

Geophysical signature of the Tunnunik impact structure, Northwest Territories, Canada

Yoann Quesnel, W. Zylberman, P. Rochette, Minoru Uehara, Jérôme

Gattacceca, G R Osinski, P. Dussouillez, C. Lepaulard, C. Champollion

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Quesnel, Y. et al. (2020), Geophysical Signature of the Tunnunik Impact Structure, Northwest 1 Territories, Canada. Meteoritics and Planetary Science, 1-16, doi:10.1111/maps.13447 2 3 Geophysical Signature of the Tunnunik Impact Structure, 4 Northwest Territories, Canada 5 6 Y. Quesnel¹, W. Zylberman^{1,2}, P. Rochette¹, M. Uehara¹, J. Gattacceca¹, G. R. Osinski^{2,3}, 7 Dussouillez, P.¹, Lepaulard, C.¹ and C. Champollion⁴ 8 ¹Aix-Marseille Univ, CNRS, IRD, INRAe, Coll France, CEREGE, Aix-en-Provence, France. 9 ²Institute for Earth and Space Exploration, University of Western Ontario, London, Ontario N6A 10 5B7, Canada 11 ³Department of Earth Sciences, University of Western Ontario, London, Ontario N6A 5B7, 12 13 Canada ⁴Geosciences Montpellier, UMR5243, Université de Montpellier – CNRS, Place E. Bataillon, 14 34095 Montpellier, France 15 16 17 Corresponding author: Yoann Quesnel (<u>guesnel@cerege.fr</u>) 18 **Key Points:** 19 A 3 mGal negative gravity anomaly and a 120 nT positive magnetic anomaly were 20 measured over the center of the Tunnunik impact structure 21 A numerical model constrained by laboratory measurements on rock samples suggests an 22 uplifted magnetic crystalline basement 23

• The fracturing/brecciation extends down to 1 km in depth

26 Abstract

In 2011, the discovery of shatter cones confirmed the 28 km-diameter Tunnunik complex impact 27 structure, Northwest Territories, Canada. This study presents the first results of ground-based 28 electromagnetic, gravimetric and magnetic surveys over this impact structure. Its central area is 29 characterized by a ~10 km wide negative gravity anomaly of about 3 mGal amplitude, roughly 30 corresponding to the area of shatter cones, and associated with a positive magnetic field anomaly 31 32 of ~120 nT amplitude and 3 km wavelength. The latter correlates well with the location of the deepest uplifted strata, an impact-tilted Proterozoic dolomite layer of the Shaler Supergroup 33 exposed near the center of the structure and intruded by dolerite dykes. Locally, electromagnetic 34 field data unveil a conductive superficial formation which corresponds to an 80-100 m thick sand 35 layer covering the impact structure. Based on measurements of magnetic properties of rock 36 samples, we model the source of the magnetic anomaly as the magnetic sediments of the Shaler 37 38 Supergroup combined with a core of uplifted crystalline basement with enhanced magnetization. More classically, the low gravity signature is attributed to a reduction in density measured on the 39 brecciated target rocks and to the isolated sand formations. However, the present-day fractured 40 zone does not extend deeper than ~1 km in our model, indicating a possible 1.5 km of erosion 41 since the time of impact, about 430 Ma ago. 42

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44 Plain Language Summary

This study reveals the geophysical signature of the buried structure of an eroded impact crater, 45 Tunnunik, located in Northwest Territories, Canada. A positive magnetic anomaly was detected 46 at the center, showing the uplift of some deep geological formations and the possible presence of 47 strongly-magnetized basement. A negative gravimetric anomaly is also observed, mostly 48 corresponding to the fracturing/brecciation of the impacted rocks inside the crater. Using 49 numerical models constrained by laboratory measurements on rock samples, the physical 50 properties and geometry of the buried geological formations are estimated. An important 51 implication of this study is the link between the geophysical remains of an impact crater and the 52 post-impact erosion. 53

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INTRODUCTION

With about 190 impact structures confirmed, the Earth presents currently the lowest number of 56 craters of all the terrestrial planets. Indeed, craters on Earth are continuously removed by erosion 57 and hidden by sedimentation, and/or tectonic activity (Grieve 2006; Hergarten and Kenkmann 58 2015). Therefore, geophysical methods are particularly relevant to the study of terrestrial impact 59 craters, as they can reveal buried or eroded structures (Pilkington and Grieve 1992; Gudlaugsson 60 1993; Grieve and Pilkington 1996; Pilkington and Hildebrand 2000; Kjaer et al. 2018). They also 61 help to unveil the physical effects of syn- and post-impact processes which are still not 62 completely understood, particularly for mid-size to large impact structures (Osinski and Pierazzo 63 2013). Moreover, they help to reveal the present-day geometry of the different layers of impact 64 lithologies. 65

