Parameterization of Surface Roughness Based on ICESat/GLAS Full Waveforms: A Case Study on the Tibetan Plateau

JUNCHAO SHI, MASSIMO MENENTI, AND RODERIK LINDENBERGH

Department of Geoscience and Remote Sensing, Delft University of Technology, Delft, Netherlands

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ABSTRACT

Glaciers in the Tibetan mountains are expected to be sensitive to turbulent sensible and latent heat fluxes. One of the most significant factors of the energy exchange between the atmospheric boundary layer and the glacier is the roughness of the glacier surface. However, methods to parameterize this roughness for glacier surfaces in remote regions are not well known. In this paper, the authors use the data acquired by Ice, Cloud, and Land Elevation Satellite (ICESat)/Geoscience Laser Altimeter System (GLAS) laser altimetry from February 2003 to November 2004 along several tracks over glaciers of the Nyainqêntanglha range in central Tibet. The authors make a study of the waveforms measured by the ICESat/GLAS laser system over mountainous and glacial areas. The surface characteristics are evaluated within laser footprints over the glacier outlines based on the glaciological inventory of the Tibetan Plateau constructed by the Cold and Arid Regions Environmental and Engineering Research Institute (CAREERI), Chinese Academy of Sciences. For this purpose, the authors extract waveform parameters: the waveform width, the number of modes, and the RMS width of the waveform. These parameters are compared with surface slope and roughness obtained from the Advanced Spaceborne Thermal Emission and Reflection Radar (ASTER) Global Digital Elevation Model (GDEM). Through this analysis, the impact of morphology on the returned laser waveform is shown for the Nyainqêntanglha range. The roughness and the slope of the surface can be quite significant and may contribute from several meters to tens of meters to the pulse extent. The waveform analysis results indicate that the received waveforms are capable representations of surface relief within the GLAS footprints.

1. Introduction

Recent meteorological and glaciological studies show that a significant proportion of the total energy absorbed by glacier surfaces comes from turbulent fluxes (Munro 1989; Oerlemans 2000; Denby and Greuell 2000; Brock et al. 2000). When evaluating the energy exchange over glaciers, the atmospheric boundary layer (ABL) is generally supposed to be stratified (Arya 1973, 1975). This energy transfer process is commonly parameterized by the “bulk eddy” aerodynamic method, which assumes that airflow in the ABL is turbulent and fully adjusted to the underlying terrain. The Monin–Obukhov similarity theory is applied to parameterize the relation between the meteorological elements and surface fluxes (Hock and Holmgren 1996; Denby and Smeets 2000; Andreas 2002). The irregularity/undulations of the glacier surface, besides albedo, temperature, and so on, are significant controllers during the surface–atmosphere exchange of mass and energy (Arya 1977). Normally, this irregularity is defined as roughness, which is parameterized as a momentum sink for the atmospheric flow.

These surface irregularity/undulations are generally categorized according to their sizes. The large-scale undulations have usually a typical wavelength of 3–4 times the ice thickness (Budd and Carter 1971). In contrast, small-scale surface irregularities associated with sastrugi and wind-formed features form over periods of hours to days and are inherently transient and unpredictable in a deterministic sense (Van der Veen et al. 2009). The other type, containing crevasses, is developed where the stresses of the ice form variations in flow exceeding the breaking strength of the ice in tension. Glaciers with these types of features are widely distributed in mountainous areas, and their shape and flow are characterized...
by fluctuating surface relief and downhill flow. Many energy balance models use refinements of the boundary layer turbulence theory, as described by Obukhov (1971). The aerodynamic roughness \( z_0 \) is used to scale the logarithmic increase of wind speed with height in a neutrally stratified layer from a level of no motion near the surface. Furthermore, the aerodynamic roughness \( z_0 \) is experimentally determined from wind velocity and air temperature profiles. Such resulting roughness estimates are found to be in good agreement with the existing relationships linking the geometric and the aerodynamic roughness. This suggests that for natural surfaces, \( z_0 \) can be estimated on the basis of the geometric characteristics of the roughness elements.

