteristics of the isochronous layer and calculate the SMB according to the depth (Equation (5)), this article, is similar to Robert et al. [1993], which employ corresponding elevation information only when analyzing individual regions.

5. Correspondence with ice core data

To accurately calculate the SMB, the year information recorded in the ice core can be used. Zhou et al. [2006] determined the year in which the isochronous layers were deposited by comparing the trace elements produced by volcanic eruptions with the composition of the ice core. According to the non-sea-salt \( \text{SO}_4^{2-} \) level contained in the ice core, records of at least 17 volcanic eruptions are present in this core. For example, trace elements produced by the eruption of Krakatoa in Indonesia, which occurred in AD1884, appear at a depth of 30.3 m in the ice core. Thus, layer 1 (Figure 6) represents the location of the ice sheet surface in AD1884, but it is now located at a depth of 30.3 m, which can be interpreted as the material that has accumulated since AD1884. This accumulation may have been driven by snowfall, wind transport or other factors [HP Black, et al., 1964].

Thus, the isochronous layer can be defined in terms of its depth at the site where the isochronous layer that formed in AD1884. In addition, the same isochronous layer is understood to have appeared at the same time in other areas, even if the depth of the layer is variable. Such variations in depth arise because the accumulation rates between the regions are different, causing fluctuations in the layers. Similarly, the ice core data show that the eruption of Cosiguina in Nicaragua, which occurred in 1982 AD1835, is recorded at a depth of 36.2 m, and the corresponding isochronous layer 2 (Figure 6) can be related to a particular time in the past.
Figure 6. Radar data in the range of 30 km near ice core DT263. (a) Echogram obtained by signal processing and drawing isochronous layers; the black line shows the location of ice core DT263; the red point A shows the trace of the Krakatoa eruption of Indonesia in AD1884, with a depth of 30.3 m; the blue point B shows the traces of the Cosiguina volcano eruption that occurred in Nicaragua at AD1835, with a depth of 36.188 m; the green point C shows an unknown volcanic record from AD1285, with a depth of 72.403 m. A partial cross-sectional view near the yellow point C (the black box) is enlarged to obtain (b), where the four lines represent four adjacent high-density isochronous layers in Table 1. (c) Extraction of three isochronous layers drawn in (a).

| Volcanic events       | D13 | D14 | D15 | D16 |
|-----------------------|-----|-----|-----|-----|
| Year of eruption (A.D.) | 1285 | Unknown | Unknown | 1259 |
| Time in core (A.D.)   | 1286 | 1277 | 1269 | 1260 |
| Depth in core (m)     | 72.403 | 73.266 | 74.079 | 75.233 |
| Depth of volcanic record (m) | 74.173 | 75.035 | 75.848 | 77.002 |
| TWT in RES (ns)       | 745.9 | 757.9 | 770.8 | 778.2 |
| Depth in RES (m)      | 70.06 | 70.97 | 72.20 | 72.90 |
| Relative error        | 5.5% | 5.41% | 4.81% | 5.3% |

