Variability in the ice sheet elevations over Antarctica derived from repetitive SARAL/AltiKa radar altimeter data (2013–2016)

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Changes in surface elevations of polar ice sheets are the result of changes in ice dynamics and surface mass balance. Here, we present intra- and inter-annual elevation changes over the Antarctic ice sheet using the AltiKa radar altimeter's 40 Hz geophysical data record products for the period 2013–2016. Slope corrections were applied on the elevations using a digital elevation model (DEM) available from NASA's ice, cloud and land elevation satellite (ICESat). Comparison of elevations from AltiKa and ICESat's DEM yielded correlation, bias and root-mean-square-deviation values of the order of 0.99, -2.88 and 23.04 m, respectively, indicating the first-level accuracy of a former dataset. Further comparison of Airborne Topographic Mapper dataset with AltiKa derived elevation yielded 0.4 m root-mean-square-deviation over a part of Vostok subglacial lake. The intra-annual change indicates that for GY2 (glaciological year), GY3 and GY4, number of negative elevation change points exceeded the number of positive elevation change points during the Antarctic austral summer period (December–February). Inter-annual elevation changes were negative during 2013–2014 and positive during 2014–2015 over east Antarctica, whereas in west Antarctica negative elevation changes were observed for both periods.

Keywords. Antarctica; ice sheet; elevation change; SARAL/AltiKa.

1. Introduction

Study of changes in ice sheet elevations is crucial for understanding the impact of global climatic variations over the polar regions. Changes in surface elevations represent the response of an ice sheet to changes in ice dynamics and surface mass balance in an integrative way (Helm *et al.* 2014). Deriving the rate of change in elevations (dh/dt) alone does not provide a total solution to the mass balance of ice sheets but it certainly supplements ice studies leading to mass balance

estimates considering additional information such as firn densification and accumulation (Zwally *et al.* 2011; Shepherd *et al.* 2012). Over large continents, such studies are only possible with the data acquired by satellite-based instruments with repetitive coverage and precision. Altimetry is one of the remote sensing methods that can be utilized for such studies over polar ice sheets. Both laser (better accuracy, cannot penetrate through cloud) and radar altimeters (relatively poor accuracy, can penetrate through cloud) have been widely used in a complementary manner or independently for

understanding ice sheet dynamics (Pritchard *et al.* 2009; Smith *et al.* 2009; Sorensen *et al.* 2011; Ewert *et al.* 2012).

Principally, the altimeter is a nadir-looking radar, which transmits a narrow microwave pulse and receives a backscattered signal. The graph of received power *vs.* two-way travel time gives the basic altimetry signature referred to as a ‘waveform’, from which all the other parameters are derived. Different parameters can be retrieved from a waveform-like epoch at mid-height, backscatter coefficient (σ_0), leading-edge amplitude and trailing-edge slope. Altimetry was originally developed for oceanic studies; later, with the advancement of technology, its potential for ice studies was explored. However, altimeter data processing over non-oceanic surfaces is complicated because of the surface topography and dielectric constant (Bamber 1994). Surface undulations divert the leading edge of the ice sheet return waveform from the on-board altimeter tracking gate, causing an error in the telemeter range measurement which leads to inaccurate measurement of the elevation. Thus, for accurate surface elevation measurement, all altimeter data require post-processing referred to as ‘retracking’ (Bindschadler *et al.* 1989; Rosmorduc *et al.* 2011). More details of the fundamentals of altimetry and its utilization for ice sheet observation are provided in the handbook (Fu and Cazenave 2001).

Utilizing the capability of altimetry, Brooks *et al.* (1978) generated surface topography over one-third of the southern Greenland ice sheet from the radar altimeter onboard the GEOS-3 (Ku-band) satellite launched in 1975. Subsequently, ice sheet topographies of 1 m accuracy were derived from Seasat and Geosat altimeter data (Zwally *et al.* 1983, 1987). Remy *et al.* (1989) developed a three-step inversion technique to generate an elevation map up to 0.40 m accuracy over Terre Adélie (a small region of the Antarctic ice sheet) and predicted 0.35 m accuracy for elevation obtained from ERS-1 altimeter data. A day’s elevations with single point accuracy of 0.10–0.15 m are now available from the ice, cloud and land elevation satellite (ICESat; laser altimeter) (Shuman *et al.* 2006). Several digital elevation models (DEMs) have been published during the last two decades in a continuing quest to obtain more accurate topography of polar ice sheets (Liu *et al.* 2001; Bamber *et al.* 2009, 2013; Fretwell *et al.* 2013; Howat *et al.* 2014).