The most commonly used geophysical techniques for studying impact craters are gravity and 66 magnetic surveys by satellites, airplanes and/or ground measurements. Indeed potential-field 67 anomalies are often significant over impact structures, even buried or eroded. Small-size 68 (apparent diameter $D_a < 10$ km; D_a being the present-day diameter after erosion of the final crater 69 rim) impact craters are well characterized by negative potential field anomalies (Grieve and 70 Pilkington 1996), while it is not as clear for intermediate-size (10 < D_a < 30 km) complex impact 71 structures. The low gravity signature is due to a reduction in the density of target rocks because 72 of impact-induced fracturing, brecciation and melting due to the compression, excavation and 73 modification stages of the impact event (Pilkington and Grieve 1992; Osinski and Pierazzo 74 2013). Additional minor effects can also contribute to the mass deficiency, such as lower-density 75 76 post-impact sedimentary layers filling relatively fresh craters (Grieve and Pilkington 1996). The lithological and physical changes associated with the impact process can also be modified by 77 post-impact alteration. The observed weak magnetic signature over small craters ($D_a < 10$ km) 78 results from the alteration of the pre-existing regional signals (Cisowski and Fuller 1978; Clark 79 1983; O'Neill and Heine 2005), but may be also influenced by the reduction of the natural 80 remanent magnetization (NRM) through partial or complete shock remagnetization (Gattacceca 81 et al. 2010). Overall, some of the impact induced effects on rock magnetism are still debated, 82 such as the possible preservation of a shock-induced remanent magnetization (SRM; e.g., 83 Cisowski and Fuller 1978; Halls 1979; Pesonen et al. 1992; Gattacceca et al. 2007, 2008, 2010; 84 Tikoo et al. 2015). For larger craters (D_a > 10 km), central high-amplitude and short-wavelength 85 anomalies are observed (Pilkington and Grieve 1992; Morgan and Rebolledo-Vieyra 2013). 86 Their source is usually complex, and can originate from shock metamorphism (e.g., Slate Islands 87 and Charlevoix, see Halls 1979; Robertson and Roy 1979; review in Grieve and Pilkington 88 1996), cooling of impact-melt rocks and/or breccias (e.g., Ries and Morokweng; Pohl et al. 2010; 89 Henkel et al. 2002), structural deformation of target geological units during the modification 90 stage (e.g. Bosumtwi; Ugalde et al. 2007), and/or post-impact hydrothermal processes (e.g., 91 Haughton: Quesnel et al. 2013; Zylberman et al. 2017; Clearwater Lakes: Gattacceca et al. 92 2019). Henkel (1992) and Henkel and Reimold (2002) also suggested oxidation of pre-existing 93 magnetic carrier phase in the strongly fractured rocks, which may explain the reduction of 94 ground magnetic susceptibilities surrounding the central uplift. The origin of magnetic anomalies 95 with a significant negative part in complex and eroded impact structures remains unclear 96 (Pilkington and Grieve 1992; Grieve 2006). All larger craters (D_a > 40 km) show high-amplitude 97 magnetic anomalies at their center (Pilkington and Grieve 1992; Morgan and Rebolledo-Vieyra 98 2013). 99

It is notable that the magnetic and gravimetric anomaly characterization strongly depends on the 100 altitude of the measurements: no signature at satellite or airborne altitude does not mean that 101 there are no contrasts in the upper crust of the studied area. For instance, this effect has been 102 suggested to explain the lack of 'apparent' magnetic signatures of Hellas and Argyre large 103 impact basins on Mars (Acuña et al. 1998), even if the absence of a dynamo during the cooling 104 stage after these events is a better explanation (Langlais et al. 2004; Langlais and Thébault 105 2011). This may also be suggested for most of the magnetic signatures of lunar impact structures: 106 the lowest satellite magnetic field measurements were performed at minimum 30 km of altitude, 107 excluding *de facto* the mapping of small-wavelength magnetized contrasts (Nicholas et al. 2007; 108 Hemingway and Tikoo 2018), while a wealth of gravimetric details were observed by the 109 GRAIL mission (Zuber et al. 2016). On Earth, few large impact structures – including Chicxulub 110 (Gulick et al. 2013) - exhibit a gravimetric anomaly that can be observed by satellite 111

measurements. For mid-size impact structures, the characterization of gravimetric and magnetic 112 anomalies by airplane/helicopter measurements can be possible to get a broad overview of its 113 geophysical signature. However, such airborne surveys are still low-resolution, practically 114 difficult and expensive for studying remote areas. Ground surveys allow acquiring high-115 resolution gravimetric and magnetic data over impact structures in remote areas. also enable 116 sampling to be conducted for further laboratory analyses, to update the geological mapping and 117 to identify possible local geophysical anomalies (e.g., the Haughton crater center; Zylberman et 118 al. 2017). All these data will serve as constraints for modeling the geological (and possibly 119 multiple) sources of the observed gravimetric and magnetic anomalies. The central uplift itself 120 creates a gravimetric anomaly, but on Earth, the erosion level of the impact structure will then 121 play a role. In most cases, the more eroded the crater, the less significant (in wavelength and 122 amplitude) the gravimetric negative anomaly (Pilkington and Grieve 1992). On the other hand, 123 the erosion of a large thickness of post-impact sediments will better unveil density and 124 125 magnetization contrasts linked to pre-, syn- or post-impact processes in the structure.

In this study the first ground geophysical measurements acquired over the recently-discovered mid-size ($D_a \sim 28$ km) eroded Tunnunik impact structure are described, analysed and modeled in terms of geological structure and cratering processes.

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GEOLOGICAL CONTEXT

131 The Tunnunik impact structure is located in the western Canadian Arctic Archipelago (Fig. 1) iust south of the Richard Collinson Inlet ("Tunnunik" in Inuvialuit), on the Prince Albert 132 peninsula, northwestern Victoria Island (72°28'N, 113°58'W). The target sequence is composed 133 of sub-horizontal Cambro-Silurian sedimentary rocks (mostly carbonates) from the Arctic 134 Platform: from oldest to youngest, the Wynniatt (part of the Shaler Supergroup), Mount Phayre, 135 Victoria Island, and Allen Bay formations (Fig. 1; Dewing et al. 2013; Newman and Osinski 136 137 2016). Neoproterozoic diabase dykes also intrude into the Wynniatt Fm. More detailed information on the pre-impact stratigraphic sequence and geological context can be found in 138 139 Dewing et al. (2013). There are also a number of Quaternary sand and gravel formations (not shown in Fig. 1) deposited between 50 and 100 m in altitude throughout the crater interior, and 140 mainly related to the central river valley where some outcrops of the Shaler Supergroup and 141 Mount Phayre formations show a central uplift. By using remote sensing and sample analyses, 142 143 this formation was recently mapped as fluvioglacial deposits (Unit 1 of Choe et al. 2019). Sea shells were found in some outcrops of this formation, suggesting that it could correspond to post-144 glacial marine deltaic deposits, now overwater due to the isostatic rebound. Based on the 145 observation of tilted strata and the distribution of shatter cones, Dewing et al. (2013) described a 146 ~25 km-wide circular feature. Based on detailed mapping and the presence of inward-dipping 147 listric faults out to a radius of 14 km, Osinski et al. (2013) defined a 28 ± 0.5 km apparent crater 148 diameter (D_a). No crater fill breccias and/or melt rocks were mapped, indicating that the impact 149 structure is deeply eroded. Only isolated dykes of polymict impact breccias were observed 150 (Newman and Osinski 2016). Paleomagnetic analyses of these dykes has provided an age for the 151 impact event of 440 ± 10 Ma (Lepaulard et al. 2019). The present topography decreases from 152 about 200 m altitude in the southeast corner to the sea level of the Richard Collinson Inlet in the 153 northwest. This general trend is cut by two major ~N-S oriented rivers - including one in the 154

central part of the structure – and by syn-impact (mainly concentric) and post-impact (mainly in
the SW-NE direction) faults. Therefore, except for the concentric faults, there is no topographical
and geomorphological signature of an impact crater. Possible hydrothermal alteration has also
been reported within the impact structure, but evidence remains sparse and samples are still
under investigation, as it is not always clear if the described alteration is pre- or post-impact
(Marion et al. 2013).