According to the studies by Arya (1975), Andreas (1987), Oke (1987), and Stull (2009), the dimensions and density distribution of surface roughness elements are influential on \( z_0 \) when normal turbulence flows over melting snow and ice surfaces. Because of increasing height, surface area, and density of surface roughness, the value of \( z_0 \) increases until the ratio between the silhouette area (upwind face of elements) and the unit ground area reaches 0.4. After this value, a transition to "skimming" flow occurs and \( z_0 \) starts to reduce again (Garratt 1992; Brock et al. 2006). The standard method to derive \( z_0 \) is from the vertical profiles of horizontal wind speed and air temperature, using measurements at two or more heights in the ABL.

Until the past few decades, there have been two main acceptable strategies to estimate the aerodynamic roughness. On one hand, in situ measurements dependent on the bulk transfer equations were tested both for polar ice sheets and various mountainous glacier zones (Andreas 2002). The temperature, wind speed, and other relevant meteorological elements are recorded and applied for a local aerodynamic roughness approximation. On the other hand, since the end of last century, studies done by Menenti and Ritchie (1994), Su et al. (2001), and DeVries (2003) indicate that the capabilities of measurements obtained by laser and optical remote sensing are promising for the determination of the aerodynamic roughness. Compared to conventional techniques, these approaches to estimate \( z_0 \) by remote sensing measurements have the potential to achieve a resolution that is much higher than meteorological measurements in both the spatial and temporal domain.

Spaceborne laser altimetry is a relatively new instrumental possibility to assess the morphology of the illuminated objects based on the full-waveform information. The return waveform is influenced by effects like cloud attenuation, system noise, and morphology. In January 2003, the National Aeronautics and Space Administration (NASA) launched the Geoscience Laser Altimeter System (GLAS) loaded on the Ice, Cloud, and Land Elevation Satellite (ICESat), whose accuracy reaches 15 cm in the vertical direction within an approximately 65-m footprint, with a spacing of about 172 m between two adjacent shots (Abshire et al. 2005). Compared to radar altimeter missions, the emitted pulse width and laser divergence both decreased. These properties enable ICESat/GLAS to observe relatively small-scale characteristics of morphology of glaciers in the alpine zone.

Although, the track coverage in middle latitude areas (such as the Tibetan Plateau) is not as dense as over the polar zones, it still supplies enough signals for statistical analysis of the full-waveform signal. Therefore, we address the following issues in this research: (i) we estimate the theoretical relation between spaceborne laser signals and surface irregularity, and (ii) we derive the representative parameters of this irregularity as roughness from spaceborne laser altimetry.

2. Study area and data

a. Study area

In this study, we consider the glaciers Guren, Guila, Lisheng, and Bili located on the southern slope of the Nyainqêntanglha range (30°19′00″N, 90°7′50″E). It is a mountain range lying approximately 300 km northwest of Lhasa in central Tibet (Caidong and Asgeir 2010). The ice volume of about 4900 km³ is distributed over 7080 glaciers in this mountain range. Nevertheless, there are only 28 valley glaciers longer than 10 km (Zongtai and Huian 1992; Yao et al. 2007). These glaciers play a significant role in the circulation of the Yarlung Tsangpo and Nu rivers. The study area is limited by elevation ranging from 4498 to 7162 m, and the mean slope is 17.51°. A glacier inventory is a basic tool to explore issues related to glacial studies and morphology studies in glacial areas. Since the 1970s, Chinese scientists have focused on this task through aerial photographs and in situ measurements. The latest Chinese glacier inventory has been a fusion of new topographic maps, digital elevation models (DEMs), optical imagery [Advanced Spaceborne Thermal Emission and Reflection Radar (ASTER) and Landsat Thematic Mapper (TM)/Enhanced TM (ETM)\], and historical records. All four glacier outlines used below correspond to the Chinese glacier inventory by Cold and Arid Regions Environmental and Engineering Research Institute (CAREERI), Chinese Academy of Sciences (CAS) described in Shi and Liu (2000) and Shi et al. (2009).

b. ICESat/GLAS data

The ICESat/GLAS instrument consists of two channels, at 1064 nm and 532 nm. The first channel is designed...
for the measurement of surface topography. There are three lasers on ICESat/GLAS. Laser 1 started to collect data in February 2003, but it failed shortly after its launch. Lasers 2 and 3 have collected data with a number of 33-day subcycle campaigns to increase their longevity. The receiver on GLAS records the echo waveform in 200 bins over sea and 544 (1000) bins over land, depending on which of the two lasers is used, with a beam width of ~110 μrad and a pulse rate of 40 s⁻¹ (Kwok et al. 2006).