In order to derive elevation changes over the ice sheet, two main approaches, the crossover method

(Wingham *et al.* 1998; Davis and Ferguson 2004) and the repeat track method (Legresy *et al.* 2006), were used (Xiaoli *et al.* 2016). Davis and Ferguson 2004 used a crossover method and observed a strong negative trend in elevation change of west Antarctic glacier outlets consistent with increased basal melting at glacier grounding lines caused by ocean thermal forcing. Helm *et al.* (2014) derived new DEMs, elevation change maps (using the repeat track method) and volume change estimates for both the Antarctic and Greenland ice sheets from CryoSat-2 (Ku-band) data from January 2011 to 2014. These elevation changes, when compared to ICESat (2003–2009) data, revealed that in west Antarctica volume loss has increased by a factor of 3, which is partly compensated by anomalous thickening in Dronning Maud Land, east Antarctica. ICESat data acquired between 2003 and 2008 over the Antarctic ice sheet indicated snow accumulation exceeded loss from ice discharge by $82 \pm 25 \text{ Gt a}^{-1}$. Their study found positive mass balance in east Antarctica and negative mass balance in the Antarctic peninsula and west Antarctic near-coastal region (Zwally *et al.* 2015).

A Ku-band altimeter was used in most of the aforementioned studies. The study presented here used Ka (35.75 GHz) band altimetry over the Antarctic ice sheet. The first Ka-band altimeter onboard satellite with ARGOS and AltiKa (SARAL) was launched on 25 February 2013 with coverage up to 81.5°N and 81.5°S . Ka-band frequency has several advantages over traditional Ku-band altimetry, mainly that due to its high transmitting frequency, only negligible ionosphere corrections need to be applied in the range correction (Chander *et al.* 2015). Further, the Ka-band penetrates less far into the ice surface than the Ku-band, so most of the volume backscatter will be contributed from the surface (Remy *et al.* 2015). SARAL/AltiKa has 35 days repeat cycles, a pulse width of 2 ns and pulse repetition frequency of 4000 Hz. The footprint size is $\sim 2 \text{ km}$ for the flat surface with along-track spacing of 175 m for a 40 Hz dataset (Remy *et al.* 2014). Details of the SARAL/AltiKa level-2 product formats are provided in the handbook (SARAL/AltiKa product handbook 2013).

This study focuses on intra-annual and inter-annual changes in elevations over the Antarctic ice sheet. The analysis uses only slope-corrected (slope less than or equal to 1°) data to arrive at more realistic values. This study complements existing knowledge on ice sheet surface elevation changes using Ka-band altimeter data.

Table 1. Range of the atmospheric, tidal (for cycle07, 10 October–14 November 2013) and slope correction applied to derived elevation.

Corrections	DTC (m)	WTC (m)	IC (m)	SETC (m)	PTC (m)	Slope correction for SARAL/AltiKa ($H \sim 800$ km)	
						Slope (α) (deg)	Correction (m) ($0.5H\alpha^2$)
Minimum	−2.2825	−0.0629	−0.013	−0.0792	−0.0015	0.0	0.0
Maximum	−1.2837	−0.0007	−0.0028	0.0964	0.0016	1.0	120.0

2. Data and methodology

SARAL/AltiKa 40 Hz geophysical data record (GDR) products of Exact Repeat Mission (ERM) phase consisting of 35 cycles (14 March, 2013–4 July, 2016) (<ftp://avisoftp.cnes.fr/AVISO/pub/saral/gdr.t>) were utilized. The high resolution (500 m) DEM generated from the Geoscience Laser Altimeter System (GLAS)/ICESat dataset of duration 1 February 2003 to 30 June 2005 was downloaded from the US National Snow and Ice Data Center (NSIDC) website (<http://nsidc.org/data/nsidc-0304>) and was used to generate a slope map over the Antarctic ice sheet. DiMarzio *et al.* (2007) validated the ICESat DEM elevations with Airborne Topographic Mapper (ATM) data over Greenland and observed that the mean difference varies with slope. For regions with less than 0.1° slopes, the mean difference was -32 ± 43 cm; for regions with 0.1° – 1.0° slopes, the mean difference was -66 ± 61 cm.