Fig. 1. Geological map of the Tunnunik impact structure, modified from Dewing et al. (2013) and 162 163 Newman and Osinski (2016). Background corresponds to a Map Data ©2015 Google satellite image. Stratigraphy: Shaler Supergroup (Neoproterozoic), Mount Phayre formation (Cambrian), Victoria Island 164 formation (Cambrian/Ordovician), and Thumb Mountain/Allen Bay 165 formation (Upper Ordovician/Silurian). The coordinate system for the geological map is UTM Zone 12 North projection 166 with WGS84 datum, in meters. Upper left: Location of the structure on Victoria Island in the Canadian 167 Arctic, with a geographic coordinate system on a WGS84 datum. Background corresponds to the ArcGIS 168 169 online ESRI Ocean layer.

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METHODS

Ground gravity and magnetic field measurements (total paths of 300 km) were performed within the central part of the Tunnunik impact structure, in an area approximately corresponding to the extent of the Victoria Island Formation (Fig. 2). The wide N-S river on the western side prevented further measurements in that direction, while the large area and a short time in the field in this remote Arctic region also influenced our choice to mainly survey the center of the structure. Thus we obtained a well constrained map of the central part rather than extended profiles. We also conducted local subsurface electromagnetic sounding in order to determine the thickness of quaternary surficial formations that could influence the gravity signature.





The gravity field was measured using a Scintrex CG-5 Autograv gravity meter. Three main profiles in the NNW-SSE, NE-SW and WSW-ENE directions were performed with a mean spacing of 500 m between each measurement point. These profiles cross each other near the

estimated center of the structure. Additional measurements along 5 shorter profiles, with 204 acquisition spacing between 250 and 500 m, completed the central part of the gravity map. A 205 total of 113 independent measurements were acquired. For each point, the gravity field was 206 measured 4 times during 100 seconds at 1 Hz, excluding outliers and outputting the mean value. 207 Instrumental error (i.e. 1 sigma uncertainty), as defined by the standard deviation of the 4 208 successive averages, varied between 0.5 and 16 µGal with a mean value of 6 µGal, depending on 209 wind and soil stability. Depending on the number of acquired measurements, on the local 210 topography of the surveyed area, as well as on the weather conditions, drift varied from 0.5 to 211 2.6 µGal mn⁻¹ (average: 1.2 µGal mn⁻¹). For each station, the X, Y and Z locations were precisely 212 measured using a differential GNSS Trimble R8 system. The base and mobile GNSS stations 213 214 were connected by a standard radio communication and monitored with a controller. Due to the long distances and topography, a secondary radio relay station was used to extend the 215 communication with the base station. With the real-time kinematics (RTK) capability of the 216 system, a precision of ±10 cm in elevation (Z) was achieved. This precision is suitable for 217 gravity accuracy at the 0.1 mGal level as an elevation error of 10 cm produces a 0.03 mGal error 218 in free-air gravity (Featherstone and Dentith 1997). Then the Bouguer correction (0.1 mGal/m, 219 opposite sense to free-air) will reduce this error to 20 µGal, which, combined with the average 6 220 µGal experimental error of the CG-5 instrument, would result in a total error inferior to 30 µGal. 221 This total error is below 1% of the total amplitude of our Bouguer gravity anomaly (see next 222 223 sections). Gravity data were processed using the dGNSS elevation data and a series of standard corrections to remove the Earth tide, drift, latitude, altitude and topographic effects and finally, 224 to obtain the complete Bouguer gravity anomaly value. The final accuracy of the Bouguer 225 anomaly (taking into account instrumental errors, altitude and topography correction accuracies) 226 is 0.1 mGal (see the Supplementary Material for details about gravity data processing, and Figs. 227 S1 and S2; GEOINT 2008; Heiskanen and Moritz 1967; Hwang et al. 2003; Schwiderski 1980; 228 Tamura 1987; Wenzel 1996). 229

230 The variations of the intensity of the geomagnetic field vector (Total Magnetic Intensity, TMI) were mapped using a mobile Geometrics G-858 MagMapper cesium vapor probe fixed at 2 m 231 height on a pole. The internal and external field temporal variations on site were measured using 232 a fixed Geometrics G-856 proton precession base station magnetometer. The use of a base 233 station for diurnal variation of the magnetic field was necessary due to the proximity of the North 234 magnetic pole (1,700 km away): diurnal variations reached several hundreds of nT of amplitude 235 236 during our surveys, potentially completely concealing the crustal signal, which is in the range of 0 to 100 nT in amplitude. The G-856 and G-858 magnetometers have absolute precision of 0.1 237 and 0.01 nT (at 1 Hz sampling rate), respectively. A mean sampling rate of 15 s was used on the 238 field leading to a mean spatial resolution of ~20 m along each survey line. The surveyed area 239 was mainly restricted to the central part within the shatter cone area, except for one ENE-WSW 240 line which extends towards the northeast faults (Fig. 2). During the survey the IGRF predicted 241 average field was about 58140 nT in intensity - exactly in the range of our base station data - 86° 242 and 16° of inclination and declination, respectively. Given the magnetic north pole was 1,700 km 243 away, there was no need to apply the reduction-to-the-pole to our magnetic field data. The 244 magnetic field anomaly was then simply computed by subtracting the TMI value measured at the 245 base station from the TMI value acquired by the mobile magnetometer. We carefully checked 246 that there was no correlation of the anomaly spatial variations with the topography. The poor 247 spatial resolution of the available airborne data (CAN-SCAN project 1965-1976) and magnetic 248 field grids in this area (Ravat et al. 2009) prevents from a good characterization and comparison 249

of the regional influence over the impact structure. Therefore, we first applied a low-pass (100

m) filter to the interpolated data grid to exclude isolated outliers, and then detrended this grid by

a 3rd-order polynomial surface to level the anomaly map (i.e. to remove an eventual large-scale

253 regional crustal field signal).

