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*Total number of authors:*  
32

*Published in:*  
Earth System Science Data

*Link to article, DOI:*  
[10.5194/essd-14-2209-2022](https://doi.org/10.5194/essd-14-2209-2022)

*Publication date:*  
2022

*Document Version*  
Publisher's PDF, also known as Version of record

[Link back to DTU Orbit](#)

*Citation (APA):*  
Colgan, W., Wansing, A., Mankoff, K., Lösing, M., Hopper, J., Loudén, K., Ebbing, J., Christiansen, F. G., Ingeman-Nielsen, T., Liljedahl, L. C., MacGregor, J. A., Hjartarson, Á., Bernstein, S., Karlsson, N. B., Fuchs, S., Hartikainen, J., Liakka, J., Fausto, R. S., Dahl-Jensen, D., ... Løkkegaard, A. (2022). Greenland Geothermal Heat Flow Database and Map (Version 1). *Earth System Science Data*, 14(5), 2209-2238.  
<https://doi.org/10.5194/essd-14-2209-2022>

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## Greenland Geothermal Heat Flow Database and Map (Version 1)

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Received: 26 August 2021 – Discussion started: 13 September 2021

Revised: 10 February 2022 – Accepted: 10 March 2022 – Published: 12 May 2022

**Abstract.** We compile and analyze all available geothermal heat flow measurements collected in and around Greenland into a new database of 419 sites and generate an accompanying spatial map. This database includes 290 sites previously reported by the International Heat Flow Commission (IHFC), for which we now standardize measurement and metadata quality. This database also includes 129 new sites, which have not been previously reported by the IHFC. These new sites consist of 88 offshore measurements and 41 onshore measurements, of which 24 are subglacial. We employ machine learning to synthesize these in situ measurements into a gridded geothermal heat flow model that is consistent across both continental and marine areas in and around Greenland. This model has a native horizontal resolution of 55 km. In comparison to five existing Greenland geothermal heat flow models, our model has the lowest mean geothermal heat flow for Greenland onshore areas. Our modeled heat flow in central North Greenland is highly sensitive to whether the NGRIP (North Greenland Ice core Project) elevated heat flow anomaly is included in the training dataset. Our model's most distinctive spatial feature is pronounced low geothermal heat flow ( $< 40 \text{ mW m}^{-2}$ ) across the North Atlantic Craton of southern Greenland. Crucially, our model does not show an area of elevated heat flow that might be interpreted as remnant from the Icelandic plume track. Finally, we discuss the substantial influence of paleoclimatic and other corrections on geothermal heat flow measurements in Greenland. The in situ measurement database and

gridded heat flow model, as well as other supporting materials, are freely available from the GEUS Dataverse (<https://doi.org/10.22008/FK2/F9P03L>; Colgan and Wansing, 2021).

## 1 Introduction

Assessing the magnitude and spatial distribution of geothermal heat flow across Greenland is important for many reasons, such as mapping geothermal springs and energy resources; constraining a key basal boundary condition for the permafrost, glaciers, and the ice sheet; constraining the thermal structure and material properties of the lithosphere; and understanding the generation and preservation of hydrocarbon accumulations. However, the current generation of Greenland regional heat flow models still shows substantial disagreement (Rezvanbehbahani et al., 2017; Martos et al., 2018; Greve, 2019). A fundamental challenge in reliably interpolating the magnitude and spatial distribution of geothermal heat flow across Greenland is the paucity of local heat flow measurements with which to constrain regional heat flow models.

Of the 40 870 onshore heat flow measurements presently catalogued in the International Heat Flow Commission (IHFC) database, just 10 ( $\sim 0.02\%$ ) are onshore in Greenland (Fuchs et al., 2021a). As Greenland represents  $\sim 1.5\%$  of global land area, this makes the country disproportionately underrepresented in the IHFC database. While several studies have assembled additional non-IHFC heat flow measurements from published sources (Martos et al., 2018; Rysgaard et al., 2018), it is highly desirable to have a comprehensive and continuously updated open-access repository of all Greenland heat flow measurements. It is similarly desirable to have an open-access Greenland heat flow map that is self-consistent with that updating repository.