Table 1. Comparison of radar data with volcanic eruption records using ice core DT263.
337 the propagation velocity error of electromagnetic waves in 338 firn/ice also increases.
339 We measured the area where ice core DT263 was located on 340 December 24, 2015 [R. Drews, O. Eisen1, I. Weikusat, 2009], but 341 the ice core had been drilled by the inland team in AD1999, 342 during which snowfall increased the depth of the volcanic records 343 contained in the ice core. According to the SMB data from stakes 344 observation, the average accumulation rate in the region 860 km 345 from the coast from AD1999 to AD2015 was 55.35 kg/m²·a⁻¹. 346 and exhibited no temporal changes [Ding et al. 2011]. The 347 thickness of the snow added in this area was estimated to be 348 1.76 m.
349 In Figure 6a, layer 4 is selected to perform error analysis on the 350 wave velocity model. Four unknown volcanic eruption records 351 were found at depths of 72.403 m, 73.266 m, 74.079 m, and 352 75.233 m of ice core DT263 [Y Li et al., 2009]. Together with the 353 increased snow thickness during AD1999-2015, radar data can be 354 calibrated based on these volcanic records. Four closely visible 355 radar reflection layers can be found in the TWT range of 356 740-780 ns (Figure 6b). The parameters of these four isochronous 357 layers are shown in Table 1, which shows that the relative error 358 between the depth of D13-D16 and the volcanic record in the ice 359 core is approximately 5%, which is consistent with the findings 360 of M Frezzotti et al. [2005].
361 The difference between the two data sets is due in part to the 362 increased snow thickness error within AD1999-2015. The firm 363 formed during this period has not yet reached the critical density 364 (0.55 g/cm³), and the process of snow densification is obvious in 365 this range [Herron M M et al. 1980]. Therefore, the present paper 366 has a certain degree of overestimation of the inversion results. In 367 addition, although Equation (4) is an often useful empirical 368 formula for the depth-density relationship and has a high 369 correlation with the measured data of DT263, it still contains the 370 biases.
371 However, when using a constant wave velocity of 0.168 m/ns 372 (the electromagnetic wave velocity in pure ice), the relative error 373 reached 15.5%, which shows that the velocity model used in this 374 paper has higher precision when calculating depth. If the model is 375 used in a deeper range, it will exhibit greater advantages, which 376 provides a foundation for accurate calculations of accumulation 377 rate.
378 The depth of internal reflection horizons can be calibrated by 379 the corresponding ice core, and the depth-year information in the 380 vicinity of the ice core can also be determined. As shown in 381 Figure 7, the isochronous layer in the region is extended, and 382 continuous isochronous layers within a few hundred kilometers 383 can be obtained so that the SMB can be studied on a larger scale. 384 This method can also be used to plot the three-dimensional 385 distribution of a particular isochronous layer after multiple 386 investigations [MJ Siegert et al., 2010].

387 6. Presentation of isochronous layers over large areas

388 At the end of this paper, a number of typical areas are also 389 analyzed, including the Dome A area, a region 100 km away from 390 Dome A [Sun Bo et al., 2009], and the region surrounding the 391 Lambert Glacier, which is the largest glacier in the world. These 392 regions present a wide range of layers.
393 In Figure 7, the randomly selected images of different depths 394 and layers in these regions show the shallow ice structure and the 395 depth variation in the horizontal direction within ~130 m. The 396 depth information of the isochronous layers is shown in Figure 7. 397 Therefore, we can, as in the Section 5, track and map specific 398 isochronous layers using the corresponding ice core data and 399 determine the ages corresponding to specific depths.
400 The isochronous layers of Dome A and the Lambert Glacier are 401 continuous and can be traced over large areas. Analysis of the 402 isochronous layers in these regions shows that the elevation 403 changes are relatively small in the Dome A region, and the 404 structure of the shallow isochronous layers also shows excellent 405 stability. The Lambert Glacier area displays relatively stable 406 elevations; however, the isochronous layers are more volatile, 407 although they are still relatively continuous and the layers are 408 folded in local areas.
409 Through analysis of the relief (depth variation) of the 410 isochronous layers shown in Figure 7 (a) and (b), the maximum 411 depth, the minimum depth, the average depth and the standard 412 deviation of the depths of the different layers are obtained 413 (Table 2, Table 3). The standard deviation of the isochronous 414 layers’ depth within the Lambert Glacier region is more than 3 415 times that of the Dome A. This result shows that the isochronous 416 layers are relatively flat in the Dome A, whereas the changes in 417 the Lambert Glacier region are more pronounced.

| Layer   | Layer 1 | Layer 2 | Layer 3 | Layer 4 |
|---------|---------|---------|---------|---------|
| Minimum depth (m) | 22.7 | 31.0 | 38.8 | 51.5 |
| Maximum depth (m) | 41.6 | 54.5 | 69.0 | 90.6 |
| Average depth (m) | 33.1 | 43.0 | 54.9 | 70.8 |
| Standard deviation (m) | 3.4 | 4.4 | 5.6 | 9.4 |

Table 2. Information on the depth of layers in the Dome A

| Layers   | Layer 1 | Layer 2 | Layer 3 |
|----------|---------|---------|---------|
| Minimum depth (m) | 39.3 | 51.6 | 56.9 |
| Maximum depth (m) | 67.1 | 86.1 | 94.9 |
| Average depth (m) | 46.6 | 57.6 | 64.1 |
| Standard deviation (m) | 12.5 | 14.8 | 16.4 |