2.1 Elevations over the Antarctic ice sheet

Elevations (with respect to reference ellipsoid WGS84) over the Antarctic ice sheet were derived using the following relationship (Helm *et al.* 2014)

Elevation = Altitude – Range

- Dry Troposphere Correction (DTC)
- Wet Troposphere Correction (WTC)
- Ionosphere Correction (IC)
- Solid Earth Tide Correction (SETC)
- Pole Tide Correction (PTC). (1)

It has been observed that even though the threshold retracker (sea ice retracker) provides better results over ice sheets, the Ice-2 retracker is more appropriate for ice sheet studies, as threshold algorithms are not based on a physical model (Legresy *et al.* 2005; Remy *et al.* 2014; Chander

Figure 1. Illustration of slope-induced elevation correction applied to obtain corrected elevation measurement.

et al. 2015). Hence, the Ice-2 retracker derived range was used to obtain elevation.

As this study is restricted to the continental ice sheet, the surface type 2 flag was selected, which corresponds to the continental ice sheet. Surface type flags 0, 1, and 3 correspond to the ocean, closed water body and land surface, respectively.

2.2 Slope-induced elevation correction (ΔH)

Correction to be incorporated in elevation measurement due to the terrain slope is referred as ‘slope-induced elevation correction’. The magnitude of slope-induced correction in comparison to other corrections like atmospheric and tidal is as shown in table 1. A terrain of slope 1° can introduce a hundreds of metres vertical error in elevation measurement at 800 km altitude of the satellite (Hurkmans *et al.* 2012).

We have incorporated the slope-induced elevation correction (Brenner *et al.* 1983) using ICESat DEM derived slope (α) information. Geometry for applying slope-induced elevation correction is as shown in figure 1, and the following formulation was used to estimate the slope-induced elevation correction (ΔH)

$$\Delta H = H - H \cos \alpha, \quad (2)$$

$$\Delta H = H(1 - \cos \alpha), \quad (3)$$

$$\Delta H = 0.5H\alpha^2, \quad (4)$$

where α is small and measured in radians and H is the altitude of the satellite.

The ICESat DEM was used to incorporate slope correction. Using this DEM, a slope map over Antarctica was generated in the ArcGIS spatial analyst tool (desktop.arcgis.com). This tool uses an average maximum technique to compute slope. Derived elevations, which were in the geographic coordinate system, were re-projected into polar stereographic map projection in order to grid at 500 m resolution (as the ICESat DEM dataset was at 500 m with polar stereographic projection). Slope values were obtained at the required location (as per elevation data) using the slope map, and thereafter slope-induced elevation corrections were calculated. The flowchart of the methodology adopted to estimate slope-corrected elevation is shown in figure 2. These slope corrected elevations were compared with ICESat DEM derived elevation for entire ice sheet and with ATM dataset over Vostok subglacial lake.

Figure 2. Flowchart incorporating slope-induced elevation correction in SARAL/AltiKa-derived gridded elevation measurement to obtain gridded slope-corrected elevation measurements.

2.3 Intra-annual and inter-annual changes

The data were grouped into four glaciological periods (table 2). Elevation changes between datasets were obtained within each glaciological year (intra-annually), and from one glaciological year to another to analyse the inter-annual changes. We segregated all 35 cycles of the ERM phase dataset in the temporal context of the glaciological year (GY). The GY includes one melt and one freeze season of the year taken as the period between 1 October of year n and 30 September of year $n+1$ (Foresta *et al.* 2016). As the AltiKa ERM dataset was available from 14 March 2013 (start of cycle 01), GY1 includes the dataset of duration 14 March 2013 to 5 September 2013 instead of the October 2012 to September 2013 dataset. Similarly, the whole of GY4 does not include the dataset from 4 July 2016 to September 2016 as the ERM ended on 4 July 2016.