Local electromagnetic soundings were performed using the Geonics Ltd EM34-3 terrain 254 conductivity meter, which measures the apparent conductivity (sensitivity of 1000 mS.m⁻¹) of the 255 ground by means of a pair of coils. We measured 104 data points to characterize the lateral and 256 vertical extent of quaternary sand deposits at different locations nearby the main central river 257 (Fig. 2). Both horizontal and vertical dipole configurations were implemented at distances of 10, 258 20, and 40 m between the transmitter and receiver coils, to increase the investigated depth (see 259 Supplementary Material; McNeill 1980). The mean distance between each measurement point 260 was 250 m. 261

262 In addition to these geophysical data, we also sampled all accessible lithologies to measure some petrophysical properties in laboratory in order to constrain numerical models of the geological 263 264 sources of the observed geophysical anomalies. The bulk density of these samples was measured using a Quantachrome Helium stereopycnometer. Magnetic parameters used to constrain the 265 model are described in Lepaulard et al. (2019): the remanent magnetization intensity was 266 obtained from measurements with a Superconducting Quantum Interference Devices (SQUID) 267 760R (2G Entreprises) magnetometer while the magnetic susceptibility was measured with an 268 AGICO Kappabridge MFK1. 269

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RESULTS AND INTERPRETATION

The complete Bouguer gravity anomaly map is shown in Fig. 3A. It was generated using a 272 minimum curvature interpolation and masking areas with no data. Despite the influence of radial 273 line surveys, the map reveals a general negative gravity anomaly of ~3 mGal amplitude and ~10 274 km of wavelength over the center of the Tunnunik impact structure. In the absence of data 275 outside of the surveyed area, we cannot infer that this is the maximum amplitude of the anomaly 276 observed over this impact structure (i.e. 3 mGal is a lower limit for the amplitude of the 277 anomaly). In detail, the shape of this central anomaly does not appear to be circular but seems 278 slightly more extended in the N-S direction. The minimum value of -4.6 mGal is also not at the 279 center but is located in a ring of about -4 to -4.5 mGal anomalies surrounding a less negative (-280 2.5 mGal) isolated central anomaly; this is considered to be located near the center of the 281 structure (402000, 8044000). Fig. 3A shows that most of the concentric negative extrema 282 correspond to the Quaternary sand deposits, suggesting that this porous formation may locally 283 amplify the general low gravity signal at the center. Fig. 3B shows that 5 km away from the 284 center, the anomaly gradients are less important, but still a small increase is observed at the 285 borders, arguing in favor of a more extended anomaly. Therefore, we conclude that a central 286 negative anomaly is visible, with a possible asymmetric shape and with a possible extension 287 beyond our survey area. This gravity anomaly is not observed in the regional gravity data 288 acquired by the National Research Council of Canada (see Supp. Mat. for details of calculations 289 and regional trends in Fig. S1). 290



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Fig. 3. (A) Complete Bouguer gravity anomaly map over the center of the Tunnunik impact structure. The thin pink lines correspond to the gravity anomaly profiles selected for modeling and shown in (B). Faults are indicated by full black lines. The area where shatter cones were observed and collected is delimited by the dashed red line (Osinski and Ferrière 2016), while the limits of the Quaternary Sand Formation correspond to the thin green lines. Grayscaled background corresponds to the digital elevation model shown in Fig. 2. Same coordinate system as for Fig. 1. (B) Complete Bouguer gravity anomaly data along the selected profiles in (A).

The interpolated map of the total-field magnetic anomaly over the center of the Tunnunik impact 302 structure is shown in Fig. 4A. The data distribution limits the analysis to the central part (i.e. 303 within the shatter cone area) only. The total amplitude of the interpolated and filtered signal 304 reaches 120 nT, with minimum and maximum values of -20 and 96 nT, respectively. The 305 positive part of the anomaly is restricted to the central area of the structure and centered at 306 401000, 8045000. It correlates well with the Shaler Supergroup (with intruding diabase dykes) 307 outcrops (Fig. 4A). Compared to the gravity anomaly, the center of the positiveanomaly is 308 located ~500 m away in the NW direction, but on average, the observed central negative 309 gravimetric and positive magnetic anomalies overlap. Although less extended than the Bouguer 310 gravity anomaly, the central positive magnetic field anomaly also seems to be slightly 311 312 asymmetrical with a major axis in the NNW–SSE direction and a minor axis in the WSW–ENE direction, with wavelengths of ~6 and ~4 km respectively. Fig. 4B shows the shape of the 313 anomaly in the N-S and W-E directions, unveiling this weak asymetry. 314

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317 Fig. 4. (A) Magnetic field anomaly map of the center of the Tunnunik impact structure. The green solid

318 line corresponds to the locations of the data selected for modeling. Faults are indicated by full black lines.

319 The area where shatter cones were observed and collected is delimited by the dashed red line (Osinski and

Ferrière 2016). The area where the Shaler Supergroup Formation is exposed is delimited by the dotted purple line. Grayscaled background corresponds to the digital elevation model shown in Fig. 2. Same

coordinate system as for Fig. 1. (B) Magnetic field anomaly along the selected profiles in (A).

Thus, both gravity and magnetic field signals show a significant anomaly of several km wavelength in the surveyed area. Their correlation argues in favor of a possible single geological source at or nearby the center of the impact structure.