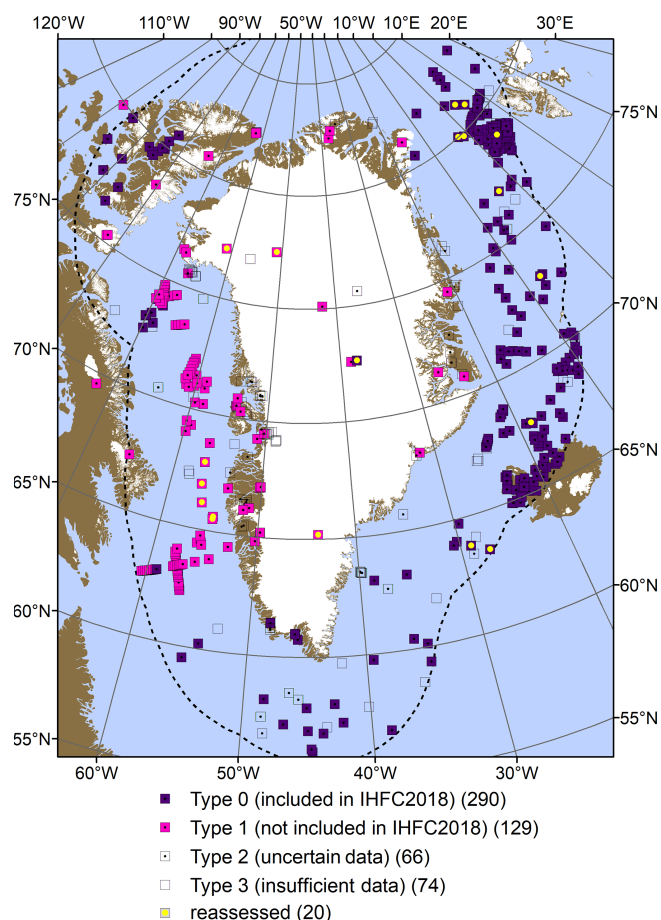
The goal of this study is to develop and describe this first version of the Greenland Geothermal Heat Flow Database and Map. First, we collect existing and new heat flow measurements in and around Greenland into a database with uniform metadata. Second, we apply machine learning to this database, along with other geophysical datasets, to produce a regional model of the magnitude and spatial distribution of Greenland heat flow that is consistent with these measurements. These data products and their supporting information are publicly available. We anticipate updating the heat flow database and map as new measurements become available. Here, we describe the development of these data products and discuss their implications for advancing our understanding of Greenland's geothermal heat flow.

## 2 Methods

### 2.1 Heat flow measurement database

We use the 2018 IHFC database as the foundation upon which to build a region-specific update of Greenland heat flow measurements. This 2018 version of the IHFC database, which contains 58 536 measurements (both onshore and offshore) with no quality documentation, is an updated version of the 2013 IHFC database (Global Heat Flow Compilation Group, 2013; <https://doi.org/10.1594/PANGAEA.810104>). Within our Greenland domain, the IHFC 2018 database contains 290 heat flow measurements, the majority of which are offshore (Fig. 1). In 2021, the fundamental structure of the IHFC database was revised for the first time since 1976 (Jespersen et al., 1976; Fuchs et al., 2021b). While the previous 1976 IHFC database structure contained 17 data fields for each heat flow measurement, the new 2021 IHFC database structure now contains 59 data fields for each heat flow measurement, 18 of which are “mandatory”, 32 of which are “recommended”, and 9 of which are “optional”. The novel heat flow database we present here has 26 data fields for each heat flow measurement. Eighteen of these fields have 100 % data coverage, but these do not align with all 18 “mandatory” IHFC 2018 fields (Table A1). Four fields, associated with the temperature gradient, depth range, and thermal conductivity, have better than 81 % data coverage. The last four fields, associated with topographic correction of heat flow, have 34 % data coverage.

