Ice loss in the Northeastern Tibetan Plateau permafrost as seen by 16 years of ESA SAR missions

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Abstract

InSAR time series of surface deformation from 16 years of Envisat (2003-2011) and Sentinel-1 (2014-2019) ESA satellite radar measurements have been constructed to characterise spatial and temporal dynamics of ground deformation over an 80,000 km² area, in the permafrost of the northeastern Tibetan Plateau. The regional deformation maps encompass various types of periglacial landforms and show that seasonal thaw effects are controlled by the sediment type and local topography. High seasonal ground movements are concentrated on shallow slopes and poor-drainage areas in unconsolidated, frost-susceptible and fine-grained sediments within glacier outwash plains, braided stream plains, alluvial deposits or floodplains. Fast subsidence due to thaw settlement takes place during June/July while frost heave is intense during December/January when two-sided freezing of pore water under pressure causes prolonged ice segregation near the permafrost table. The analysis reveals pervasive subsidence of the ground of up to ~ 2 cm/yr, and increasing by a factor of 2 to 5 from 2003 to today, in high-relief and well-drained areas. The findings suggest that seasonal thaw increasingly affects ice-rich layers at the permafrost table, as well as high-rates of widespread mass movements of non-consolidated sediments, the latter amplified by an increase of effects from frost heave/thaw settlement.

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1. Introduction

Tibet is not simply a flat and elevated plateau but alternates mountain ranges with wide internally-drained plains, where tectonic activity, erosion, and sedimentation are interacting to reshape the Earth's surface [1, 2]. Above $\sim 3500\text{m}$ -height, where the mean ground surface temperature is below 0° , permafrost is present in the ground [e.g., 3, 4, 5] and excess ice in frost-susceptible materials of the active layer thaws during summer and fall months and heaves in winter to spring months, resulting in cyclic surface movements [e.g. 6]. Evidence of frost-related processes is widespread in the Tibetan landscape but varies depending on the relief [1]. On slopes, water is drained and the cover of allochthonous material is reworked by mass wasting processes such as those driven by glacial/fluvial activity, deep-seated creep of frozen materials, solifluction, or debris flows. On lowlands and undrained regions, polygonal cracks, thermokarst lakes and ice-cored mounds attest to the ice-rich layers near the permafrost table.

Recent studies, mainly at local scales of a few km^2 , have highlighted the potential of multitemporal radar interferometry (MT-InSAR) to detect and monitor the seasonal ground displacements attributed to the freeze and thaw cycles of the water stored in the shallow layers of the ground [e.g., 7, 8, 9, 10, 11, 12, 13]. The recent launch of Sentinel-1 satellites with systematic acquisitions in ascending and descending orbits and with a shorter revisit time (down to 6 days) allows scientists to investigate in much more detail the seasonal and inter-annual deformation response of the frozen ground to climate forcing at regional scale.

The impacts that permafrost has on engineering constructions, infrastructure stability (e.g. along the Qinghai-Tibet Plateau Engineering Corridor [e.g., 14, 15, 16, 17, 18, 19]) and on gravitational processes that can pose a hazard to communities [e.g., 20, 21, 12] makes the understanding of how its thermal regime

28 responds to temperature increase very relevant. Moreover, little is known about
29 the effects of permafrost degradation on hydrological processes, moisture and
30 related ecosystem or carbon dynamics [e.g., 22, 23, 24]. Due to the complex
31 relationship between climate and permafrost [e.g., 25], it is therefore essential
32 to provide continuous measurements of surface displacements at large scale and
33 over a long-time period, indicative of the permafrost dynamics.

34 Here, we conduct a large-scale MT-InSAR analysis in the northeastern part
35 of the Tibetan Plateau using both Envisat (2003-2011) and Sentinel-1 (2014-
36 2019) C-band SAR data (Fig. 1) to measure the response of the permafrost's re-
37 lated ground deformation to atmospheric temperature increase over Tibet [e.g.,
38 26, 27, 28].

39 The study area is located between the North Qaidam thrust system, in the
40 south, and the Qilian Nan Shan thrust belt, in the north (Fig. 1). In this
41 arid and remote part of the Tibetan Plateau with elevation ranging between
42 1600 m and 5700 m, the mean annual ground surface temperature is between
43 0 and -5° , the annual precipitation between 50 mm and 400 mm [27, 29] and
44 large-scale coarse models [3, 4, 5] predict cold and continuous permafrost with
45 a relatively thin active layer between 1 and 1.5 m. Investigations of permafrost
46 in the Qilian Mountains of the Heihe River Basin, show that the lowest obser-
47 vation of permafrost is at 3650 m and 3400 m for the south and north facing
48 slopes, respectively, and that the active layer thickness (ALT) can be strongly
49 variable [e.g., 27].

50 In this analysis, the high-resolution data set provides large-scale measure-
51 ments in both ascending and descending geometries from 2003 to 2019 that
52 allow us to analyse the deformation signals for different topographic, geomor-
53 phologic, lithological and hydrological settings and advance our understanding
54 of the dynamics of periglacial surface processes, in an area where deployment
55 of in-situ traditional investigation methods is difficult.