Table 3. Information on the depths of layers on the Great Icefield of the 421 Lambert Glacier
In addition, the sharply shrinking layer (contraction zone) cannot be tracked. It can be found that in the area around Dome A, where the detection length is 84 km, we cannot effectively trace the distribution of the isochronous layers due to the large number of shrinkage areas. In the area where Profile 3 (Figure 1) is located, the elevation increases more rapidly, and a phenomenon in which multiple isochronous layers converge is frequently observed, thus impeding the continuity of the isochronous layers.

However, during the extraction process, a sharper slope along the internal layer (the area shown in Figure 8 and Profile 3 in Figure 1) was produced. Since the layers are compressed, making it impossible to clearly distinguish the periods of the layers and affect the calculation of the depth-year relationship [J. J. Legarsky and X. Gao, 2006].

The shrinkage of the layer will be concentrated in certain areas, and there are many explanations for the appearance of this phenomenon. For example, after the ice flows into different regions, the tributary and the main stream are concentrated by the lateral compression of the mechanical anisotropic ice [PD Bons et al., 2016]; the fold produced by the ice flowing through the basal topography of the onset region is preserved [King, E., 2011]. It is also possible that the interaction of topography, altitude and the wind field impacts the accumulation rate [S Fujita et al., 2011].

Analysis of all radar data showed that this shrinkage phenomenon occurs along the coast and in individual inland areas. The folding of the layer appears multiple times on the 80 km scale, as shown in Figure 8. According to the data reported by Z Shengkai et al. [2008], the ice flow velocity in the Profile 3 region gradually increases, and the flow direction substantially matches the surface elevation profile; therefore, it is likely to represent local layer mixing and compression caused by the ice flow. The surface topography of Profile 3 is very complex and has high spatial variability [D Minghu et al., 2011]. Wind-driven processes also have a great influence on the distribution of accumulation rate.

After analyzing these three regions, it can be found that the distribution of isochronous layers differs among regions. These phenomena are closely related to the spatial transformation of the accumulation rate. Therefore, further analyses can be conducted using geographic and meteorological factors.

Figure 7. Presentation of the isochronous layers over large-scale areas; (a) a display of the Great Icefield of the Lambert Glacier; (b) a display of the Dome A.

Figure 8. A display of the area 100 km from Dome A and the corresponding elevation. The location at which the layers are folded is often accompanied by dramatic changes in elevation.
7. Conclusion

This paper mainly describes the shallow exploration radar observation work conducted by CHINARE 32 from Zhongshan Station to Kunlun Station in 2015. The process of collecting data for the entire section is described, detailed processing was then carried out, the isochronous layers were drawn, and the correspondence of the results with the ice core were demonstrated. The results are summarized below.

1) FMCW sounding radar is a high-resolution radar system. After signal processing, a clear echogram of the isochronous layers can be obtained. And layers within \( \sim 100 \) m can be distinguished and extracted.

2) To calculate the SMB by isochronous layer, the depth of each layer needs to be accurately calibrated, but the density of the firm/ice changes significantly, resulting in a large change in the velocity of the electromagnetic waves in the shallow region. Therefore, according to the empirical formula, the velocity model of electromagnetic waves in firm/ice was established to calculate the depth of the layer. Based on the volcanic records stored in ice core DT263, the error analysis and calibration of the depth showed the superiority of the model results.

3) During the entire exploration transect, the elevation changes by more than 4000 m; therefore, the elevation of the transect route cannot be ignored. However, the elevation can only be matched in a small range. When the elevation change is much greater than the detection depth, it is difficult to observe the distribution of shallow isochronous layers.

4) Isochronous layers of different depths sometimes converge and can still be tracked in constricted areas. This folding phenomenon occurs along the coast and in individual inland areas, and many explanations exist for the appearance of this phenomenon. However, an analysis of radar operations indicates that the constricted regions are not a result of radar instability [A. M. Ilisei and L. Bruzzone, 2014]. This phenomenon is probably due to the folding and mixing of different layers caused by ice flow movement and wind-driven processes.