The ICESat/GLAS data used in this study were acquired within campaign L2A (September–November 2003), L2B (February–March 2004), L2C (May–June 2004), and L3A (October–November 2004). Among 15 GLAS data products, we investigate the products of Level 1A raw data and Level 2 global land surface elevation data, releases 428 and 429, available from the National Snow and Ice Data Center (NSIDC). The GLAS products used in the assessment are GLA01 (L1A Global Altimetry) and GLA14 (Land Surface Altimetry). The former stores the transmitted and received waveforms from the altimeter while the latter contains land surface elevation, land footprint geolocation, and reflectance, as well as geodetic, instrumental, and atmospheric corrections for range measurements. These two datasets can be linked by the record index and shot number (1–40). A total of 1316 shots is selected for further evaluation in this area. The area illuminated by the ICESat/GLAS pulse will vary significantly during the span of each campaign, over the course of one orbit, and even shot by shot (shown in Fig. 1).

c. ASTER GDEM data

ASTER imagery is captured from a multispectral imaging sensor on board the NASA Terra platform, launched in December 1999. It is designed to monitor surface geodynamic processes, such as glacier movements, landslides, and flood disaster assessment, with higher resolution and better coverage than other previous spaceborne optical instruments. The main spectral and geometric capabilities of the imaging system are three visible and near-infrared (VNIR) bands, six shortwave infrared (SWIR) bands, five thermal infrared (TIR) bands, and one NIR along-track stereo band with 15, 30, 90, and 15-m spatial resolution, respectively (Yamaguchi et al. 1998).

For normal ASTER DEM generation, the stereo band 3B is often combined with 3N images (the same spectral range 0.76–0.86 μm) (Fujisada et al. 2005). After ground control point selection, epipolar image selection, image matching, and geocoding, the final ASTER DEM is established. Then, ASTER DEM standard data products are resampled with 30-m postings and with Z accuracies generally between 10 and 25-m root-mean-square error (RMSE) (Toutin and Cheng 2001).

3. Methods

a. Slope and roughness derived from ASTER GDEM

To evaluate the GLAS performance over mountainous glacial surfaces of varying relief, slope and roughness values for representative geomorphological terrains in our study area were quantified using ASTER GDEM. These two parameters were defined at a length scale of 90 m, approximating the footprint diameter of the GLAS laser beams by calculating the best-fitting local mean plane to a 3 × 3 grid of elevation values (see Fig. 2). The slope calculation [Eqs. (1)–(5)] used the algorithm described by Burrough et al. 1998:

\[
slope_{\text{radians}} = \arctan(\Delta S),
\]

\[
\Delta S = \sqrt{(dz/dx)^2 + (dz/dy)^2},
\]
\[
\Delta H = \sum_{i=1}^{9} [H_i - (r_{x_i} + s_{y_i} + t)]^2.
\] (8)

All these implementations are done in the commercial software ArcInfo.

b. ICESat full-waveform fitting process

The quality of the full waveform is essential to determine the range from the satellite to the features on the ground. Normally, the emitted pulse is considerably stable and introduces little error in the elevation calculation. However, the echo waveform can be modified by changes in emitted pulse energy, cloud effects, surface reflectivity, and characteristics of the surface illuminated within a single laser footprint (Yi et al. 2005). To shape the features of echo waveforms, we used a waveform-fitting procedure to remove noise. Waveforms were removed from further analysis for one of the following reasons: 1) the raw waveform cannot be decomposed by the Gaussian fitting algorithm or 2) the deviation between ICESat GLA14 and ASTER GDEM is greater than 100 m.