Figure 1: Surface traces of the 4 overlapping Envisat descending tracks, the Sentinel-1 descending and the two Sentinel-1 ascending tracks in the northeastern part of the Tibetan plateau superimposed on the SRTM 30 m hill-shaded Digital Elevation Model [30] and Permafrost Zone Index (PZI) [3], ie. the probability of permafrost occurrence. Dashed-rectangles A and B delimit the two enlarged areas in the following analysis (Figs. 5, 6). Black lines are regional active faults [31]. Bottom right inset: regional setting and location of the study area (red box).

2. Method

2.1. InSAR data processing

We process the complete Envisat data archive along four ~ 300 -km-long and 100-km-wide overlapping descending tracks (90, 319, 047, and 276) between 2003 and 2011, as well as three Sentinel-1 tracks (004, 172, 099) acquired in interferometric wide-swath mode (~ 250 km-width) from 2014 to 2019, with the New Small Baselines Subset (NSBAS) processing chain [32, 33], which includes routines of the ROIPAC software [35] (Fig. S1). Extensive details on the processing of the Envisat data can be found in Daout et al., 2019 [31]. Here, we provide information about the Sentinel-1 data processing (Fig. 2), which follows a similar approach until the unwrapping step (Fig. 2c). However, fine registration based on the enhanced spectral diversity method as explained in Grandin 2015 [36], is also performed to the Sentinel-1 radar images. To unwrap both short and long-baseline interferograms (Fig. S1), NSBAS allows a series of cor-

Figure 2: Illustration of the processing strategy adapted for this large mountainous area with examples from Sentinel-1 descending track 004 around the rectangle B in Fig. 1. Interferograms are processed with the New Small Baselines Subset (NSBAS) processing chain [32, 33]. (a). Before the critical step of phase unwrapping, we apply empirical phase-elevation and ramp corrections in order to reduce the variability of the phase (b). After phase multi-looking, filtering and unwrapping (c), the previous corrections are re-introduced (d) to construct the cumulative time series of the complete interferometric signal referred to the first date (e). Cumulative LOS displacements are then corrected from predictive ERA-5 atmospheric models from ECMWF [34] in order to remove the tropospheric delays (f). After correction, corrected cumulative displacement time series are decomposed into a seasonal and long-term trend (g). Cropped interferograms are displayed in radar geometry.

reactions prior to phase unwrapping, thus significantly improving unwrapping over large areas [e.g., 37, 33, 31]. An empirical phase-elevation correction on the wrapped phase is applied to all interferograms to reduce the variability of phase. This step already reveals the signal associated with freeze-thaw cycles of the active layer in the wrapped interferograms. The heave and thaw settlement deformations of the active layer are usually associated with reversible and high-fringe rates on localised and sharp boundary areas (Fig. S2). The signal is usually spatially asymmetric, with higher gradients on one side of the basin from the other (Fig. S2). The asymmetry of the phase is a clear challenge for unwrapping as it easily creates unwrapping errors if the unwrapping path goes through the high gradients, where the signal is often subjected to aliasing, and, therefore, requires careful processing [9]. To avoid unwrapping propagating into incoherent areas (e.g snow-capped mountain ridges or areas of very high fringe rates linked for instance to permafrost’s related processes), the unwrapping is performed by starting from a coherent reference pixel, and expanding progressively into areas of lower coherence [37].

After unwrapping, we re-introduce the empirical correction previously removed (in step Fig. 2b) from each interferogram (Fig. 2d), refer all interferograms to a low-elevation area within the Qaidam basin, and construct the cumulative time series, referred to the first acquisition date, on a pixel-by-pixel basis [33] (Fig. 2e). The calibration point, at a low elevation within the Qaidam basin, is selected in an area with high coherence, low phase gradient and where low atmospheric turbulence or deformation is expected. Misclosure of the unwrapped interferogram network is then computed and interferograms with high residuals are re-unwrapped with high priority bridges set manually by visual inspection of the interferograms. Final time series results are masked from pixels with low misfits and elevations lower than 3400 m, where no permafrost is expected. Cumulative LOS displacements ($\text{LOS}(t)$) are, afterwards corrected for tropospheric effects using the ERA-5 atmospheric models from ECMWF [34] (Fig. 2f). ERA-5 provides estimates of temperature and water vapour partial pressure as a function of the geopotential height every hour at 137 pressure lev-

els on a 30 km grid from 1979 to present. Integrated path delays at acquisition
 times, computed from both dry and wet delay contributions are derived at ERA-
 5 points encompassing a SAR scene, using vertical profiles of these variables [34].
 The delay is then mapped onto the radar scene using a Digital Elevation Model
 (DEM) and a bilinear interpolation in the horizontal direction on the vertical
 profiles. Jointly with the atmospheric model, a best-fitting cubical ramp is also
 removed from the cumulative LOS displacement in order to tight the correction
 to the LOS displacements and refer all maps to the same low-elevation area
 within the Qaidam basin. This step also aims at removing the 4-6 mm/yr of
 long-wavelength tectonic convergence, between the Qaidam Basin and the north
 Qilian Shan, associated with the distributed slow inter-earthquake deep-seated
 loading of the thrusts in this area [38, 31]. The corrected displacement time
 series are then decomposed into a seasonal and long-term trend (Fig. 2g), such
 as $LOS(t) = V * t + A * \cos(w * t + \phi)$, where V , A and ϕ are the long-term
 rates, the amplitude and the timing of the seasonal displacements, respectively.