5) As seen from Table 2 and Table 3, there is a difference in the standard deviations of the layer depth in the Dome A and the Lambert Glacier. However, the isochronous layers cannot even be effectively extracted in the region 100 km away from Dome A. It can be found that the distribution of isochronous layers differs among regions. Meanwhile, it can be speculated that among these regions, the accumulation rates differ, and the reasons for these phenomena deserve additional analysis [T.T. Wang and B. Sun, 2016].

Acknowledgements. This work was supported by the National Natural Science Foundation of China (41776199) and was carried out during the 32nd China Antarctic Scientific Expedition. Thanks to Yuzhong ZHANG and Xueyuan TANG for providing us with the radar principle and data processing help. Taiyuan University of Technology and the Polar Research Institute of China provided logistical support.

References

222 J.G. Paren and G. de. Q. Robin (1975). Internal Reflections in Polar Ice Sheets, Journal of Glaciology, 14 (71), 251-259.

225 M J. Siegert (1999). On the Origin, Nature and Uses of Antarctic Ice-sheet Radio-Echo Layering, Progress in Physical Geography, 23 (2), 159-179.

228 S Fujita, H Maeno, S Uratsuka, T Furukawa, S Mae (1999). Nature of Radio Echo Layering in the Antarctic Ice Sheet Detected by a Two-
604 cascaded with adaptive filter to enhance peak signal to noise ratio
605 from single image. IETE Journal of Research, 55(3), 173-179.
606 Bao, Q. Z., Li, Q. C., & Chen, W. C. (2014). GPR data noise attenuation
607 on the curvelet transform. Applied Geophysics, 11(3), 301-310.
608 Uslu, E., & Albayrak, S. (2014). Curvelet-based synthetic aperture radar
609 image classification. IEEE Geoscience and Remote Sensing Letters,
610 11(6), 1071-1075.
611 Lang, S., Liu, X., Zhao, B., Chen, X., & Fang, G. (2015). Focused
612 synthetic aperture radar processing of ice-sounding data collected
613 over the east antarctic ice sheet via the modified range migration
614 algorithm using curvelets. IEEE Transactions on Geoscience and
615 Remote Sensing, 53(8), 4496-4509.
616 Lang, S., Zhao, B., Liu, X., & Fang, G. (2015). Two-dimensional imaging
617 of ice sheets of airborne radar sounder via a combined modified
618 range migration algorithm based on ISFT and beamforming using
619 curvelet. IEEE Journal of Selected Topics in Applied Earth
620 Observations and Remote Sensing, 8(1), 76-89.
621 Uribe, J. A., Zamora, R., Gacitúa, G., Rivera, A., & Ulloa, D. (2014). A
622 low power consumption radar system for measuring ice thickness
623 and snow/firn accumulation in Antarctica. Annals of Glaciology,
624 55(67), 39-48.
625 Kovacs, A., Gow, A. J., & Morey, R. M. (1995). The in-situ dielectric
626 constant of polar firn revisited. Cold Regions Science and
627 Technology, 23(3), 245-256.
628 Cuffey, K. M., & Paterson, W. S. B. (2010). The physics of glaciers.
629 Academic Press.
630 Ren, J., Qin, D., & Xiao, C. (2001). Preliminary results of the inland
631 expeditions along a transect from the Zhongshan Station to Dome A,
632 East Antarctica. J. Glaciol. Geocryol, 23(1), 51-56.
633 Fahnestock, M., Abdalati, W., Luo, S., & Gogineni, S. (2001). Internal
634 layer tracing and age - depth - accumulation relationships for the
635 northern greenland ice sheet. Journal of Geophysical Research:
636 Atmospheres, 106(D24), 33789-33797.
637 Das, I., Scambos, T. A., Koenig, L. S., Broeke, M. R., & Lenaerts, J.
638 (2015). Extreme wind - ice interaction over Recovery Ice Stream,
639 East Antarctica. Geophysical Research Letters, 42(19), 8064-8071.
640 Frezzotti, M., Pourni, M., Flora, O., Gandolfi, S., Gay, M., Urbini, S., ...
641 & Severi, M. (2005). Spatial and temporal variability of snow
642 accumulation in East Antarctica from traverse data. Journal of
643 Glaciology, 51(172), 113-124.
644 Robert, W. J., Anthony, M. G., David, L. G., Steven, M. H., & David, L.
645 W. (1993). Interpretation of radar-detected internal layer folding in
646 West Antarctic ice streams. Journal of Glaciology, 39(133), 528-537.
647 Fujita S, Holmlund P, Andersson I, et al. (2011). Spatial and temporal