We briefly introduce and describe the fields of our database in Table 1. While our database generally exceeds the information required by the IHFC 1976 database structure, it does not contain 8 of the 18 mandatory fields of information required by the IHFC 2021 database structure (Table A1). Conforming and assimilating our database into the IHFC 2021 database and structure remains a near-term but non-trivial goal. While translating the “measurement type” of this study into “geographical environment” of IHFC 2021 is relatively straightforward, assessing IHFC 2021 fields like “TC saturation” or “TC pT conditions” (where TC refers to thermal conductivity, and pT refers to pressure and temperature) requires reviewing historical measurements on a site-by-site basis. Similar to the IHFC databases, however, our present database has a “one borehole, one estimate” philosophy. This merges multiple estimates for a single borehole into a single best or consensus estimate for that borehole. Generally, for the predominantly subglacial sites where multiple estimates have been published, the consensus estimate is reached by expert elicitation within our author team. Further, our database only



**Figure 1.** Overview of site locations and types in the heat flow measurement database. Reassessed heat flow values described in Tables 3 and 4. Dashed line denotes our study boundary, 500 km from Greenland’s coasts. The Meighen and Barnes ice caps lie outside this boundary, but we still report these subglacial measurements here. Projection is EPSG:3413. See Fig. 6 for the sites overlaid on a bathymetric map.

includes direct heat flow estimates: those derived from temperature gradient measurements. Following the IHFC, we exclude heat flow estimates derived from remotely sensed approaches (i.e., Cox et al., 2021).

Generally, our present database has a greater focus on three-dimensional positional accuracy than the IHFC databases. We report positions, elevations, and uncertainties in both latitude–longitude and the popular EPSG:3413 coordinate system adopted by many Greenland-focused and Arctic-wide data products. The EPSG:3413 system, also known as the NSIDC Sea Ice Polar Stereographic North projection, has its latitude of origin at 70° N and its central meridian at 45° W and provides a polar projection for data products (see: <http://epsg.io/3413>; last access: 8 February 2022). This focus on three-dimensional positional uncertainty is intended to facilitate future investigations of local heat flow corrections, such as those associated with paleocli-

mate or topography. As Greenland is a high-relief land mass with a complex climate history, these corrections are likely more important in modifying contemporary heat flow measurements in Greenland than lower-relief and more temperate regions of Earth.

## 2.2 Heat flow measurements

Heat flow measurements typically rely on knowledge of local temperature gradient and local thermal conductivity, with the product of these two terms yielding heat flow. There can be substantial diversity in the depth interval over which temperature gradient is measured, from kilometer-scale deep boreholes to meter-scale heat probes. Similarly, there can be substantial diversity in how thermal conductivity is estimated, ranging from relatively precise continuous-core laboratory analysis to less precise estimations based on tabulated data from analogous rock types. For this reason, it is desirable to present a heat flow measurement with sufficient metadata for users to assess the relative differences in data quality between sites (Table 1). In this database, we assess uncertainty in heat flow based on the quality of the local temperature gradient and thermal conductivity data comprising the heat flow. In the following, all mentions of “conductivity” relate to “thermal conductivity”. We divide the heat flow measurements into four types: Type 0 contains measurements already included in IHFC 2018, Type 1 contains “new” measurements that are not included in IHFC 2018, Type 2 contains sites for which we are uncertain if sufficient data exist to estimate heat flow, and finally Type 3 contains sites where we feel there are insufficient data to estimate heat flow with available methods.

### 2.2.1 Type 0 – included in IHFC 2018

The IHFC 2018 database contains 290 heat flow measurements within our Greenland domain (Fig. 1). Here, we provide the first systematic quality control of these existing IHFC data, as part of a broader community effort to bring all historical IHFC into compliance with the most recent IHFC data standards (Fuchs et al., 2021a). During this quality control, we reassess 10 of these historical IHFC heat flow values in our present database (Table 2). The remaining 280 historical IHFC heat flow measurements are unchanged.

The largest reassessment is site V28-11. The heat flow at this site –  $871 \text{ mW m}^{-2}$  – is a clear outlier within the IHFC 2018 database, exceeding the next highest value within the Greenland region by a factor of 2. It is also more than an order of magnitude greater than the contemporaneous measurement at V30-96, located 72 km away. While all temperature gradients measured during the V28 and V30 cruises range between 35 and  $181 \text{ K km}^{-1}$ , the V28-11 temperature gradient is listed as  $1000 \text{ K km}^{-1}$  in IHFC 2018. Such an extreme heat gradient seems implausible (Bons et al., 2021). We attribute this to a transposed decimal place and revise the gradient to a more reasonable  $100 \text{ K km}^{-1}$ . This accordingly re-