2.2. ERA-5 atmospheric corrections

To validate the approach and quantify the effects of the predictive ERA-5
 corrections on our measurements, we compare the time series with tropospheric
 corrections with time series that have not been corrected with ERA-5. Without
 tropospheric corrections, two seasonal peaks between September and December
 and between January and February dominate the multi-annual LOS displace-
 ments, relative to a stable low-elevation point in the Qaidam basin, for pixels
 with seasonal delays higher than 3.5 mm (Fig. 3a,c). In addition, the amplitude
 of the uncorrected seasonal signal show a clear correlation with topography at
 a large-scale, with highest values at higher elevations (Fig. 3a). This behaviour
 reflects the long wavelength and spatially-correlated signal associated with the
 layering of the troposphere [12]. One could also note that, contrary to sea-
 sonal ground deformation, the sign of the seasonal delays associated to change
 of atmospheric conditions is dependent on the reference point's elevation. For a
 low elevation calibration point, atmospheric delays will increase with elevation

131 in January, whereas, for a high-elevation calibration point, atmospheric delays
 132 will increase with decreasing elevation in July. A comparison of the seasonal
 133 delays with two reference points at two different elevations, therefore, provides
 134 a first estimation of the tropospheric delays (Fig. S3). After ERA-5 corrections
 135 of the atmospheric prediction (Fig. 3b,d), higher seasonal amplitudes are con-
 136 centrated on specific areas with no correlation with mountain ridges and with
 137 an homogeneous timing of maximum delays, which draws a normal distribution
 138 around November. This highlights the removal of the previously dominating,
 139 likely atmospheric seasonal pattern with timing of maximum delays in January-
 140 February.

141 **3. Results**

142 Four descending Envisat tracks made of 102 sporadic acquisitions, as well
 143 as two ascending and one descending Sentinel-1 tracks, with a total of 248
 144 acquisitions, illuminate at regional scale the seasonal ground displacements from
 145 2003 to 2019 over an 80,000 km² area of northeastern Tibet (Fig. 4).

146 The good spatial agreement in map view between the three independent
 147 datasets (Fig. 4) and the similarity of the LOS amplitude and velocity profiles
 148 in the overlapping areas of the two Sentinel-1 ascending tracks 172 and 099
 149 (Fig. S4), as well as in the overlapping areas of the two Envisat descending
 150 tracks 319 and 047 (Fig. S5), confirm the reliability of the results. Histograms
 151 of differences reflect measurement errors and show normal distributions cen-
 152 tered on zero with standard deviations of about ~ 0.6 mm and ~ 0.6 mm/yr for
 153 Sentinel-1 LOS amplitude and velocities, respectively (Fig. S4c). For Envisat
 154 data, the histogram of differences for LOS amplitudes shows a bimodal distri-
 155 bution with a standard deviation of ~ 0.9 mm, while for Envisat LOS velocities
 156 uncertainties are about ~ 0.7 mm/yr. Some clear outliers are also visible for En-
 157 visat data reflecting a noisier data set than Sentinel-1 (Fig. S5c). Areas where
 158 the average seasonal amplitudes exceeds 3.5 mm coincide with local valleys and
 159 basins between mountain ranges (Figs. S4b and S5b).

Figure 3: a) Amplitude and timing of the estimated seasonal displacements from descending Sentinel-1 track 004 without ERA-5 correction of the tropospheric delay. b) Same as (a) but after applying ERA-5 correction of the tropospheric delay. c) Histograms of timings of the maximum delays, for pixels with seasonal delays higher than 3.5 mm, without ERA-5 correction showing two seasonal peaks in February and November, associated with the atmosphere⁹ and the freeze/thaw cycles, respectively. d) Histograms of timings of the maximum subsidence after ERA-5 corrections and comparison with the 2 m temperatures extracted from ERA-5 (blue dots) every 6h, at latitude 38°, longitude 96.5°, and an elevation of ~4380 m, and its best-fit seasonal model (red line).

Figure 4: Left: Map of the amplitude of seasonal ground movements in the line-of-sight (LOS) directions for the ascending Sentinel-1 tracks 099 and 172 (top), descending Sentinel-1 track 004 (middle) and Envisat tracks 90, 319, 47 and 276 (bottom). Zones with a seasonal signal greater than 3.5 mm are discontinuously distributed on the flanks of the mountain ranges or at the shore of the internally-drained lakes. Right: Map of the pluri-annual ground velocities. Rates are large and on average away from the satellite.

Seasonal displacement maps are very similar in Sentinel-1 descending (3.1 [0.5–7.7] mm with 95% confidence intervals) or ascending geometries (2.6 [0.5–7.2] mm) suggesting a mainly vertical seasonal deformation.