Based on previous waveform studies by the NASA ICESat/GLAS team, the assumption is that the waveform is decomposable as a number of Gaussian components and a noise component (Abshire et al. 2003). In our case, the glaciers of interest are located in a continental summer-precipitation climate. In general, the peaks of the echo waveform represent relevant surface morphological features. Waveform extent is defined as the vertical distance between the first and the last elevation at which the waveform exceeds a threshold level corresponding to the peak and 4 times the standard deviation of the background noise (Abshire et al. 2005). The decomposition of the waveform is formulated as follows:

\[
w(t) = \sum_{m=1}^{N_p} W_m + \epsilon, \tag{9}
\]

\[
W_m = A_m e^{-(t-t_m)^2/2\sigma_m^2}, \tag{10}
\]

where \(w(t)\) is the amplitude of the waveform at time \(t\), \(W_m\) is the contribution from the \(m_{th}\) Gaussian mode, \(N_p\) is the number of Gaussian modes found in the waveform, \(A_m\) is the amplitude of the \(m\)th Gaussian, \(t_m\) is the Gaussian position, \(\sigma_m\) is the 1/e half-width (standard deviation) of the \(m\)th Gaussian, and \(\epsilon\) is the bias (noise level) of the waveform.

The original data from GLA01 is formatted in binary and counts. We converted them to the American Standard Code for Information Interchange (ASCII) and
voltage for further processing. In the fitting step, Gaussian components are fitted to the return waveform.

A robust least squares method is used to compute the model parameters. That is, the values of \(v, A_m, t_m, \) and \(s_m\) are obtained by fitting the theoretical model to the observed waveform in such a way that the difference between model and observation is minimized in the least squares sense. The parameter’s \(n\) modes (the number of fitted peaks) and the full-waveform width are generated for further analysis in later sections (Duong et al. 2009). Both of them encode specific physical information: \(n\) modes, the number of waveform peaks, roughly corresponds to the number of elevation levels (including the objects and the earth’s surface), and the width of the full waveform represents the elevation range between lowest and highest surface elements within an ICESat/GLAS footprint (Fig. 3).

c. Theoretical impact of morphology on the ICESat full waveform

In this paragraph the theoretical response of an ICESat waveform on the reflecting terrain is described (Gardner 1982). This theory assumes that the average waveform shape due to terrain reflection is related to the probability density of variation of terrain elevations within the laser footprints. We use the series of equations below, where \(N\) is the photon count, \(T_p\) is the centroid time of the received pulse, and \(\sigma_{SR}\) is the RMS pulse width. These are the functions of temporal moments of the pulse, and they are defined as follows:

\[
N = \int_0^\infty p(t)\,dt, \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \quad \ quad
where $\sigma_F$ is the transmitted laser RMS pulse width; $\sigma_h$ is the RMS width of receiver impulse response; $\text{Var}(\Delta \xi)$ is the RMS roughness parameter of the statistical surface; $c$ is the speed of light; $S_x$, $S_y$ are the along-track and across-track slope, respectively; $q_t$ is the laser beam divergence angle; and $f$ is the laser-pointing angle off nadir.

For the ICESat/GLAS system, both the pointing uncertainty (1.5 arc sec) and the normal off-nadir pointing angles ($< 1^\circ$) are small; thus, the above equation was simplified as

$$
\text{Var}(\Delta \xi) = \frac{c^2}{4} \left[ \sigma_{SR}^2 \left( \frac{1}{\sigma_F^2} + \frac{1}{\sigma_h^2} + \frac{4\text{Var}(\Delta \xi)}{c^2} + \frac{4z^2 \tan^2 q_t}{c^2} \cdot (\tan^2 q_t + \tan^2 S) \right) \right].
$$

From these equations, the laser system contributions to the width of the received pulse include the widths of the transmitted pulse and the receiver impulse response. The received pulse is broadened by the surface roughness and slope of the ground target and by the beam curvature.

The relevant procedures of this study are summarized in a scheme in Fig. 4. Part 1 is mainly about the full-waveform-fitting algorithm. We use the methodology mentioned before to fit all the selected waveforms in the study region. Part 2 aims at comparing the waveform parameters obtained from the fitting procedures to the
topographic slope and roughness as obtained from ASTER GDEM.