Fig. 5A shows the spatial variations of the subsurface electrical conductivity derived from the 326 electromagnetic measurements performed in the central area of the impact structure (see Supp. 327 Mat. for results of EM surveys performed outside this central part). It clearly reveals that the hills 328 around the center and nearby the river contain porous liquid water loaded material (> 20 mS/m) 329 that contrasts with the resistive frozen soil outside these areas (< 5 mS/m; see Todd et al. 1991 330 for a comparison in the same environment). The conductive material corresponds to the sand 331 formations either originated from fluvioglacial processes (Choe et al. 2019) or deposited by sea 332 level variations during Quaternary era (Fig. 5B). Within these deposits, variations of the 333 conductivity between 10 and 30 mS/m are observed, reflecting possible variations in the amount 334 335 of pore water and its salinity eventually due to presence of fossil sea water. Presence of permafrost below the liquid water layer is also a possibility, while permafrost formation above it 336 may have been prevented by a thick dry sand layer (Fig. 5C). The conductivity remains high at 337 338 the maximum possible depths reached by our measurements, indicating that this sand formation is thicker than 60-80 m. This observation is important because such a thick, weakly dense (2,000 339 kg m⁻³), material will decrease the Bouguer gravity value for data acquired on top of (or nearby) 340 341 this formation. In fact, most of the lowest gravity anomalies correspond to the location of these deposits (see Fig. 5A for an example of the correlation with gravity low values). Still, on 342 kilometric scale, there is a general, centered and negative ~3 mGal Bouguer gravity anomaly 343 344 over the center of the impact structure.



Fig. 5. (A) Map of the electrical conductivity variations at about 20 m in depth, in the central area of the impact structure. Bouguer gravity data are also shown to illustrate the correlation between some isolated low gravity values and the Quaternary Sand deposits. Dashed black line corresponds to the cross-section shown in (C). (B) Image of the conductive formation composed of Quaternary sediments nearby the central river. (C) W-E cross-section resulting from interpolation of conductivity data (vertical exaggeration ~ 2; points: data from different coil orientations at each location) acquired on the sand formation in the same area, along the dashed line in (A).

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DISCUSSION

358 Diameter and erosion

Tunnunik is a mid-size impact structure with a ~10 km wavelength negative Bouguer gravity 359 anomaly of ~3 mGal amplitude over its center. Mid-size complex impact craters (see Osinski and 360 Grieve 2013 for a review) are typically characterized by circular negative gravity anomalies that 361 extend to or beyond the crater diameter. The size and amplitude of the anomaly are expected to 362 increase with an increasing apparent crater diameter D_a (Pilkington and Grieve 1992; Grieve and 363 Pilkington 1996). Since our measurements are limited to the center of the impact structure, they 364 do not reach the outermost faults (outside the map in Fig. 2), which represent the outer limit of 365 D_a, as defined in Osinski et al. (2013). Thus, we can expect a larger amplitude and wavelength of 366 the low gravity signal observed at Tunnunik: the Bouguer gravity data towards the rims still 367 show – in all directions – a small increase before reaching the regional signal (here leveled to 0 368 mGal). The minimum amplitude of the anomaly (~3 mGal) could reflect a deeply-eroded 369 370 structure, because impact structures of comparable size generally have a higher amplitude (e.g., 10 mGal for Keurusselkä; Raiskila et al. 2013). Alternatively, this amplitude could be due to the 371 type of target rocks: impact structures in sedimentary targets generally have lower maximum 372 negative Bouguer gravity anomalies (Δg) than those in crystalline targets (Pilkington and Grieve 373 1992). 374

Impact structures can be classified using the erosional level index (E) established by Dence 375 (1972; see also Grieve and Robertson 1979) and based on the preservation of their ejecta, crater-376 fill products and exposure of crater floor (Pilkington and Grieve 1992). Most of the indicated 377 diameters in impact studies are apparent crater diameters D_a (Osinski and Pierazzo 2013). 378 Indeed, the majority of complex impact craters on Earth are eroded (i.e., they are impact 379 380 structures) and therefore only D_a can be determined in most cases. Tunnunik seems to correspond to the worst state of preservation: indeed the crater floor has been removed and the substructure 381 exposed, with no remnant of ejecta or crater-fill deposits preserved (Dewing et al. 2013; Osinski 382 et al. 2013; Newman and Osinski 2016). The only preserved impact breccia found in the field are 383 in the form of very local and thin (< 1 m of width) dykes injected into the sedimentary target 384 rocks (Osinski et al. 2013; Newman and Osinski 2016). Slightly modifiying the index definition, 385 386 Osinski and Ferrière (2016) classified the Tunnunik structure as erosion state 6, because of these still visible isolated breccia dikes. Such a deep level of erosion can be either due to a particularly 387 old impact structure or to an especially high rate of erosion, or both. In the present-day 388 periglacial environment of the Canadian High Arctic archipelago (e.g., Lee and Osinski 2005), 389 the 23 Ma Haughton impact structure, located about 900 km away, has been relatively well 390 preserved. Haughton is characterized by a $D_a \approx 23$ km and a smaller estimated rim-to-rim 391 diameter (D) of 16 km (Osinski et al. 2005). Its negative Bouguer gravity anomaly has a 12 392 mGal amplitude and a 24 km wavelength, the latter being very similar to its D_a (Pohl et al., 393 1988). It has an erosional level of 2, which means that it is relatively fresh with ejecta partly 394 preserved (Osinski and Ferrière 2016). As mentioned earlier, the diameter of the gravity anomaly 395 should be roughly equal to D_a in relatively fresh craters such as Haughton, while the amplitude -396 more affected by erosion - is not a relevant criterion (Pilkington and Grieve 1992). Therefore, 397

our gravimetric data suggest that the Tunnunik impact structure is severely eroded, in agreement

with its old age. Indeed, using paleomagnetism, Lepaulard et al. (2019) estimated that this impactevent occurred 430-450 Ma ago.

A similarly eroded impact structure is Gosses Bluff in Australia (Barlow 1979), which also formed in sedimentary target rocks. Its apparent crater diameter D_a has been recently revised to 32 km (Osinski and Ferrière 2016). Its total gravimetric anomaly is $\Delta g = -5.5$ mGal with a central peak (CP) of -3.5 mGal ($\Delta g = -3$ to -4 mGal and CP = -2.5 mGal at Tunnunik). This 142.5 Ma impact structure has been attributed an erosion level similar to Tunnunik (E = 6–7; Pilkington and Grieve 1992; Osinski and Ferrière 2016), while the presence of crater-fill breccias should revise this value to E = 5 (Milton and Sutter 1987). Nevertheless, it shows that the relatively

408 weak amplitude of the remaining gravity anomaly at Tunnunik is not a unique case.