Timings of maximum subsidence for areas with seasonal amplitude higher than 3.5 mm are homogeneously distributed between the beginning of September and December, highlighting the removal of atmospheric artefacts (Fig. 3b,d, Fig. S3). 2 m air temperatures extracted from ERA-5 models have an annual average of -8°C , a minimum of $\sim -32^{\circ}\text{C}$ in late January and a maximum of $\sim 12^{\circ}\text{C}$ in July (Fig. 3d). Contrasts between the coldest and warmest hours of the day are 20°C . The average surface thawing and freezing onsets are on 22nd of May and 05th of September, respectively. Thaw settlement stops between

171 September and December, thus after the average freezing onset derived from
172 atmospheric models.

173 Measurements also show an anti-correlation between the amplitude of the
174 seasonal signal and the terrain slope, as well as a correlation between the slope
175 and the timing of maximum subsidence, with a later active layer response to
176 the downward thawing front for areas located on steeper slopes (Fig. S6).

177 Negative Sentinel-1 LOS velocities dominate in areas with large seasonal
178 amplitude (>3.5 mm) (Fig. 4), corresponding to a movement away from the
179 satellite due to local subsidence, a horizontal motion away from the satel-
180 lite or a combination of both. Pluri-annual absolute LOS velocities are lower
181 ($-1.1 [-6.4 - 1.7]$ mm/yr) for Envisat descending track 047 than Sentinel-1 de-
182 scending track 004 LOS velocities ($-2.1 [-9.0 - 2.2]$ mm/yr), indicating rates of
183 subsidence two times higher for Sentinel-1 than for Envisat.

184 In the following, we will focus on the interpretation of the results on two
185 specific landscapes types, encompassing lowland (enlargement A in Fig. 1) and
186 high-relief (enlargement B in Fig. 1) permafrost-related processes.

187 4. Interpretations

188 4.1. Lowland deformation

189 The first landscape example is the area surrounding Hala lake extending
190 35 km in the EW direction and 28 km in the NS. The area has elevations rang-
191 ing between 3840 m and 5720 m. The lake is fed by rivers collecting water
192 from the surrounding braided stream plains (Fig. 5a). The southern part is
193 composed of alluvial sediments in the immediate proximity of the lake, where
194 thermokarst phenomena and polygonal cracks attest the presence of ice-rich
195 permafrost (Fig. 5a1). The measured seasonal displacements support previous
196 studies [e.g 6, 21, 9, 12, 11, 29], highlighting a water availability control of the
197 amplitude of the freeze and thaw cycles. Amplitude map (Fig. 5b) and profile
198 (Fig. 5c) show a concentration of large amplitudes on shallow slopes and poor-
199 drainage areas close to the lake shores on frost-susceptible fine-grained lacustrine

Figure 5: Enlargement around the large internally-drained Hala lake in the northeastern part of the study area (rectangle A in Fig. 1). a) Google Earth images and enlarged area a₁. On the northern side, bedrock of the Southern Qilian thrusts is exposed. b) Amplitude map of the seasonal freeze/thaw cycles for descending Sentinel track 004. c) Seasonal amplitudes (red line) and elevation (black line) along profile PP' shown in (b) surrounding the lake in an anti-clockwise direction. Dashed lines indicate the start and end of segments of PP'. d) Map of the long-term ground velocity for descending Sentinel track 004. e) LOS displacement time series for the circles in (d) with best fit sine function with varying timing and a linear trend.

200 or poorly-sorted fluvial sediments. For example, on the north-west side of the
 201 lake, high seasonal displacements follow very sharply the boundary of a debris
 202 fan (see Fig. 5b) and to the south of the lake, larger seasonal displacements are
 203 observed along braided streams. High amplitude concentrations are also found
 204 at the mouth of gullies collecting melt waters, along the river floodplains and on
 205 flat areas at higher elevations. We observe relatively higher amplitudes farther
 206 from the high energy transport area and lower amplitudes closer to the high
 207 energy region. In these areas, lower stream power is expected to transport finer
 208 soil particles, such as silt or fine sands, that have the highest susceptibilities for
 209 ice lensing and to frost heave due to their favourable combination of capillarity
 210 and permeability, and thus ability to retain water within the pores.

211 Several areas with large seasonal amplitudes (Figs. 5b,c) are affected by an
 212 overall trend away from the satellite (red areas in Figs. 5d,e). This is likely
 213 linked to a long-term cumulative thaw settlement due to a gradual loss of wa-
 214 ter/ice volume that is not fully recovered during yearly cycle [7, 9, 10]. The
 215 absence of significant changes in seasonal amplitudes in comparison to the large
 216 pluri-annual trends implies that the thawing of the ice-rich permafrost near the
 217 permafrost table produces little or no changes in the excess water within the
 218 active layer. The latter may be drained away into the lake or might not provide
 219 enough water/ice content in the shallow layers to create changes of heave/thaw
 220 settlement in the next freezing and thawing seasons. In addition, a component
 221 of sediment compaction also contributes to the long-term thaw settlement and
 222 may explain the absence of proportionality between long-term subsidence and
 223 seasonal amplitude increase [e.g 39].