**Table 1.** Database fields in this study.

Field [units]	Description
Site name [text]	Unique text string identification for each entry in the database. For existing IHFC 2018 sites, names are retained. Ambiguous IHFC 2018 names are made unique by adding alphanumeric suffixes based on measurement year. For new sites, names are derived from the primary literature and/or personal communications.
ID [unitless]	Unique numeric identification for each entry in the database: 1000-level denotes submarine sites, 2000-level denotes subaerial sites, and 3000-level denotes subglacial sites. Generally equivalent to the discontinued “data number” field in IHFC 2018.
IHFC status [unitless]	Code to identify each entry in the database as either “0” (existing entry in IHFC 2018), “1” (new entry, not contained in IHFC 2018), “2” (uncertain data availability at this site), or “3” (no heat flow can be calculated at this site).
Type [text]	Code to identify each entry in the database into one of three classes: “subaerial”, “subglacial”, or “submarine”.
Latitude [° N]	The decimal degree latitude for each entry in the database. For existing IHFC 2018 sites, existing latitude is retained. For new sites, latitude is derived from the published literature and/or personal communications.
Longitude [° E]	The decimal degree longitude for each entry in the database. For existing IHFC 2018 sites, existing longitude is retained. For new sites, longitude is derived from published literature and/or personal communications.
Latitude and longitude uncertainty [°]	Order-of-magnitude estimate of the positional uncertainty in each entry in decimal degrees: 0.01, 0.001, or 0.0001°. For existing IHFC 2018 sites, this uncertainty is estimated from a combination of reported decimal places and measurement year. For new sites, this is based on reported decimal places and/or personal communications.
Easting [m]	The local easting position of each entry in the database in the EPSG:3413 projection, derived from longitude using the MATLAB “polarstereo_fwd” conversion tool (Bliss, 2021).
Northing [m]	The local northing position of each entry in the database in the EPSG:3413 projection, derived from latitude using the MATLAB “polarstereo_fwd” conversion tool (Bliss, 2021).
Easting and northing uncertainty [m]	Order-of-magnitude estimate of the positional uncertainty in each entry in meters: 1100, 110, or 11 m. Converted from latitude and longitude uncertainty in decimal degrees.
Elevation [m]	Elevation above sea level of each entry. For existing IHFC 2018 sites, elevations are retained where available. For new sites, elevations are derived from the primary literature and/or personal communications. Where elevation is not available, it is interpolated from either BedMachine v3 or ETOPO1 digital elevation models (Amante, and Eakins, 2009; Morlighem et al., 2017) and noted in the comments section. For subglacial sites, this is the elevation of the ice–bed interface.
Elevation uncertainty [m]	Uncertainty in elevation above sea level of each entry. Unless otherwise noted, this is assumed to be 5 m where elevation is reported and 20 m where elevation is interpolated from a digital elevation model.
Year [CE]	Common era (CE) measurement year of each entry. For existing IHFC 2018 sites, this is retained as the “year of publication”, or the latest “year of publication”, when a range is provided. For new sites, this is derived from the published literature and/or personal communications.
Minimum depth [m]	The minimum depth, relative to site elevation, of the temperature profile used to compute heat flow for each entry in the database. For existing IHFC 2018 sites, this is retained as “minimum depth” where available. Where unavailable in IHFC 2018, it remains unavailable in this database. For new sites, this is derived from the published literature or personal communications. For subglacial sites, temperature profiles are collected above the bedrock, so depths are negative.

Table 1. Continued.