For all 35 cycles, 500 m gridded and slope-corrected elevation maps were generated. Residues were derived for each cycle by subtracting elevation from glaciological yearly grid average elevation. Due to the unavailability of ERM phase data, GY1 and GY4 do not contain a complete period as per the definition of GY, so the average elevations of GY2 and GY3, respectively, were used to generate their residual plot. After comparing AltiKa dataset with ATM elevation 0.4 m RMSD was obtained thus the elevation changes which are outside ± 0.4 m are considered significant for change else the region will be considered as stable. Hence to generate residual plot (figure 5) only those residual elevation changes are considered which are greater than or equal to 0.4 m and less than or equal to -0.4 m.

For inter-annual change analysis, cycle 07 (10 October–14 November 2013), cycle 17 (25 September–30 October 2014) and cycle 28 (15 October–19 November 2015) were selected. These cycles were gridded (500 m), had slope-induced elevation correction performed on them and elevation change maps were generated.

Table 2. Details of number of cycles included in each GY and their duration.

Duration	Number of cycles included	Glaciological year
14 March 2013–5 September 2013	Cycle01–cycle05	GY1
5 September 2013–25 September 2014	Cycle06–cycle16	GY2
25 September 2014–10 September 2015	Cycle17–cycle26	GY3
10 September 2015–4 July 2016	Cycle27–cycle35	GY4

Figure 3. Slope-induced elevation correction (ΔH) vs. slope, where slope values were derived from ICESat DEM slope map and altitude of 800 km for SARAL/AltiKa.

3. Results and discussion

3.1 Elevations over the Antarctic ice sheet

Realizing the importance of slope-induced elevation correction in altimetry data, a relationship was derived between slope up to 1° and the magnitude of correction required in the elevations derived from AltiKa data. This is shown in figure 3.

In order to check our slope-induced elevation correction data we compared the elevation values with ICESat-derived high-resolution DEM elevations, although the two datasets do not correspond to each other in a time domain. This comparison helped in determining the significance of slope-corrected data. Statistics were derived for three cycles (cycle 07, 17, 28), and the results obtained for cycle 07 (October–November 2013) are discussed as being representative of the three datasets. Applying slope correction for the entire continent irrespective of the constraint on the slope produced inferior statistical results. Correlation reduced to 0.30 from 0.99 (correlation for slope-uncorrected elevation), bias increased to -108.50 m from 9.76 m and root-mean-square-deviation (RMSD) increased to 3723.51 m from 42.63 m. But after applying slope-correction to only those locations, where the slope was less than or equal to 1° , we observed a dramatic improvement in elevation values compared to the DEM values.

Although the correlation was 0.99 for uncorrected and slope-corrected elevation, bias and RMSD were improved to -2.88 and 23.04 m, respectively. The results showed that the applied slope correction method works better over regions where the slope is less than or equal to 1° . The statistic also indicated that applying the slope less than or equal to 1° criterion reduced the total number of points to 3,073,412 from 3,243,499 ($\sim 5\%$ of the data loss). Thus, for further analysis, only those datasets were used where the slope is less than or equal to 1° . Further we compared the ATM (<https://atm.wff.nasa.gov/data-products>) elevation dataset of date 26 November, 2013 with SARAL/AltiKa 500 m gridded and slope corrected dataset of cycle 08 (14 November–19 December, 2013) over small region of Vostok subglacial lake so that dataset will overlap in temporal context and RMSD was found to be 0.4 m.

As far as elevation change is concerned slope-induced elevation correction will cancel out but it leads to underestimation of the area-integrated volume change (Hurkmans *et al.* 2012), hence, it is mandatory to incorporate slope-induced elevation correction, as this study will lead to volume change estimation in the future.

Figure 4 shows the slope (less than or equal to 1°) corrected elevation map over the Antarctic ice sheet, derived using cycle 07 (October–November 2013). The elevations were found to vary between 0 and ~ 4000 m. Elevations in most east Antarctic regions ranged between 3000 and 4000 m, while in west Antarctica the majority of elevations lay below 3000 m. In the central part, elevations

Figure 4. Slope-corrected elevation map of the Antarctic ice sheet derived from SARAL/AltiKa cycle 07 of duration 10 October to 14 November 2013.