224 At the southern margin of Qilian Shan-Nan Shan thrust belts, in the glacier
 225 fore-fields, deep-seated cohesive viscous creep in perennially frozen, ice-rich de-
 226 bris is indicated by the "porridge-like" appearance of thick morainic deposits
 227 (Figs. 5a, Fig. 7a). The vegetation cover at the surface indicates a rather old age
 228 of the landform with low activity (Fig. 7a). South of those periglacial landforms,
 229 a signal towards the satellite (in blue patterns) follows very sharply the bound-
 230 aries of alluvial fans and the northeastern shore of the lake in both ascending

231 and descending views (Figs. 5c,e, Fig. 7a). If this signal contained a horizon-
 232 tal component, this would indicate a motion up-slope. Since this is unfeasible
 233 and given that ascending and descending geometries have very similar signs of
 234 surface displacements, we interpret the data as a long-term surface rise. This
 235 is likely associated with an increase of water or ice volume within areas charac-
 236 terised by high water accumulation, fed by increase in precipitation, permafrost
 237 thawing, or glacier’s retreat at higher elevations [28].

238 4.2. High-relief deformation

239 The high-relief Paleozoic range is bounded, in the north, by a glacier out-
 240 wash plain and, in the south, by alluvial and lacustrine quaternary deposits
 241 (Fig. 6a). The elevation ranges between 3600 m and 5650 m. The measurements
 242 underline again the control exerted on the magnitude of the freeze/thaw cycles
 243 by the geological material and topographic effects. The higher seasonal ampli-
 244 tudes are mainly concentrated on the unconsolidated sediments in the glaciers
 245 outwash plain, in the alluvial deposits or on the floodplains (Fig. 6b). However,
 246 a dependency on slope orientation of the long-term rates is visible at large-
 247 scale along the various valleys, which is linked to the along slope difference of
 248 horizontal velocities induced by the slow downslope, gravitational, mass wast-
 249 ing processes (Fig. 6c). A large-scale horizontal motion is highlighted by the
 250 correlation between the observed LOS velocity and the terrain slope direction
 251 (Fig. 6d).

252 Long-term cohesive deformation of thick frozen debris masses is visible in
 253 glacier fore-fields, north of the enlarged mountain (Fig. 6a, Fig. 7b). In many
 254 cases, they are characterised by thermokarst ponds and proglacial lake forma-
 255 tion, which emphasise the recent effects of increasing temperature trends on
 256 surface and subsurface ice in the region [e.g 26, 27, 28]. Geomorphological
 257 markers and LOS surface displacements suggest slow gravitational deformation
 258 of a few cm/yr or less of the ice-rich perennially frozen, ice-cored morainic ma-
 259 terials as well as of peripheral glacier ice (Fig. 7c), in agreement with expected
 260 thin and slow flowing cold glaciers in dry and high-elevated continental climatic

261 conditions [40].

262 Comparison of time series (Fig. 6e) reveals a net increase of the displacement
263 rates, the Sentinel-1 velocities over the period 2014-2019, being two to five
264 times higher than those measured with Envisat data over the period 2003-2011
265 (Fig. 6e). In the floodplain, an increase of the seasonal amplitude is visible on
266 some time series, highlighting an increase of water content in a saturated active
267 layer and/or an increase of the ALT itself (Fig. 6d). However, a regional increase
268 of inter-annual seasonal amplitude is not measured. The fact that the amplitude
269 increase is only observed on low-angle slope pixels in the floodplains emphasises
270 the clear control on the deformation exerted by the drainage conditions of the
271 active layer. Anomalies are also observed, more particularly in summer 2018,
272 where displacement time series indicate larger subsidence movements (Fig. 6e).

273 On the southern side of the range, the slope, away from the range, is on the
274 whole much shallower ($\sim 3\%$ slope) than the transversal valley flanks ($\sim 12\%$
275 slope) separating the individual ridges and oriented roughly NE-SW. Decom-
276 position of both ascending and descending LOS velocities into a vertical and
277 east-west ($N110^\circ E$) horizontal velocities is possible, here, as the horizontal ve-
278 locity gradients are expected to be roughly oriented in a $N110^\circ E$ direction,
279 which is the most favourable orientation for detection by the satellite. Results
280 show that for pixels with seasonal freeze/thaw cycles higher than 3.5 mm, the
281 $N110^\circ E$ long-term velocities range between -26 and 18 mm/yr, while the ver-
282 tical long-term velocities are within -24 and 7.5 mm/yr. The map of Fig. 8a
283 shows pervasive subsidence affecting all valley flanks and sedimentary basins in
284 the whole area, in comparison to surrounding stable bedrock. The east-west
285 motion is clearly localised on the steeper slopes of the transversal ridges, il-
286 luminating the downslope sliding of valley flanks (Fig. 8b,c). The sign of the
287 horizontal signal correlates with the orientation of the slope and with solifluction
288 sheet patterns visible on the optical images, from our analysis (Fig. 8b,d). Such
289 geomorphological landforms indicate shallow mass movement of thin detached
290 layers of homogenous and fine-grained material with the slipping plane corre-
291 sponding to one of the fissures created by the ice [42, 6, 21]. The anti-correlation

Figure 6: Enlargement around a high-relief range in the western part of the study area (rectangle B in Fig. 1). a) Google Earth imagery. b) Amplitude of the seasonal freeze/thaw cycles from descending Sentinel track 004. c) Long-term ground velocity from track 004. d) Scatter plots showing the correlation in enlargement area between the LOS velocity of track 004 and terrain slope along radar range direction (left) and the amplitude of seasonal movements (right) for pixels with slope higher than 3%. e) LOS displacement time series for the two circles in c) derived from both Envisat (2003-2011) and Sentinel-1 (2014-2019) data with best fit sine function with varying timing and a linear trend, emphasising the increase of both pluri-annual rates and seasonal amplitudes from 2003 to today.