Field [units]	Description
Maximum depth [m]	The maximum depth, relative to site elevation, of the temperature profile used to compute heat flow for each entry in the database. For existing IHFC 2018 sites, this is retained as “maximum depth” where available. Where unavailable in IHFC 2018, it remains unavailable in this database. For new sites, this is derived from the published literature or personal communications. For subglacial sites, temperature profiles are collected above the bedrock, so depths are negative.
Gradient [K km <sup>-1</sup> ]	The temperature gradient used to compute heat flow for each entry. For existing IHFC 2018 sites, this is retained as “gradient” where available. Where unavailable in IHFC 2018, it remains unavailable in this database. For new sites, this is derived from the published literature or personal communications.
Conductivity [W m <sup>-1</sup> K <sup>-1</sup> ]	Thermal conductivity used to compute heat flow for each entry. For existing IHFC 2018 sites, this is retained as “conductivity” where available. Where unavailable in IHFC 2018, it remains unavailable in this database. For new sites, this is derived from the published literature or personal communications. Where conductivity has been assumed rather than measured is noted in the comments section.
Heat flow [mW m <sup>-2</sup> ]	Heat flow computed for each entry. For existing IHFC 2018 sites, this is retained as “heat flow”. For new sites, this is derived from the site gradient and site conductivity reported in the database.
Heat flow uncertainty [mW m <sup>-2</sup> ]	Uncertainty in heat flow for each entry. For both existing IHFC 2018 sites and new sites, unless otherwise stated, this is estimated as 5 % where both gradient and conductivity are reported; 10 % where only gradient is reported, and conductivity is assumed; and 15 % when only heat flow is reported without any information about gradient or conductivity.
Topographic correction [unitless]	Topographic correction for geothermal heat flow for each entry. This value represents the fraction by which local topography enhances measured heat flow. For both existing IHFC 2018 sites and new sites, where available, this is derived from an existing data product (Colgan et al., 2021). Where unavailable in this product, topographic correction remains unavailable in the database. Mean topographic correction is interpolated within the positional uncertainty in each entry.
Uncertainty in topographic correction [unitless]	Uncertainty in topographic correction for geothermal heat flow for each entry in the database. For both existing IHFC 2018 sites and new sites, this is derived from the Colgan et al. (2021) product where available.
Topographically corrected heat flow [mW m <sup>-2</sup> ]	Topographically corrected geothermal heat flow for each entry, where topographic correction is available.
Uncertainty in topographically corrected heat flow [mW m <sup>-2</sup> ]	Uncertainty in topographically corrected geothermal heat flow for each entry, where topographic correction is available.
Source [text]	This is the source of the heat flow value for each site. For existing IHFC 2018 sites, this is listed as “IHFC 2018”. For new sites, this is either listed as the individual who is most familiar with site-specific calculation or “as published” for previously published values.
Comment [text]	This field contains relevant additional notes for each entry. This includes ice thickness for subglacial sites, heat flow corrections where available, assumptions about conductivity, source of elevation data, or edits made to IHFC 2018 data. This field also aggregates information in less frequently used fields of the IHFC 1975 structure, such as “heat production” and “number of sites”.
Reference [text]	This is the most relevant article or report discussing the temperature profile, or other metadata, for each entry. These references have been updated since IHFC 2018 where possible. The most relevant reference will not necessarily contain a heat flow estimate for the site, but it serves as the best starting point to learn more about a site.

assesses the V28-11 heat flow to  $87 \text{ mW m}^{-2}$ . In comparison, the nearby V30-96 heat flow measurement is  $52 \text{ mW m}^{-2}$ .

There are nine other heat flow measurements where the product of reported gradient and conductivity is more than  $\pm 2 \text{ mW m}^{-2}$  different from the reported heat flow. In these instances, heat flow has been revised to reflect the product of reported gradient and conductivity. This approach assumes that gradient and conductivity are the primary measurements from which heat flow is a secondarily derived product. The discrepancy of this reassessment of reported gradient and conductivity product is greatest at Ymer 80-133, where heat flow is reassessed from 442 to  $120 \text{ mW m}^{-2}$ . Table 2 summarizes all instances of reassessed IHFC 2018 heat flows. Generally, these reassessments can be described as down-revising extreme heat flow values from oceanographic surveys conducted in the vicinity of the Mid-Atlantic Ridge in the late 1970s and early 1990s. These reassessments are also noted in the comments section of each entry in the database.