Besides, another phenomenon could explain this feature, without requiring a high amount of 409 erosion. Indeed, diagenetic pressure-solution (Sprunt 1977) could cement the impact-induced 410 fracturation in target rocks buried below a few kilometers of post-impact sediments. To our 411 412 knowledge, this post-impact phenomenon was never invoked before as a possible way to reduce the amplitude of gravity anomalies in impact structures in sedimentary carbonated rocks. Long 413 term healing of impact induced carbonate porosity by pressure-solution is a well-known 414 phenomena (Croizet et al. 2013) that may be particularly effective in the Tunnunik limestones 415 due to the several hundreds Myr of burial endured by the fractured rock. Moreover, as the impact 416 occurred around 430-450 Ma ago, soon after deposition of the Ordovician sediments, this target 417 may have been not fully lithified, resulting in a lesser amount of fracturing and/or easier healing 418 of fractures. Possibly, it could be a new factor contributing to the generally weaker maximum 419 negative Bouguer gravity values for impact structures associated with sedimentary targets 420 (Pilkington and Grieve 1992). Lastly, the effect of central uplift of underlying basement has to be 421 taken into account, as it can partly cancel the broad negative anomaly in the structure center, as 422 described for example in the 6 km diameter Jebel Waqf as Suwwan structure (Heinrichs et al. 423 2014). 424

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Modeling the Tunnunik impact structure and post-impact erosion

A possible geological solution explaining the observed gravity and magnetic field anomalies 427 over the center of the Tunnunik impact structure is shown in Fig. 6. This numerical model is 428 429 constrained by (1) gravity and magnetic field data, (2) average values of the bulk density for the Earth's layers such as the crust and the mantle (not shown in Fig. 6), (3) measured physical 430 parameters for the different geological formations identified and sampled in the field, *i.e.*, bulk 431 densities, magnetic susceptibility, natural remanent magnetization (NRM) intensity, inclination 432 (I) and declination (D), and (4) field geological data and mapping, *i.e.*, thickness of geological 433 formations (Dewing et al. 2013) and location of structures such as contacts and faults (Newman 434 435 and Osinski 2016). Table 1 shows the measured physical parameters used in the model. Consistent with geological mapping, the model shows the remains of a central uplift with the 436 Shaler Supergroup in the very centre. Clear lateral physical contrasts may reflect faults, 437 particularly if they join some mapped faults on surface. Near the center (i.e., ± 2 km from the 438 profile center), the uplift of sedimentary units likely occurred along thrust faults (some are 439 mapped and shown on Fig. 1). The crystalline basement core of this central uplift underlying the 440

Shaler formation seems affected by brecciation and/or fracturing, as well as by an enhancement 441 of its magnetization intensity, required to explain the central magnetic anomaly detected on Fig. 442 4. The shape of the strongly-magnetized crystalline basement body contributes to the skewness 443 of the anomaly (see profiles on Fig. 4B). Nearby this basement core, the deep part of the Shaler 444 Supergroup unit also possesses a magnetization intensity larger than the overlying and 445 surrounding parts, but about 10 times weaker than the one of the crystalline basement. More 446 dolerite dykes may be present at depth in this unit. Overall, ~5 km from the center, coinciding 447 with the shatter cone area, and from 0.7 to 1 km in depth at the center, the rocks seem more 448 brecciated and fractured than outside, with 20 to 100 kg/m³ density reduction (*i.e.* a maximum of 449 4% of porosity increase). This is probably the most striking result of this study: even if the 450 Tunnunik impact structure is highly eroded and even if the Quaternary Sand Formation biases 451 the general low gravity signal by producing isolated very low (down to -4.6 mGal) gravity 452 values, a ~1 km-thick zone of brecciated and fractured uplifted rocks (including basement) seems 453 to be present. The presence of crystalline basement rocks (denser than the overlying units despite 454 brecciation) in the central uplift also accounts for the local gravity 'high' of -2.6 mGal at the 455 center of the structure, within the general gravity low. There is also a transition zone of density 456 reduction (~10 kg/m³) between 5 and 10 km of lateral distance from the center, inside the rim of 457 local faults that surrounds our studied area (see Figs. 1 and 2). 458

A second model was built along the W-E direction profiles and is shown in the Supplementary 459 Material (Fig. S3). It reveals a structure similar to the one shown in the N-S model of Fig. 6, 460 461 again with ~1 km-thick brecciated/fractured rocks at the center, and with an associated enhancement of magnetization. The combination of these 2 perpendicular profiles does support a 462 small assymetry of the brecciated rocks towards N (Fig. 6), but not really towards W (Fig. S3). 463 Of course, in the absence of data with higher spatial resolution (especially gravity data) and of 464 borehole data, more detailed geological models (including 3D representation) are not possible. 465 The pre-impact distribution of densities within the target rocks is also a key parameter in those 466 467 models (Pilkington and Grieve 1992). Despite stratigraphic information from Dewing et al. (2013), our models show that sharp and significant (e.g. ~300 m for some layers in the central 468 area) thickness variations are necessary to explain the observed anomalies. This variability is due 469 to faulting which offsets formations, including the Mount Phayre Formation which has a weak 470 471 density (~2,500 kg/m³). The porous Quaternary sand units, possibly thicker than 100 m (as shown by EM34 measurement interpretation) and correlated with the very low gravity values, 472 473 also play a role in decreasing the RMS residuals (of about 0.2 mGal) between observations and predictions of the Bouguer gravity signal. 474



477 Fig. 6. N-S profile modeling of the Tunnunik impact structure constrained by magnetic and gravity data, 478 as well as by laboratory measurements on samples. Top: comparison of observed (black) and predicted 479 (gray) magnetic and gravity anomaly data. Location of the gravity measurements are shown by the solid points. Bottom: corresponding forward numerical model of the crust over the center of the Tunnunik 480 impact structure. The physical parameters of each layer are shown in Table 1. Fracturing and brecciation 481 (symbolized by gray hatching) decrease the bulk density values of all geological layers from the outer 482 483 parts of the crater to the center. Topographic variations are from the global digital elevation model 484 version 2 (GDEM2) of ASTER (NASA-METI). No vertical exaggeration.