Figure 7: Enlargement at the bottom of Qilian Shan-Nan Shan thrust belts, north of the Hala lake of the Fig. 5 (a) and at the northern flank of the mountains of Fig. 6 at the fore-field of the glaciers (b). a) Google Earth imagery side-by-side of the LOS velocities of the descending track 004 (top-right) and the ascending track 099 (bottom-right). Ascending and descending velocities are very similar indicating a mainly vertical deformation. b) Google Earth imagery side-by-side of the LOS velocities of the descending track 004 (top-right) and the ascending track 172 (bottom-right). Ascending and descending velocities are almost of opposite signs indicating a mainly horizontal deformation. c) Two time series of the descending LOS surface displacements marked as circles on the velocity map in (b). Creep rates of both glaciers and frozen debris landforms are of the order of the cm/yr or less.

Figure 8: Vertical (a) and east-west (N110°E) (b) decomposition of the velocity of the area B in Fig. 1 obtained from inversion of the two ascending and the descending velocities of the Sentinel-1 data. c) N110°E velocities (eastern motion in blue, western motion in red) along profiles AA' shown in (b) illuminating the downslope sliding of solifluction sheets within the individual transversal ridges of valley flanks. d) Enlargement of the east-west velocity, side-by-side to a Google Earth imagery, on individual ridges showing the correlation of the signal with solifluction sheet patterns.

292 between LOS velocity and the amplitude of the seasonal movements (Fig. 6d)
 293 highlights the control played by the water concentration and the amplitude of
 294 the freeze/thaw cycles on the intensity of the mass movements [6]. Here again,
 295 the soil properties and water content in the layers of loose sediments appear to
 296 directly influence the rates of the horizontal ground deformations.

297 **5. Discussions and conclusion**

298 *5.1. Permafrost's freeze/thaw cycles monitoring with InSAR*

299 With an unprecedented spatial and temporal sampling, MT-InSAR data pro-
 300 vide insights into the response of the permafrost to temperature increase over
 301 Tibet. The higher spatial continuity of the Sentinel-1 results in comparison to
 302 Envisat highlights the significant benefits and potential of the newly launched
 303 Sentinel-1 data to drastically decrease temporal decorrelation affecting the in-
 304 terferometric signal for fast-moving surface processes and to monitor the spatial
 305 and inter-annual variations of deformation at high-resolution.

306 Moisture cycle is likely the main source of error when measuring the freeze-
 307 thaw cycle with InSAR [43, 44]. The increase of moisture during a thaw event
 308 decreases the penetration of the electromagnetic wave into the ground by a fac-
 309 tor that can reach 10-20% of the wavelength, and therefore under-estimate the
 310 amplitude of the measured thaw settlements. Large fluctuations of soil water
 311 content in coarse or high bulk density material with low organic matter [e.g
 312 45] will, therefore, increase this effect. In addition, the asymmetry of the mois-
 313 ture deformation, with fast soil wetting and slow soil drying, creates systematic
 314 decorrelation or possible unwrapping errors of the interferometric signal during
 315 wet events, which may lead to time series analysis biases. In order to explore
 316 the stability of the time series analysis, and investigate the possible presence
 317 of biases, we perform time series inversions with various interferometric net-
 318 works (Fig. S7, S8). The analysis show good robustness of the linear velocity
 319 retrieval after changing the interferometric network and, more particularly, af-
 320 ter removing short-temporal baselines interferograms. Major differences only

321 arise when performing a time series analysis with an interferometric network
 322 constructed only with short-baselines interferograms connecting two successive
 323 acquisitions (Fig. S7, S8). This confirms the analysis of Andari et al., 2020 [46]
 324 showing that short baseline interferograms are the least reliable for deformation
 325 retrieval. Biases induced by the integration of short temporal baselines are not
 326 only observed in high-moisture regions, but also on the mountain peaks, likely
 327 linked to snow and to systematic decorrelations or unwrapping errors induced
 328 by the abrupt changes of ground properties.