Figure 5. Histogram of residual elevation change (m) for (a) GY1, (b) GY2, (c) GY3, and (d) GY4 displaying number of elevation changes outside the range ± 0.4 m. In each GY, respective cycle number is mentioned on the top of the individual bar plot.

were ~ 4000 m and went on decreasing towards the coast.

3.2 Intra-annual changes

Figure 5(a) indicates the intra-annual residual elevation changes from cycle 01 (March–April 2013, summer end) to cycle 05 (August–September 2013, winter end). Probably as all these 5 cycles fall in winter period, positive number of elevation changes were more than the number of negative elevation changes.

Figure 5(b) displays intra-annual residual elevation changes from cycle 06 (September–October 2013, winter end) to cycle 16 (August–September

2014, winter). For cycle 06 no changes were observed outside ± 0.4 m possibly indicating the stability of the study region. From cycle 07 (October–November 2013) to cycle 08 (November–December 2013) both, number of positive and negative elevation changes increased. For cycle 09 (December 2013–January 2014), number of negative changes was further increased and surpassed the number of positive elevation changes, indicating maximum number of negative elevation changes during GY2 probably due to summer month. During cycle 10 (January–February 2014) both positive as well as negative number of points decreased and negative number of points were more than positive and situation persisted for cycle

11 (February–April 2014). During cycle 12 (April–May 2014) positive points took over the negative number probably indicating onset of winter. For cycle 13 (May–June 2014), 14 (June–July 2014), 15 (July–August 2014) and 16 (August–September 2014), positive number of points remained more than the negative number of points probably because these cycles fall in winter period.

Figure 5(c) shows intra-annual residual elevation changes for the duration between cycle 17 (September–October 2014, winter end) and cycle 26 (August–September 2015, winter). We observed an increase in both the elevation changes during cycle 17 and 18 (October–December 2014) but number of positive elevation changes was more than number of negative elevation changes. During cycle 19 (December 2014–January 2015) number of negative elevation changes surpassed the number of positive elevation changes indicating maximum number of negative elevation change during GY3 probably due to summer month. During cycle 20 (January–February 2015) negative elevation changes were more than positive elevation changes probably indicating that the summer situation persisted for cycle 20 but negative elevation underwent decrease in number with respect to cycle 19. During cycle 21 (February–March 2015) negative changes decreased and positive number surpassed the negative number of points. During cycle 22 (March–April 2015) both number were decreased but positive changes still exceeded the negative

changes. During cycle 24 (May–July 2015), 25 (July–August 2015) zero elevation changes were observed outside ± 0.4 m indicating stability of the study region. During cycle 26 the number of positive elevation points were more than the number of negative points.

Figure 5(d) displays intra-annual residual elevation changes as we proceed from cycle 27 (September–October 2015, winter end) to cycle 35 (June–July 2016, peak winter). As observed in GY3 similarly in GY4 both elevation changes were increased during cycle 27 and 28 (October–November 2015) with positive number of points more than the negative number of elevation changes. During cycle 29 (November–December 2015) negative elevation changes were increased while positive changes were decreased with number of positive elevation change points more than number of negative elevation changes. During cycle 30 (December 2015–January 2016) numbers of negative and positive elevation changes were decreased but negative changes exceeded the positive changes probably due to summer. During cycle 31 (January–March 2016) both the changes increased but positive elevation changes exceeded the negative number probably indicating onset of winter. During cycle 33 (April–May 2016), 34 (May–June 2016) and 35 (June–July 2016) number of positive elevation changes remained greater than the number of negative elevation changes probably due to winter. After 4 July 2016 ERM mission was

Figure 6. Elevation change map of (a) cycle 17–cycle 07, (b) cycle 28–cycle 17, and (c) cycle 28–cycle 07. Site 1 indicates elevation changes near Pine Island and Thwaites glaciers and site 2 indicates elevation changes over east Antarctica.

changed to drift orbit mode, led to incompleteness of cycle 35 and as a result of these it contains less number of data points.