This study adds significant metadata to many of the existing IHFC 2018 entries. First, all 290 existing IHFC 2018 entries are coded as either “subaerial”, “submarine”, or “subglacial”. Heat flow uncertainties are also estimated for all 290 sites. Where site-specific measurements of both temperature gradient and thermal conductivity are available, an uncertainty of  $\pm 5 \%$  is assumed. Where only site-specific temperature gradient is measured, and thermal conductivity is assumed, an uncertainty of  $\pm 10 \%$  is assumed. Where only heat flow is reported, without a specific temperature gradient or thermal conductivity, an uncertainty of  $\pm 15 \%$  is assumed. This approach is applied to all Type 0 (and Type 1) sites for which formal uncertainties are not previously reported. For 50 of the existing Type 0 IHFC 2018 sites, a previously unavailable elevation of measurement is interpolated from a digital elevation model. Within the immediate proximity of Greenland, the higher-resolution BedMachine v3 elevation model is used (150 m horizontal resolution), while for areas more distal to Greenland, the lower-resolution ETOPO1 elevation model is used (1 arcmin or  $\sim 1.8 \text{ km}$  horizontal resolution; Amante, and Eakins, 2009; Howat et al., 2014; Morlighem et al., 2017). For a further 10 sites, corresponding to the “OU” cruise, measurement elevations were revised from a positive water depth to a negative ocean bottom elevation. Finally, for 51 of these existing sites, a previously unavailable measurement year is now estimated from the IHFC 2018 “year of publication” field.

In addition to these batch additions of metadata, we also revised existing site-specific metadata in several instances. This includes renaming two non-unique “Akureyri” sites as “Akureyri73” and “Akureyri91”, non-unique “Grundarfjörður” sites as “Grundarfjörður91A” and “Grundarfjörður91B”, and finally two non-unique “Hvammstangi” sites as “Hvammstangi73” and “Hvammstangi91”. The numeric suffixes are based on measurement year. We remove an apparent IHFC 2018 double entry of “V29-155”. For the GRIP (Greenland Ice core Project) ice core site, the measured el-

evation is revised from sea level (0 m elevation) to the observed ice-bed interface at 203 m elevation. For GRIP, the heat flow was also reassessed with the approach described in Sect. 2.2.2 to be consistent with the heat flow assessed for other subglacial sites.

## 2.2.2 Type 1 – not included in IHFC 2018

Our database includes 129 heat flow values that were not included in IHFC 2018 (Fig. 1). Of these, 54 heat flow values are collected from previously published gray-literature reports or peer-reviewed articles (Classen, 1977; Colebeck and Gow, 1979; Van Tatenhove and Olesen, 1994; Müller et al., 2006; Taylor et al., 2006; Damm, 2010; Harper et al., 2011; Rysgaard et al., 2018; Hartikainen et al., 2021). In the database, the “source” field of these heat flow values is listed as “as published.” The remaining 75 heat flow values are previously unreported. While heat flow values have been previously published at a handful of these sites, we reassess these heat flows using a consistent methodology. In the database, the “source” field of these heat flow values is listed as the author of this study who is most familiar with this calculation. Below, we provide a brief overview of these 75 heat flow values.

The first type of new heat flow measurements in the database is submarine measurements (44 of 75 new values). Of these, 30 measurements were collected during cruise 2005-040 of the Canadian Coast Guard Ship *Hudson*. These measurements comprise east–west and north–south transects in water depths of between 1015 and 2550 m within the Davis Strait. These heat flow values were calculated from site-specific measurements of temperature gradient and thermal conductivity measured by a 4 m heat probe deployed in the uppermost layers of submarine sediments. For these heat-probe measurements, site-specific heat flow uncertainty is propagated from gradient and conductivity uncertainties.

The remaining 14 new submarine measurements are derived from deep marine exploration wells and cored boreholes drilled offshore West Greenland between 1976 and 2011. These wells are between 1148 and 4385 m deep and at water depths between 104 and 1508 m. For these sites, the temperature gradient is calculated from Horning-corrected bottom-hole temperature and an assumed top-hole temperature of  $4^\circ\text{C}$ , reflecting assumed bottom water temperature. We assume that any seasonal cycle in bottom water temperatures at these sites is small in comparison to the temperature difference between the top- and bottom-hole temperatures. For the 14 deep exploration wells presented here, the average top-to-bottom temperature difference is  $86^\circ\text{C}$ . In the absence of a method to systematically convert borehole stratigraphy into depth-integrated conductivity, we assume a bulk thermal conductivity of  $2.00 \text{ W m}^{-1} \text{ K}^{-1}$  for all these deep boreholes. This approximates the back-calculated bulk thermal conductivity inferred by Rolle (1985) at five of these wells. We ac-