329 The measurements presented in this study show amplitudes and timings
 330 of maximum subsidence that vary spatially and inter-annually, indicating that
 331 other factors than air temperature are important in determining the freeze/thaw
 332 state of the active layer [39, 21, 47]. To emphasise this point, we provide in
 333 Fig. S9, time series of surface displacements modeled with the Stefan equations,
 334 where the deformation is proportional to the square root of the cumulative de-
 335 gree day of thawing or freezing extracted from the ERA-5 measurements of
 336 Fig. 3d. A degree-day model imposes uniform heave and thaw onsets on the
 337 whole area based on one single temperature, and, therefore, does not approx-
 338 imate correctly the data at a regional scale. In addition, on steep ground as
 339 the expansion is normal to the slope and the collapse is more or less vertical,
 340 the LOS surface displacements deviate from the Stefan prediction. Care must,
 341 therefore, be taken when prescribing the evolution of the deformation in the time
 342 series analysis with a degree-day model because of the spatial variability of the
 343 freezing and thawing onset lags. Gridded temperatures records from reanalyses
 344 or remote sensing data could be used to better capture the spatial variability
 345 of the ground temperature. However, and as shown from the time series of
 346 Figs. 5e, 6e, maps of the timing of maximum seasonal displacements (Fig. 3b),
 347 or observed correlations between seasonal amplitudes, local slope, and timing
 348 of maximum seasonal subsidence (Fig. S6), there is a spatial variability of sea-
 349 sonal delays over the same regions that cannot solely be explained by changes in
 350 surface temperatures. In addition, the large DEM errors found by some studies
 351 using Sentinel-1 data, which have, however, a good baseline control, reflect the

trade-offs of the deformation model with other parameters of the inversion, such as a DEM error term. The analysis of this paper shows that a simple sinusoidal model allows to map and interpret the lateral variability of maximum seasonal deformation and separate seasonal signals associated with tropospheric delays from seasonal ground deformation (Fig. 3).

5.2. Seasonal effects

The analysis highlights a late point of change from subsidence to heave at the end of the year (Fig. 3b,d). To further illustrate this, we compute the de-trended and normalised seasonal signal over a 12 months period, $(LOS(t) - V * t)/A$, where $LOS(t)$ are the time series of LOS displacements for each pixel and A , V , are the inverted amplitudes and velocities previously obtained (Fig. 9). During the frozen period between February to June, the surface remains stable. In May/June (once surface snow disappears), as surface temperatures rapidly increase above 0°C, large vertical subsurface temperature gradients and heat fluxes build up and rapid one-sided downward thawing in the active layer with corresponding thaw settlement takes place (Figs. 5e, 6e, 9). Around September/October, maximum thaw depths are reached and little to no more phase change occurs. Then, the two-sided freezing starts in November/December and lasts until February/March. The process is fast downwards from the surface because of high temperature gradients and correspondingly high heat losses, but slow upwards from the permafrost table with low temperature gradients and heat fluxes. When the downward penetrating freezing front approaches the upward migrating freezing front at depths, pore water in the still unfrozen zone between the two freezing fronts becomes pressurised, stabilising and prolonging the phase change with corresponding latent heat and extended zero-curtain effects. While mostly interstitial ice with little or no expansion of total soil volume had formed in the top soil during early fall, ice segregation takes place during December/January at depth (near the permafrost table), involving volume expansion and frost heave. Eventually, as the remaining unfrozen water between the two freezing fronts gets under pressure and lowers its freezing temperature,

the process of ice segregation and complete freezing is further delayed. Comparison between the high-relief region and the lowland area in Fig. 9, also indicate a more delayed frost heave for the high relief area. Around the Hala lake, the frozen period is between February and June, while the cumulative maximum ground settlement is reached in October. For the high-relief area, the frozen period is shrunk compared to the lowland region between April and June, and the maximum settlement is around November.

In comparison, Liu et al., 2017 [48] have shown with in-situ observations of the active-layer properties in the Tien Shan that active-layer freezing at depth, near the permafrost table, can be delayed by several months with respect to the onset of freezing from the surface and takes place around the end of the year. Boreholes measurements in the northern and more vegetated part of the Qilian Shan [29], also show a downward penetration of the freezing fronts in the active layer of about 1.2 m in November-December, in agreement with the timing of lowest surface elevation measured in this study (Figs. 3d, 9). Several studies also highlight the presence of ice-rich layers in the ground, and more particularly, of segregated ice for silty clay and sandy gravel in valley-bottom geomorphological settings [e.g 14, 15]. Permafrost in shallow areas in the studied region can reasonably be assumed to be similarly rich in ice, especially at the permafrost table, in the presence of fluvial or lacustrine frost-susceptible fine or poorly-sorted materials (silt, sand, sandy gravels).

In this northeastern part of the Tibetan Plateau, large-scale coarse models predict cold and continuous permafrost [3, 4, 5]. However, the analysis reveals that large (>3.5 mm) frost heave and thaw settlements are only concentrated on undrained basins or on south-exposed flanks covered by sedimentary deposits (Fig. 4). We suggest that the spatial localisation of large freeze/thaw cycles reflects the heterogeneous distribution of the excess water in frost-susceptible materials that is controlled by the soil property and by the contrasts between slopes and nearly flat areas of the Tibetan Plateau. Permafrost's active layer freezing and thawing is widespread but frost heave and thaw subsidence magnitude depends on excess subsurface ice. In this dry and desertic continental area,