648 variability of snow accumulation rate on the East Antarctic ice
649 divide between Dome Fuji and EPICA DML, The Cryosphere, 5
650 (4), 1057-1081.
651 Zhou, L., Li, Y., Jihong, C. D., Tan, D., Sun, B., & Ren, J., et al.
652 (2006). A 780-year record of explosive volcanism from d263 ice
653 core in east antarctica. Chinese Science Bulletin, 51(22), 2771-2780.
654 Black, H. P., & Budd, W. (1964). Accumulation in the region of Wilkes,
655 Wilkes Land, Antarctica. Journal of Glaciology, 5(37), 3-15.
656 R. Drews, O. Eisen1, I. Weikusat (2009). Layer disturbances and the
657 radio-echo free zone in ice sheets, The Cryosphere, 3 (2), 195-203.
658 Minghu D, Cunde X, Yuansheng L, et al. Spatial variability of surface
659 mass balance along a traverse route from Zhongshan station to
660 Dome A, Antarctic[J]. Journal of Glaciology, 2011, 57(204): 658
660-666.
661 Li, Y., Cole - Dai, J., & Zhou, L. (2009). Glaciochemical evidence in an
662 East Antarctica ice core of a recent (AD 1450-1850) neoglacial
663 episode. Journal of Geophysical Research: Atmospheres, 114(D8).
664 Herron M M, Langway C C. Firn densification: an empirical model[J].
665 Journal of Glaciology, 1980, 25(93): 373-385.
666 MJ Siegert, M Pakar, JA Dowdeswol, T Benham (2010). Radio-echo
667 layering in West Antarctica: a spreadsheet dataset. 30 (12) :1583-
668 1591.
669 Sun Bo, Martin J. Siegert, Simon M. Mudd, David Sugden, Shuiji Fujita,
670 Cui Xiangbin (2009). The Gamburtsev mountains and the origin and
671 early evolution of the Antarctic Ice Sheet, Nature, 459 (7247) :690-3.
672 J. J. Legarsky and X. Gao (2006). Internal Layer Tracing and Age-Depth
673 Relationship From the Ice Divide Toward Jakobshavn, Greenland,
674 IEEE Geoscience and Remote Sensing Letters, 3 (4), 471-475.
675 Bons, P. D., Jansen, D., Munchel, F., Bauer, C. C., Binder, T., Eisen, O.,...
676 & Weikusat, I. (2016). Converging flow and anistropy cause large-
677 scale folding in Greenland's ice sheet. Nature communications, 7,
678 11427.
679 King, E. (2011). Ice stream or not? Radio-echo sounding of Carlson Inlet,
680 West Antarctica. The Cryosphere, 5(4), 907-916.
681 Fujita, S., Holmlund, P., Andersson, I., Brown, I., Enomoto, H., Fujii, Y
682 & Hara, K. (2011). Spatial and temporal variability of snow
683 accumulation rate on the East Antarctic ice divide between Dome
684 Fuji and EPICA DML. The Cryosphere, 5, 1057-1081.
685 Shengkai Z., Dongchen E., Zemin W., Yuansheng L., Bo, J., &
686 Chunxia, Z. (2008). Ice velocity from static GPS observations along
687 the transect from Zhongshan station to Dome A, East Antarctica.
688 Annals of Glaciology, 48, 113-118.
689 Minghu, D., Cunde, X., Yuansheng, L., Jiawen, R., Shugui, H., Bo, J.,
690 & Bo, S. (2011). Spatial variability of surface mass balance along a
691 traverse route from Zhongshan station to Dome A, Antarctica.
692 Journal of Glaciology, 57(204), 659-666.
693 A. M. Bisiel and L. Bruzzone (2014). A Model-Based Technique for the
694 Automatic Detection of Earth Continental Ice Subsurface Targets in
695 Radar Sounder Data, IEEE Geoscience and Remote Sensing Letters,
696 11 (11) :1911-1915
697 T.T. Wang and B. Sun (2016). Spatio-temporal variability of past
698 accumulation rates inferred from isochronous layers at Dome A,
699 East Antarctica, Annals of glaciology, 57 (73) :87-93.
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| Volcanic events | D13          | D14          | D15          | D16          |
|-----------------|--------------|--------------|--------------|--------------|
| Year of eruption (A.D.) | 1285         | Unknown      | Unknown      | 1259         |
| Time in core (A.D.) | 1286         | 1277         | 1269         | 1260         |
| Depth in core (m) | 72.403       | 73.266       | 74.079       | 75.233       |
| Depth of volcanic record (m) | 74.173       | 75.035       | 75.848       | 77.002       |
| TWT in RES (ns)  | 745.9        | 757.9        | 770.8        | 778.2        |
| Depth in RES (m) | 70.06        | 70.97        | 72.20        | 72.90        |
| Relative error   | 5.5%         | 5.41%        | 4.81%        | 5.3%         |