**Table 2.** Reassessed heat flows between IHFC 2018 and this study.

Site	Latitude [° N]	Longitude [° E]	IHFC 2018 [mW m <sup>-2</sup> ]	This study [mW m <sup>-2</sup> ]	Change [mW m <sup>-2</sup> ]	Change [%]
V28-11	63.750	-28.970	871	87 ± 4	-784	-90
787	63.367	-27.333	280	278 ± 14	-2	-1
807	67.750	-18.500	197	241 ± 12	+44	+22
AS-151	77.000	-3.667	78	86 ± 4	+8	+10
AS-210	72.967	-7.433	193	196 ± 10	+3	+2
Svalbard84-42	78.828	4.499	117	122 ± 6	+5	+4
Ymer 80-131	81.108	3.272	343	125 ± 6	-218	-64
Ymer 80-132	80.786	5.107	212	113 ± 6	-99	-47
Ymer 80-133	79.997	-2.363	442	120 ± 6	-322	-73
Ymer 80-134	79.853	-1.175	147	111 ± 6	-36	-24
GRIP	72.600	-37.600	51	61 ± 2	+10	+20

cordingly assume a 10 % uncertainty in heat flow at these sites.

Issler and Beaumont (1987) previously estimated heat flows at five of these deep wells (Hellefisk-1, Ikermiut-1, Kangâmiut-1, and Nukik-1/2) that are generally higher than our heat flows (Table 3). While Issler and Beaumont (1987) do not provide the gradients or conductivities used in their calculations, the gradients we calculate are nearly identical to those published contemporaneously by Rolle (1985). We therefore speculate that the disparity between our heat flows and those of Issler and Beaumont (1987) results from differences in assumed conductivity.

The second type of new measurements in the database is subaerial measurements. This category specifically denotes ice-free onshore sites (11 of 75 new values). They are generally derived by applying an assumed thermal conductivity to a measured borehole temperature gradient. At three sites (DH-GAP01, Skaergaard\_89-09B, and Skaergaard\_90-11), thermal conductivity has also been measured (Balling and Brooks, 1991; Harper et al., 2011). At these sites, we estimate the uncertainty in heat flow from site-specific uncertainties in gradient and conductivity.

At the remaining eight sites, thermal conductivity has not been measured, so we must therefore assume a value (Roethlisberger, 1961; Dam and Christiansen, 1994; Van Tatenhove and Olesen, 1994; Bjerager et al., 2018). At sites located within the Precambrian Shield of southern Greenland (Alcoa\_Site7e-I, Alcoa\_Site7e-P, Alcoa\_Site6g-P, SIS2019-02) we assume a thermal conductivity of 2.50 W m<sup>-1</sup> K<sup>-1</sup>, based on the thermal conductivity measured near Kangerlussuaq (Harper et al., 2011). At sites located within the basaltic intrusions stretching across central Greenland from Disko Island to Geikie Plateau (G02\_Paakitsoq, Marraat-1, Blokely-1), we assume a thermal conductivity of 2.25 W m<sup>-1</sup> K<sup>-1</sup>, based on thermal conductivity measured in the Skaergaard Formation (Balling and Brooks, 1991). At Thule (Thule\_1002feet) we similarly assume a thermal conductivity of 2.25 W m<sup>-1</sup> K<sup>-1</sup>, although with no guidance from

measurements in analogous geological formations (Roethlisberger, 1961; Davies et al., 1963; Dawes, 2009). At all eight sites, we assume an uncertainty in heat flow of 10 %.

For all subaerial sites, where possible, we calculate temperature gradients below 75 m depth. As recent near-surface warming can decrease the near-surface temperature gradient and thus decrease near-surface heat flow in comparison to deeper values, we prefer not to use ground temperature data from depths shallower than 75 m. This approach minimizes the influence of recent, meaning post-1990, pronounced atmospheric warming on the near-surface temperature gradient (Balling and Brooks, 1991). This can be considered a basic paleoclimatic correction to remove the influence of recent climate change over the past century (Beltrami and Mareschal, 1991; Mareschal and Beltrami, 1992).