Figure 9: a) Timings of the freeze/thaw cycles for the lowland (left, Fig. 5) and high-relief (right, Fig. 6) areas, for Sentinel-1 descending track 004, masked from pixels with amplitude of seasonal displacement inferior to 3.5 mm. b) Normalised seasonal evolution at data acquisition times (filled circle are median values for all pixels while thin vertical bars are standard deviation for all pixels) wrapped between January and December over the lowland (left) and high relief (right) areas, for pixels with seasonal displacements higher than 3.5 mm. The seasonal models are in red dashed lines and estimates of frozen and thawed periods are the blue and red shaded regions, respectively. During the frozen period (1), the surface remains more or less stable until the onset of thaw/subsidence (2) from the surface with positive air temperatures in about early June. During the thawed period (3), maximum thaw depths are reached and little to no more phase change occurs. The two-sided freezing (4) starts in November/December and lasts until February/March, when maximum heave rates are observed from ice segregation at depth.

large seasonal amplitudes detected with InSAR reflect soils with high subsurface water contents or where the deepening of the thawing front reaches ice-rich layers. The findings also suggest that the seasonal thawing front deepening at the permafrost table is occurring over a widespread region, with local variations depending on snow cover, hydrology, micro-topography or vegetation cover, and that the delay observed in active-layer freezing is not only a localised occurrence.

5.3. Long-term ground subsidence

The analysis shows significant long-term subsidence trends over areas that undergo large seasonal heave and thaw settlement at a regional scale (Figs. 4, 8a). If this subsidence were linked to an increase of ALT alone, we would observe simultaneously a regional increase of inter-annual seasonal amplitude at regional case, which is not the case else than few areas. In light of the large seasonal thaw-subsidence delayed at the end of the year, of the high long-term subsidence rates, and of the increase in long-term trends from Envisat to Sentinel-1 measurement, we infer that the seasonal thawing front is likely reaching the ice-rich layers at the bottom of the active layer and link the observed pervasive subsidence rates to an ice-loss in the Northeastern Tibetan Plateau permafrost. This phenomenon, also called isotropic thaw subsidence, has been documented by other authors in Alaska [49, 50], who highlighted limited correlation between increasing air temperature and ALT above ice-rich permafrost, but widespread and important ($\sim$ cm/yr) subsidence of the ground surface associated with warmer winter, increased snow depth, and deeper thaw penetration into the ice-rich transient layers in summer. Permafrost degradation rate and ice volume loss cannot be inferred from the measures subsidence rates as it depends on many factors such as ice content or soil compaction. However, this MT-InSAR study highlights the benefits of such regional measurements to map the lateral variability of the ground deformation dynamic. More particularly, it shows that the pluri-annual subsidence and thaw settlement of the ground is stronger for regions where relief is higher, as there, the water is well-drained and the loss of ice/water volume causes ground subsidence at rates of up to

2 cm/yr (Fig. 8a). Conversely, measurements reveal that at low-elevation and internally-drained areas, the deformation is more complex due to the combination of temperature and moisture increase and that local measurements must be interpreted carefully. In lowland regions, seasonal amplitudes may increase and/or uplift may happen as the excess meltwater pools and triggers an increase of segregation ice near the permafrost table (Fig. 5d, Fig. 6c).

The complexity of the data emphasises that caution is required in interpreting the signal in the context of atmospheric temperature increase over Tibet due to the multiple interactions between temperature increase, drainage properties, vegetation and moisture capacity of the terrain [25]. Care must also be taken when drawing direct links between deformation amplitude and ALT, as also recently proposed in many studies, due to the spatial variability of the drainage and water capability of the soil of any thickness. Our findings show that a key parameter for permafrost related deformation seems to be the slope, which controls the distribution and the evolution of the excess water and may now, therefore, be included in large-scale models of permafrost distribution and degradation.

5.4. *Relation with the long-term evolution of the Tibetan Plateau*

The flatness of the Tibetan plateau has been the center of numerous debates on the long-term mechanisms that smooth out irregularities in topography and crustal thickness [e.g 2]. The importance of surface processes has been sometimes dismissed because of the low erosion rates and low drainage networks associated with the aridity of the area. However, under those climatic conditions, periglacial mass-wasting processes may dominate other erosional mechanisms. Our findings show that in this active and rugged part of the Tibetan plateau, where relief reduction is in competition with active tectonics [2] and where there is very little deep-rooted vegetation to prevent downslope mass-movements, periglacial surface processes, with LOS ground motion of the order of a centimeter per year, are stronger than any tectonic signals that could be expected along the North Qaidam thrusts or the southern part of the Qilian

473 Shan [38]. This result shows that it is critical to delineate the areas affected by
474 periglacial related deformation to avoid erroneous interpretations of the tectonic
475 loading along faults in cold regions. It also demonstrates that, in addition to
476 low energy rivers transport and the internal drainage characteristic of this part
477 of the Tibetan Plateau that disconnect inter-mountains catchment basins form
478 the foreland outlets, periglacial processes, such as solifluction, participate in re-
479 lief smoothing by moving loose sediments from the high ranges to the lowland
480 regions.

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487 Data Availability

488 The maps of Fig. 3 are provided as KMZ in the supplements.

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