4. Results

In this section, we show several typical example waveforms for different locations in the area and analyze the variation according to their topography and landscape type. Moreover, we do a case study to investigate the distribution of parameters on the named glaciers Guren, Guila, Lisheng, and Bili. We divided each of the four glaciers into ablation and accumulation zone dependent on the glaciological inventory of Nyainqêntanglha by CAREERI, CAS. The attributes of these glaciers are mentioned in Table 1.

Next, we compare the ICESat-derived parameters, elevation, waveform width, and waveform number of modes to the ASTER GDEM elevation, roughness, and slope. Finally, the experimental waveform response is estimated with the predicted response, as discussed in section 3c.

a. ICESat full-waveform examples over the study area

Accumulation and ablation zones can be divided into much more specific zones, these being distinct zones with characteristic features in the surface layers of a glacier. As most laser scanning systems are operating in the near-infrared part of the electromagnetic spectrum, some differences on the surface by different surface types, for example, snow, firn, and ice, can be observed.

In the view of glaciomorphology, over 90% of the glaciers in this study area are mountain glaciers. First of all, we set several criteria to describe the morphological characteristics of a glacier: the general shape, the terminus location, the longitudinal profile, supply source of the glacier, and activity of the ice tongue. Moreover, we geolocated the ICESat/GLAS samples on the glaciers using the CAREERI glacier mask. Finally, we pick up four larger glaciers (Guren, Guila, Lisheng, and Bili) to analyze further.

The four named glaciers are close to each other and span an altitude range of 796.5 m on average. Guren glacier (30°12.78′N, 90°14.31′E) lies in a valley, and the orientation of exposed area and ablation zone are both northern. It is a cirque from morphology, whose terminus is debris covered and downhill, and the longitudinal profile is regular down to the foot of the hill and is not very steep. The tongue retreats slowly (Oerlemans and Fortuin 1992; Yao et al. 2007). Guila (30°11.92′N, 90°16.35′E) is located southeast of Guren and is also a valley glacier consisting of multiple basins on the upward side. Its terminal is debris covered, but the longitudinal profile reveals some ice falls, crevasses, and tower-shaped ice; the main source is avalanche and relevant snow. The retreat of the tongue is still not very fast. Lisheng and Bili are mountain glaciers; thus, frontal characteristics are expended. The main mass source of the glacier is the same as for Guren, and the glacier front has retreated slightly.

Mountainous terrain, where topographic relief within a footprint is small compared to landscape sizes, typically yields a multimodal ICESat/GLAS waveform. The intensity is a good indicator for glacial optical properties. Ice, snow, and surface irregularities (mainly crevasses) show a good differentiation in terms of geometry and reflectance. Figures 5a and 5b show shots 34 and 21 geolocated in the accumulation zone of Guila, whose slope and roughness are 9.9°, 1.01 m for shot 34 and 49.5°, 1.54 m for shot 21. The distribution is Gaussian in both cases, shot 34 is fitted with only one main peak, and the amplitudes are both close to 1.2 V. Although the waveform width of shot 21 is broadened by the slope effect and is fitted with three peaks, it still comes from a glacier. We considered two other shots (22 and 26) in the ablation zone of Guila and Lisheng (Figs. 5c,d). The most remarkable feature is their larger widths, about 11.27 and 12.01 m, respectively. There are five peaks in shot 22; the peak of the last Gaussian mode responds to the lower ground surface within the ICESat 70-m footprint. Specifically, the number of peaks in shot 26 is less than six as well, the peak of the last Gaussian mode corresponds to the surface elevation, and the maximum intensity is also much lower than the values in Figs. 5a and 5b. In combination with the glacier inventory by CAREERI and Landsat TM imagery 2003, the waveform can be interpreted as illuminated on the crevasses. Similarly, shot 26 is interpreted as coming from debris overlaying the glacier. We find more oscillations in the return pulse, which involved more morphological features.