*Table 1. Comparison of radar data with volcanic eruption records using ice core DT263*
Table 2. Information on the depth of layers in the Dome A

| Layer     | Layer 1 | Layer 2 | Layer 3 | Layer 4 |
|-----------|---------|---------|---------|---------|
| Minimum depth (m) | 22.7    | 31.0    | 38.8    | 51.5    |
| Maximum depth (m)  | 41.6    | 54.5    | 69.0    | 90.6    |
| Average depth (m)   | 33.1    | 43.0    | 54.9    | 70.8    |
| Standard deviation (m) | 3.4    | 4.4    | 5.6    | 9.4    |
Table 3. Information on the depths of layers on the Great Icefield of the Lambert Glacier

| Layers          | Layer 1 | Layer 2 | Layer 3 |
|-----------------|---------|---------|---------|
| Minimum depth (m) | 39.3    | 51.6    | 56.9    |
| Maximum depth (m) | 67.1    | 86.1    | 94.9    |
| Average depth (m) | 46.6    | 57.6    | 64.1    |
| Standard deviation (m) | 12.5    | 14.8    | 16.4    |
| Figure | Description |
|--------|-------------|
| 1      | The location of the radar detection trajectory extending from the base to Kunlun Station. The locations of four typical areas (Profile 1 to Profile 4), an ice core and nearby stations are shown. |
| 2      | Sounding radar system and antenna arrangements on the snowmobile. |
| 3      | Preprocessing flow chart. |
| 4      | A visualization of the data after application of the curvelet transform. |
| 5      | Echogram of the isochronous layers generated after application of the corresponding relative elevations in the data. |
| 6      | Radar data in the range of 30 km near ice core DT263. (a) Echogram obtained by signal processing and drawing isochronous layers; the black line shows the location of ice core DT263; the red point A shows the trace of the Krakatoa eruption in Indonesia in AD1884, with a depth of 30.3 m; the blue point B shows the traces of the Cosiguina volcano eruption that occurred in Nicaragua in AD1835, with a depth of 36.188 m; the green point C shows an unknown volcanic record from AD1285, with a depth of 72.403 m. A partial cross-sectional view near the yellow point C (the black box) is enlarged to obtain (b), where the four lines represent four adjacent high-density isochronous layers in Table 1. (c) Extraction of three isochronous layers drawn in (a). |
| 7      | Presentation of the isochronous layers over large-scale areas; (a) a display of the Great Icefield of the Lambert Glacier; (b) a display of the Dome A. |
| 8      | A display of the area 100 km from Dome A and the corresponding elevation. The location at which the layers are folded is often accompanied by dramatic changes in elevation. |
Figure 1
Raw data

Remove coherent noise

FFT

Hanning window

Lowpass filtering

Median filtering

Remove direct waves

Curvelet

Final image

Figure 3
Figure 5
Figure 8