3.3 Inter-annual changes during 2013–2015

For inter-annual change analysis, we used three similar season cycles during 2013–2015: cycle 07 (10 October–14 November 2013), cycle 17 (25 September–30 October 2014) and cycle 28 (15 October–19 November 2015).

To examine yearly changes, we subtracted cycle 07 from 17, and cycle 17 from 28. Elevation changes in east Antarctica (site 2) were negative, as seen in west Antarctica (site 1), during 2013 and 2014 (figure 6a). During the next year (2014–2015) east Antarctica showed positive elevation changes while west Antarctica still showed negative changes (figure 6b).

Based on the variability observed in the final three plots (figure 6) of inter-annual changes, 15 regions of interest (ROIs) were selected for further discussion. Four ROIs were selected in west Antarctica, seven were selected in east Antarctica and the remaining four from the Antarctic coastal region (table 3). Statistical variability was studied in detail to observe the exact change in ice sheet elevation. Outliers in the elevation changes obtained over ROIs were removed by applying the inter-quartile range criterion, and later on mean values were derived. During 2013–2014, ROIs 1, 2, 5, 6, 9, 11, 13, 14 and 15 revealed a decrease in elevation change, with a maximum decrease of 2.11 m over ROI-2 (Thwaites glacier) and a minimum decrease of 0.006 m over ROI-11 (Vostok subglacial lake). The maximum increase of 0.41 m was observed over ROI-8 in east Antarctica, while the minimum increase of 0.029 m was observed over ROI-10. ROIs 1, 2, 14 and 15 showed a decrease in elevation change again during 2014–2015, whereas ROIs 5, 6, 9, 11 and 13 displayed an increase, with a maximum increase of 0.38 m over ROIs 5 and 6 and a minimum increase of 0.027 m over ROI-11. Overall during 2013 and 2015 over ROI-1 (Pine Island glacier), a 1.98 m decrease in elevation was observed while over ROI-2 the elevation decrease was higher (3.32 m). Similar thinning at Pine Island glacier tributaries exceeding the value of 1 m/yr was reported by Helm *et al.* (2014) detected during 2011–2014 from Cryosat-2. Over ROI 3, an increase in elevation change of 0.54 m was observed, while elevation change over ROI-4 was observed to be 0.11 m. Increases of the order

Table 3. Mean elevation change values obtained over ROIs during 2013–14 (cycle17–cycle07), 2014–15 (cycle28–cycle17) and 2013–15 (cycle28–cycle07).

Sl. no.	Location	Cycle17–cycle07			Cycle28–cycle17			Cycle28–cycle07		
		Total number of points	Mean elevation change (m)	Total number of points	Mean elevation change (m)	Total number of points	Mean elevation change (m)	Total number of points	Mean elevation change (m)	Total number of points
1	Near Pine Island Glacier (74.5–75.0°S, 94–94.5°W)	111	−0.65	64	−1.29	97	−1.98	97	−1.98	97
2	Near Thwaites Glacier (75.5–76.0°S, 107.0–107.5°W)	50	−2.11	31	−0.54	45	−3.32	45	−3.32	45
3	78.5–79.0°S, 95.0–95.5°W	85	0.21	87	0.34	99	0.54	99	0.54	99
4	78.0–78.5°S, 114.0–114.5°W	114	0.38	68	0.075	80	0.11	80	0.11	80
5	71.0–71.5°S, 50.0–50.5°E	42	−0.49	18	0.38	58	0.58	58	0.58	58
6	70.5–71.0°S, 59.0–59.5°E	69	−0.04	98	0.38	94	0.57	94	0.57	94
7	73.0–73.5°S, 16–16.5°E	65	0.037	78	−0.24	79	0.48	79	0.48	79
8	72.5–73.0°S, 15.0–15.5°E	32	0.41	26	−3.30	61	0.43	61	0.43	61
9	80.0–80.5°S, 59.0–59.5°E	149	−0.12	122	0.19	127	0.19	127	0.19	127
10	71.5–72.0°S, 15.0–15.5°E	35	0.029	75	−0.6	58	−0.48	58	−0.48	58
11	Vostok (76.5–77°S, 105–105.5°E)	84	−0.006	36	0.027	97	0.012	97	0.012	97
12	69.0–69.5°S, 100–100.5°E	117	0.12	50	−0.1	63	−0.1	63	−0.1	63
13	69.0–69.5°S, 120.0–120.5°E	21	−1.06	41	0.35	61	−0.56	61	−0.56	61
14	69.0–69.5°S, 150.0–150.5°E	74	−0.63	44	−0.63	66	−2.64	66	−2.64	66
15	79.0–79.5°S, 32.0–32.5°W	98	−0.36	33	−0.67	59	−1.43	59	−1.43	59