<table>
<thead>
<tr>
<th>Glacier</th>
<th>Area (km²)</th>
<th>Slope (°)</th>
<th>Roughness (m)</th>
<th>Length (km)</th>
<th>Max Elevation (m)</th>
<th>Min Elevation (m)</th>
<th>Area (km²)</th>
<th>Accumulation Zone (km²)</th>
<th>Ablation Zone (km²)</th>
<th>Exposed Area (km²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Guren</td>
<td>5.6</td>
<td>9.9</td>
<td>1.01</td>
<td>4.2</td>
<td>6260</td>
<td>5830</td>
<td>5400</td>
<td>78</td>
<td>108</td>
<td>48</td>
</tr>
<tr>
<td>Guila</td>
<td>14.69</td>
<td>12.78</td>
<td>1.6</td>
<td>13.69</td>
<td>6360</td>
<td>5924</td>
<td>5380</td>
<td>5900</td>
<td>72</td>
<td>32</td>
</tr>
<tr>
<td>Lisheng</td>
<td>4.4</td>
<td>4.32</td>
<td>1.6</td>
<td>3.3</td>
<td>6360</td>
<td>5924</td>
<td>5488</td>
<td>5900</td>
<td>72</td>
<td>32</td>
</tr>
<tr>
<td>Bili</td>
<td>5.6</td>
<td>4.2</td>
<td>1.4</td>
<td>4.2</td>
<td>6080</td>
<td>5855</td>
<td>5630</td>
<td>48</td>
<td>0.07</td>
<td>0.4</td>
</tr>
</tbody>
</table>
FIG. 5. Typical waveform examples were picked out over different features: (a),(b) accumulation zone of Guila; ablation zone of (c) Guila and (d) Lisheng; (e) location in the rare mountainous area; and (f) the lake surface. Note that to describe the complexity of bare mountainous area, we set number of modes larger than six, when the specific plotting is implemented in (e).
Through ICESat track coverage of the four glaciers’ analysis, we selected Guila and Guren as the most interesting regions for a case study. Then we generated the distribution of these waveform parameters on both ablation and accumulation zones (see Table 2). The measurements are located in the ablation zone, near the tongue of the glacier, and the air temperature is above pressure freezing point, thus speeding up the melting of glaciers. In this area, there is no snow and possibly also a thin layer of water covering the glacier. Clearly, the reflectivity of this area is lower than surrounding bedrocks. The quantity nmodes is sensitive to the roughness of the surface, and the value of nmodes increased by the roughness of the four named glaciers (1.03 ± 0.04–1.15 ± 0.17 m) is larger than three on average. Then the parameters in glacier and mountainous zones are computed, and the results are shown in Table 3. To link the ICESat waveforms and the morphology further, a total of 1188 points is selected in the mountains.

b. Comparison of parameters derived from ASTER GDEM and ICESat data

To compare the quality of the ICESat elevations, we make a scatterplot of its elevation against that from ASTER GDEM. It shows a linear trend in Fig. 6a. The differences of elevations between ASTER GDEM and ICESat are also visualized in the histogram in Fig. 6b. In terms of absolute magnitude, it indicates that the accuracy of ICESat elevation is higher (ASTER GDEM Validation Team 2009; Zwally et al. 2002). The largest difference is 126.10 m; the difference is on average around 33.96 ± 18.17 m over all samples.

The surface features’ roughness and slope are chosen as the shaping factor of the waveforms (see Fig. 7). Next, we compute the cumulative histograms of ASTER GDEM roughness and slope compared to the ICESat/GLAS full width of the return waveform (Fig. 8).