485 Table 1. Physical parameters of the Tunnunik geological layers

Geological unit	D ^{1,2} (kg m ⁻³)	K ¹ (10 ⁻³ SI)	NRM ¹ (A.m ⁻¹)	$\mathrm{I}^{1}\left(^{\circ} ight)$	D^1 (°)
Quaternary sand formation	2000	-	-	-	-
Thumb Mountain / Allen Bay formation	2740-2760	0.02	< 10 ⁻³	-	-
Victoria Island formation	2720-2750	0.01	< 10 ⁻³	-	-
Mount Phayre formation	2470-2580	0.07	< 10 ⁻³	-	-
Shaler Supergroup (with dolerite dykes) ³	2630-2650	0.04 - 0.2	0.01 - 0.03	80-90	0
Crystalline basement ⁴	2650-2700	50	0.1 – 0.5	90	0

486 ¹d, bulk density; K, volumic magnetic susceptibility; NRM, natural remanent magnetization; I and D, inclination and declination

487 of NRM, respectively

488 ²Minimum bulk density values are associated with rock samples collected nearby the center (i.e. distance < 5 km from modeling

489 profile center), while maximum values correspond either to samples collected far from the center (i.e. d ~ 8-10 km) or to values

490 expected from the forward modeling (e.g., for the upper crust basement).

491 ³The Shaler Supergroup rocks are intruded by dolerite dykes of $K = 4 \ 10^{-2}$ SI, NRM = 0.7 A.m⁻¹, I~30°, D~110° (Lepaulard et al.,

492 2019). Therefore we adjusted the parameters of the corresponding modeled layer by considering about 5% of dolerite dykes.

493 ⁴The basement with an enhanced magnetization is correlated to the basement affected by brecciation (see Fig. 6).

494

495 Despite the simplicity of the models, there is a good agreement with computed estimations of brecciation/fracture depth extent in the presence of a low gravity anomaly. Indeed, according to 496 the infinite slab model formula (see Supp. Mat. for calculation) for structures of $D_a \ge 20-30$ km 497 (Pilkington and Grieve 1992), a maximum density contrast of 100 kg m⁻³ (Table 1), associated 498 with a maximum negative gravity value $\Delta g = -3$ to -4 mGal, gives a maximum depth extent Z = 499 715 to 954 m, a value in the range of the depth extent (0.7-1 km) of the brecciated/fractured zone 500 of our models (Figs. 6 and S3). A density contrast of about 150 kg m⁻³ has been determined at 501 Gosses Bluff (Barlow 1979), a structure similar to Tunnunik (see discussion above). However, 502 the determination of density contrasts between unfractured and fractured target rocks remains 503 sparse (see Pilkington and Grieve 1992 for a list). Similarly, for deeply eroded structures (E = 6-504 7; Grieve and Robertson 1979), the amount of removal of the disturbed zone beneath the crater 505 floor is poorly constrained, allowing for significant variations in the corresponding gravity effect 506 (Pilkington and Grieve 1992). 507

At Haughton, the expected post-impact erosion (~150 m; Osinski et al. 2005) together with the 508 age of the structure (~ 23.5 Ma; Young et al. 2013) lead to an average erosion rate of 6.4 m Ma⁻¹ 509 since the time of the impact. Because Tunnunik is at high latitudes in an expected relatively 510 stable tectonic environment of the Canadian Arctic since about 150-200 Ma, we can suppose as 511 first approximation the same erosion rate as Haughton to obtain a minimum estimate of the 512 erosion at Tunnunik. Using the paleomagnetic age of ~430 Ma for the Tunnunik impact structure 513 (Lepaulard et al. 2019) with Haughton's erosion rate of 6.4 m Ma⁻¹, it gives ~2.7 km of post-514 impact erosion. Using only 200 Ma with the same rate, it leads to \sim 1.3 km. One could argue that 515 this post-impact erosion is responsible for the weakness of the observed negative Bouguer 516 gravity anomaly. However, because the impact indeed happened ~430 Ma in the Silurian or in 517 the late Ordovician, the target was "only" composed of the Allen Bay (Ordovician-Silurian), 518 Victoria Island (Ordovician), Stripy Unit (Cambrian) and Wynniat (Proterozoic) formations 519 520 which are observed in the structure (Dewing et al. 2013). Therefore, only the younger impacted unit (Allen Bay) could have been partially eroded away. Its thickness is supposed to be of 521 maximum ~1 km (Dewing et al. 2013). Combined with the possible uncertainty in paleomagnetic 522 dating, this could give a maximum thickness of about 1 km for the target rocks possibly eroded 523 away. One cannot assess the thickness of post-impact deposits, but between 430 and ~200 Ma, 524 the Tunnunik impact structure moved from equatorial to high latitudes. It implies possible higher 525 erosion rates during this period, which may compensate the deposition of sediments. The 526 apparent crater diameter D_a shrinks with increasing depth of erosion Z_E by $D_a = D - 1.15 Z_E$, 527 because of the inclination of crater rim normal faults which dip toward the center of the crater by 528 generally 60° (Kenkmann et al. 2013). According to this relation, and supposing $Z_{Emin} = 1$ km 529 (erosion of target rocks), $Z_{Emax} = 2.7$ km (total erosion) and a $D_a = 28$ km, the final (rim-to-rim) 530 diameter D of the fresh Tunnunik impact crater could have been between ~29 km and ~31 km. 531