cy28 (15th Oct – 19th Nov 2015)– cy07 (10th Oct – 14th Nov 2013)

Figure 7. Interpolated elevation change map of cycle 28–cycle 07. Blue arrow indicates elevation changes near Pine Island and Thwaites glaciers. Red arrow indicates elevation changes over east Antarctica.

of 0.58 and 0.57 m were observed over ROIs 5 and 6, respectively, and 0.48 and 0.43 m increases were observed over ROIs 7 and 8, respectively. A 0.48 m elevation decrease was observed over ROI-10 in east Antarctica. At ROI-11 over Vostok subglacial lake, elevation change of 0.012 m was observed indicating its stability. Zwally *et al.* (2015) also found that Vostok subglacial lake in east Antarctica is dynamically stable region and dh/dt varies from approximately $+0.0 \pm 1$ to $+3.0 \pm 1$ cm/yr using ICESat (2003–2008). Four ROIs, 12, 13, 14 and 15, were selected along the Antarctic coast, and the elevation decreases observed were 0.1, 0.56, 2.6 and 1.43 m, respectively.

From figure 6(c) and ROI analysis it was observed that most of the elevation changes lie between -2 m and $+0.6$ m, hence interpolated plots of elevation change during 2013–2015 were again generated with a range of -2 to $+0.6$ m as shown in figure 7.

4. Conclusions

This study has focused on the derivation of ice sheet elevations and their variability in the spatial and temporal domain over Antarctica using SARAL/AltiKa 40 Hz GDR products of ERM phase with slope correction applied for areas having a slope less than or equal to 1° . The slope corrections were made using an ICESat-based DEM as reference data. Though a correlation of 0.99 was found between AltiKa data and ICESat data, the slope correction improved the bias and RMSD to

the order of -2.88 and 23.04 m, respectively. The test was performed on the data of three cycles, used in the inter-annual comparisons. The data reaffirm the general topography of the Antarctic continent known so far through other datasets. Along with comparison of ATM dataset with AltiKa derived elevation yielded 0.4 m RMSD over a part of Vostok subglacial lake. The intra-annual change observed during GY2, GY3 and GY4 shows negative number of elevation change points exceeded the positive elevation change points for cycle 09, 10, 11, 19, 20 and 30, which can be attributed to Antarctic summer. For GY2 and GY3 (which are complete GY) number of negative elevation change displayed increasing trend from September–October (winter end) to December–January (summer).

Inter-annual changes obtained during 2013–2015 have shown the variability of positive–negative elevation changes. Most of east Antarctica is showing an increase in elevation, with a maximum of ~ 0.6 m changes. There are a few regions which show negative elevation changes of about ~ 0.48 m. An elevation change of 0.012 m was observed over Vostok subglacial lake (ROI-11) for the period 2013–2015, indicating a stable region in east Antarctica. In west Antarctica, particularly near Thwaites and Pine Island glaciers, more than 2 m reductions in elevation were observed, which can be attributed to warm circumpolar deep water (CDW) in the Amundsen sea. Similar melting behaviour around Thwaites and Pine Island glaciers has also been reported by Stanley *et al.* (2011), Mouginot *et al.* (2014) and Smith *et al.* (2017). The changes observed between the two study periods, 2013–2014 and 2014–2015, indicated positive (2014–2015) and negative (2013–2014) changes in east Antarctica, whereas there were negative elevation changes near Thwaites and Pine Island glaciers, west Antarctica. Hence, results suggest contrasting behaviour of the two major parts of the continental ice sheet, with mass gain in the eastern part and loss in the western part.

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