Based on the cumulative histograms above, we explore the variability and spreading (or dispersion) of the ASTER GDEM slope and roughness with ICESat/GLAS number of modes and waveform width. Figures 9 and 10 show the scatterplots of the ICESat waveform number of modes and the ICESat waveform width versus the ASTER GDEM slope and roughness, respectively. As expected, both the number of modes and the waveform width increase, on average, with the corresponding ASTER GDEM slope and roughness. Indeed, both surface slope and roughness have a widening effect on the waveform: as the vertical distribution of scatterers increase with both growing slope and roughness, the waveform widens, as there is a greater time difference between return time of the lowest and highest scatterer in the illuminated footprint. The relation between surface slope and roughness versus number of modes is less direct. But clearly, in the high-relief mountainous terrain, the possibility of scattering at different heights contributes to the increase of the number of modes.

c. Analysis of ICESat full-waveform parameters as a function of morphology

In this section, we will measure surface roughness at scales on the order of the footprint size. As described before, the shape of the laser echo pulse reveals much about the topography of the region within the laser footprint. Different elevations within the area of footprint will reflect the light with different travel times, leading to pulse spreading. A height difference of $h$ between two points would result in reflected photons returning with a time difference of $2h/c$, $c$ being the speed of light.

Based on the theory in the methodology part, the terrain characteristics (slope $S$ and roughness $\sigma_R$) can influence the spreading of the echo pulse shape combined with the RMS width of the receiver impulse response and the RMS laser pulse width.

Here, we use three parameters: $\sigma_{SR}$, the RMS pulse width of the echo ICESat laser signal (nanoseconds); $\sigma_b = 0.425\text{FWHM}$ ∼ 1.7, the RMS width of the receiver signal. In Table 3, we list the attributes of culled out waveforms in both regions from February 2003 to November 2004.

<table>
<thead>
<tr>
<th>Glaciers</th>
<th>Mountainous area</th>
</tr>
</thead>
<tbody>
<tr>
<td>nmodes</td>
<td>4.8</td>
</tr>
<tr>
<td>W (m)</td>
<td>47.9</td>
</tr>
</tbody>
</table>

Table 2. The waveform shift comparison between ablation zone and accumulation zone in Guila and Guren glaciers. The nmodes and W are chosen in the analysis: nmodes are the number of peaks in each return pulse and W is the width of full waveform (from the beginning to the end).

Table 3. The attributes of culled out waveforms in both regions from February 2003 to November 2004.
impulse response (nanoseconds); and $\sigma_F = 6$, the transmitted ICESat laser pulse RMS width (nanoseconds).

In the beginning, we fill the gaps of the $\sigma_{SR}$ series by 1D interpolation, caused by attenuation of the filtering procedure. Then, using the simplified expression (Gardner 1992), we set the interpolated $\sigma_{SR}$ as initial estimates to optimize the $\sigma_{SR}$ by nonlinear least squares robust method. According to this process, all the possible singularities, due to $\sigma_{SR}$ being insufficiently large, can be avoided. Therefore, we obtain the final acceptable RMS pulse width of the echo ICESat laser signal, satisfying Eqs. (13) and (14). This can be interpreted as the general relation between the terrain characteristics and returned ICESat laser pulse.

Next, we classify the RMS width observation data into groups based on the slope and roughness, respectively. Specifically, we group the RMS width data into three roughness groups: 0–0.7 m, 0.7–1.0 m, and >1.0 m (see

FIG. 6. (a) Correlation in elevation between ICESat and ASTER GDEM; (b) the histogram of deviations obtained from ASTER GDEM elevation subtracts from ICESat elevation. We divide all the differences into 20 bins; the occupancies of variable deviations in different range can be found in this histogram.

FIG. 7. (a) Slope ($^\circ$) and (b) roughness (m) extracted from ASTER GDEM. The bold gray line is the outline of the four larger glaciers, Guren, Guila, Lisheng, and Bili. The values of the calculated slope and roughness are color scaled gradually from low to high.
According to the range of the slope (0°–60°), the data in each group are further grouped in slope sub-classes at 10° intervals (see Fig. 12), and nearly all the observations fall into the dashed outlined zone. Then all three different-colored RMS width groups show the approximate increasing trend with the majority of slope. Simultaneously, the average of RMS widths in Fig. 12 shows a slightly increasing trend with each group of the majority of RMS roughness in general, and group 1 (obs 0–0.7 m) and group 2 (obs 0.7–1.0 m) show a little drop in the generally increasing trend. The results may be due to the existing extreme values in the groups and the relatively small number of samples (only 8 in group 1 and 78 in group 2) compared with the other groups.