532

533 Magnetization contrasts

The positive magnetic field anomaly detected over the center of the Tunnunik impact structure 534 appears to almost exactly correlate to the outcrop area of the Wynniatt (part of the Shaler 535 Supergroup) Formation (Figs. 1 and 4). Similar to Haughton, the most magnetic rock mapped 536 and collected in the area is dolerite (K = $4 \ 10^{-2}$ SI and NRM = $0.7 \ A.m^{-1}$ in average). This mafic 537 rock is present as localized dykes intrusive into the crystalline basement and into the Shaler 538 Supergroup units (Dewing et al. 2013). To take into account these mafic dykes in our forward 539 model, we considered that 5% of the volume of this formation was composed of dolerite, 540 resulting in a maximum magnetization intensity of 0.03 A.m⁻¹ with a magnetic susceptibility of 2 541 10⁻¹ SI (Table 1). These values are several orders of magnitude higher than the NRM and K of all 542 other sedimentary rocks present in the area (therefore considered with M = 0 A.m⁻¹ in the 543 modeling). However, our models show that an additional deeper strongly-magnetized (0.5 A.m⁻¹) 544 source is needed to explain the shape and amplitude of the central magnetic anomaly (Figs. 6 and 545 S3). As mentioned in the previous sections, we suggest that this could be uplifted crystalline 546 basement rocks. Such magnetized crystalline blocks within the central uplift could be explained 547 either by the intrusion of dykes of impact melt rocks and breccias with a significant thermal 548 remanent magnetization (e.g., Shah et al. 2005), or by shock-induced magnetization of pre-549 impact rocks (Cisowski and Fuller 1978), or by an enhanced magnetization of the basement due 550 to hydrothermal alteration (Quesnel et al. 2013). The latter has been shown to increase the NRM 551 within impact melt rocks of the central uplift at the nearby Haughton impact structure 552 (Zylberman et al. 2017). Another possibility would be a concentration of mafic, pre-impact 553 magmatic rocks at local scale, possibly remagnetized by the impact. Of course, a combination of 554 these different processes is still possible. 555

556

557 Conclusion

The first geophysical measurements within the Tunnunik impact structure indicate the existence 558 of a central ~10 km wavelength negative Bouguer gravity anomaly of ~3 mGal and a positive 559 magnetic anomaly of ~120 nT amplitude. It shows that, despite the highly-eroded state of the 560 structure, the geophysical signature of the impact structure is still preserved. Using forward 561 modeling constrained by petrophysics measurements, the geometry of the sources accounting for 562 these anomalies is estimated. It reveals a fractured zone down to 0.7-1 km depth in its present-563 day state, and suggests that at least 1 km of erosion has occurred since 430-450 Ma, the time of 564 impact. The positive magnetic field anomaly is suggested to be mainly due to uplifted crystalline 565 basement, but also by the Shaler Supergroup sedimentary formation which is intruded by dolerite 566 dykes. The origin of the magnetization's enhancement for the crystalline basement is still 567 unknown, even if impact-generated hydrothermal activity is a good candidate. 568

To decipher the issues unveiled by these first measurements and by modeling, additional 569 gravimetric data would be helpful and other geophysical methods could be used across the whole 570 structure, like seismics and/or magnetotellurics which could reveal the deep 571 velocity/conductivity contrasts between fractured/brecciated and pristine crystalline basement at 572 depth. 573

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841 **Figure captions**

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Fig. 1. Geological map of the Tunnunik impact structure, modified from Dewing et al. (2013) and 843 Newman and Osinski (2016). Background corresponds to a Map Data ©2015 Google satellite image. 844 845 Stratigraphy: Neoproterozoic → Shaler Supergroup (Neoproterozoic), Cambrian → Mount Phayre formation (Cambrian), Cambrian/Ordovician \rightarrow Victoria Island formation (Cambrian/Ordovician), and 846 Thumb Mountain/Allen Bay formation (Upper Ordovician/Silurian). The coordinate system for the 847 848 geological map is UTM Zone 12 North projection with WGS84 datum, in meters. Upper left: Location of the structure on Victoria Island in the Canadian Arctic, with a geographic coordinate system on a WGS84 849 850 datum. Background corresponds to the ArcGIS online ESRI Ocean laver.

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Fig. 2. Geophysical measurements performed at Tunnunik. Background corresponds to a digital elevation model version 2 (GDEM2) from ASTER (NASA-METI). Same coordinate system as for Fig. 1.

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Fig. 3. (A) Complete Bouguer gravity anomaly map over the center of the Tunnunik impact structure. The thin pinkwhite lines corresponds to the gravity anomaly profiles selected for modeling and shown in (B). Faults are indicated by full black lines. The area where shatter cones were observed and collected is delimited by the dashed red line (Osinski and Ferrière 2016), while the limits of the Quaternary Sand Formation correspond to the thin green lines. Grayscaled background corresponds to the digital elevation model shown in Fig. 2. Same coordinate system as for Fig. 1. (B) Complete Bouguer gravity anomaly data along the selected profiles in (A).

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Fig. 4. (A) Magnetic field anomaly map of the center of the Tunnunik impact structure. The green solid
line corresponds to the locations of the data selected for modeling. Faults are indicated by full black lines.
The area where of shatter cones were observed and collected is delimited by the dashed red line (Osinski
and Ferrière 2016). The area where outcrops the Shaler Supergroup Formation is exposed is delimited by
the dotted purple line. Grayscaled background corresponds to the digital elevation model shown in Fig. 2.
Same coordinate system as for Fig. 1. (B) Magnetic field anomaly along the selected profiles in (A).

Fig. 5. (A) Map of the electrical conductivity variations at about 20 m in depth, in the central area of the impact structure. Bouguer gravity data are also shown to illustrate the correlation between some isolated low gravity values and the Quaternary Sand deposits. Dashed black line corresponds to the cross-section shown in (C). (B) Image of the conductive formation composed of Quaternary sediments nearby the central river. (C) W-E cross-section resulting from interpolation of conductivity data (vertical exaggeration ~ 2; points: data from different coil orientations at each location) acquired on the sand formation in the same area, along the dashed line in (A).

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878 Fig. 6. N-S profile modeling of the Tunnunik impact structure constrained by magnetic and gravity data, as well as by laboratory measurements on samples. Top: comparison of observed (black) and predicted 879 (gray) magnetic and gravity anomaly data. Location of the gravity measurements are shown by the solid 880 points. Bottom: corresponding forward numerical model of the crust over the center of the Tunnunik 881 impact structure. The physical parameters of each layer are shown in Table 1. Fracturing and brecciation 882 (symbolized by gray hatching) decrease the bulk density values of all geological layers from the outer 883 884 parts of the crater to the center. Topographic variations are from the global digital elevation model 885 version 2 (GDEM2) of ASTER (NASA-METI). No vertical exaggeration.