In total, the results indicate that the roughness approximately contributes to increasing RMS widths. And it is also shown that there is a dispersion of the RMS width based on roughness at low total slopes from analysis above. But at steep slope zones, RMS widths are insensitive to roughness and increase logarithmically.
with the slope. In other words, the variation of width caused by roughness is not considerably compared to the change caused by slopes. In general, surface roughness is related to the spreading of the Gaussian pulse. Thus, pulse spreading can be caused by footprint roughness or by footprint-scale sloping. For much of the terrain, the dominant cause of spreading is from roughness rather than from slope (Figs. 11, 12). Furthermore, we use the terrain response function and its variance to calculate the ICESat-derived roughness shown in Fig. 13. Rather large values occur in the ASTER GDEM-derived roughness values. However, there are not such relative

![Image](attachment:image1.png)

**Fig. 10.** The distribution of ASTER GDEM-derived slope and roughness vs width of the full ICESat waveform.

![Image](attachment:image2.png)

**Fig. 11.** RMS width of received pulse contributed by total slope and RMS roughness for the ICESat/GLAS system. Each color line represents the series RMS width for certain RMS roughness. Owing to the RMS roughness, all the relevant RMS width observation values are classified in three categories. Green dots show the roughness values that fall into 0 – 0.7 m, pink dots show the roughness is in 0.7 – 1 m, and red dots show roughness larger than 1 m. The bold curve indicates the mean of each bin (in Fig. 12).
larger anomalies found in the ICESat/GLAS–derived roughness. To quantify the changes between these two roughnesses, we compute the deviation between the ICESat-derived roughness and the ASTER GDEM–derived roughness in Fig. 14a. The results show that the ICESat-derived roughness fluctuates from 0.03 to 1.58 m, compared with the ASTER GDEM–derived roughness that varies between 0.15 and 6.87 m. Over 90% of the absolute deviation of these two roughnesses falls between 0 and 2 m. Simultaneously, the possibility density function of the roughness deviation follows a gamma function very well (see Fig. 14b).

5. Discussion and conclusions

ICESat/GLAS provides global measurements by spaceborne laser altimetry. It determined elevation of the surface based on measuring the round-trip duration of flight of a laser pulse from the satellite to the ground. The received pulse is spreading in time, including the fluctuation of the distances between the laser beam and the surface characteristics. According to the first 90-day calibration and validation, the accuracy of ICESat/GLAS is initially ensured. Then a number of studies for diverse ICESat/GLAS applications confirmed this accuracy (Lefsky and Harding 2005; Yi et al. 2005; Kwok et al. 2006; Harding and Carabajal 2005). Here, we studied the extraction of parameters of waveforms measured by the ICESat/GLAS laser system, and the examples of these parameters are given over mountainous and glacial terrain. Furthermore, we compare these full-waveform parameters derived from ICESat/GLAS (number of modes and width of the full waveform and elevation) with surface features derived from ASTER GDEM (roughness, slope, and elevation). Through this analysis, the response of the laser waveform to morphology has been studied for our region of interest. In terms of descriptive statistics, the relationship between ICESat/GLAS waveform attributes and surface parameters from ASTER GDEM is investigated as well. Compared with other global elevation products, ICESat/GLAS has a moderate footprint and wavelength to support precise elevation monitoring (Zwally et al. 2002). The moderate footprint especially indicates that the illuminated area of the laser pulse can be one spot and reduce the slope
influence on elevation, unlike the kilometer-scale footprint of radar altimetry (Abshire et al. 2005). Next, the link between the theoretical influence of surface slope and roughness on the ICESat waveform parameters is quantified. Consequently, the received laser shape can be described by the contribution of terrain characteristics.

In our study area, roughness and slope of the surface can contribute from several meters to even several tens of meters to the pulse shape. These results confirm that spaceborne laser altimetry is potentially efficient to describe the irregularities of terrain over high altitude mountainous areas. This means that satellite full-waveform laser altimetry is a new capable way to derive basic elements (the terrain irregularities) of the energy exchange process in the ABL of remote areas in the Tibetan Plateau and similar mountainous regions on Earth.

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