The third type of new measurements in the database is subglacial measurements (20 of 75 new heat flow values). This category specifically denotes onshore sites located beneath glaciers. Heat flow has been previously assessed at four of these sites (Table 3). For all subglacial sites, heat flow is calculated using a slightly modified version of the temperature gradient approach. We use ice temperature profiles from the Greenland Ice Borehole Temperature Profiles database (Mankoff, 2021; [https://github.com/GEUS-PROMICE/greenland\\_ice\\_borehole\\_temperature\\_profiles](https://github.com/GEUS-PROMICE/greenland_ice_borehole_temperature_profiles), last access: 8 February 2022). At sites where the ice temperature profile is complete, meaning it spans the full ice thickness from surface to bed (Devon73, PrinceWales05, Meighen67, Agassiz77/79A/79B/84, Penny96, CampCentury, DYE-3, GISP2, NEEM, HansTausen\_Dome, HansTausen\_Hare), we fit a second-order temperature–depth function in the bottom 10 % of the borehole (Paterson, 1968; Weertman, 1968; Paterson et al., 1977; Fisher and Koerner, 1984; Gundestrup and Hansen, 1984; Clarke et al., 1987; Fisher et al., 1988; Clausen et al., 2001; Kinnard et al., 2008; Rasmussen et al., 2013; MacGregor et al., 2015). We then adopt the temperature gradient at the deepest 1 % of this second-order polynomial fit. This

**Table 3.** Reassessed heat flows between previous studies and this study.

Site	Latitude [° N]	Longitude [° E]	Previous study [mW m <sup>−2</sup> ]	This study [mW m <sup>−2</sup> ]	Change [mW m <sup>−2</sup> ]	Change [%]	Reference
Hellefisk-1	67.8781	−56.7392	50	49 ± 5	−1	−2	Issler and Beaumont (1987)
Ikermiut-1	66.9367	−56.5906	51	58 ± 6	+7	+13	Issler and Beaumont (1987)
Kangâmiut-1	66.1503	−56.1900	55	54 ± 5	−1	−2	Issler and Beaumont (1987)
Nukik-1	65.6317	−54.7669	56	42 ± 4	−14	−25	Issler and Beaumont (1987)
Nukik-2	65.5267	−54.7606	54	39 ± 4	−15	−28	Issler and Beaumont (1987)
CampCentury	77.1797	−61.1097	42	40 ± 2	−2	−5	Hansen and Langway (1966)
			40		0	0	Weertman (1968)
GRIP	72.600	−37.600	51	61 ± 2	+10	+20	Dahl-Jensen et al. (1998)
			59		+2	+3	Martos et al. (2018)
NEEM	77.45	−51.07	58	65 ± 5	+7	+12	Martos et al. (2018)
DYE-3	65.18	−43.82	25	26 ± 3	+1	+4	Greve (2005)

approach standardizes the approximation of temperature gradient across sites by accounting for differences in the depth interval of temperature measurements. It also acknowledges the characteristic non-linearity of basal ice temperature profiles (Hooke, 2019).

At some subglacial sites, however, the ice temperature profile is incomplete, meaning it does not reach the ice–bed interface (Devon72/98, Renland88, FladeIsblink06) (Paterson et al., 1977; Hansson, 1994; Kinnard et al., 2006; Lemark and Dahl-Jensen, 2010). At these sites, we must extrapolate the temperature profile to the ice–bed interface to approximate the temperature gradient in the deepest 1 % of the borehole. For this extrapolation, we generate a second-order polynomial fit to the deepest portion of the measured temperature–depth profile that is the same thickness as the depth range that is missing measurements, i.e., the depth range that must be extrapolated. For example, if ice temperatures are not available in the bottom 10 % of the borehole, then we fit the temperature–depth profile to the deepest available 10 % of the borehole where ice temperatures are available: the bottom 20 % down to the bottom 10 % of the borehole. This ensures a 1 : 1 ratio between the observation and extrapolation depth increments. To quantify the uncertainty associated with this extrapolation, we then repeat the extrapolation of the deepest 1 % gradient 10 times but adjust the shallowest depth of the data range up the ice column 1 % each time. We make available a sample of this code for the Devon72 extrapolation (Fig. 2) in the provided database associated with this article.

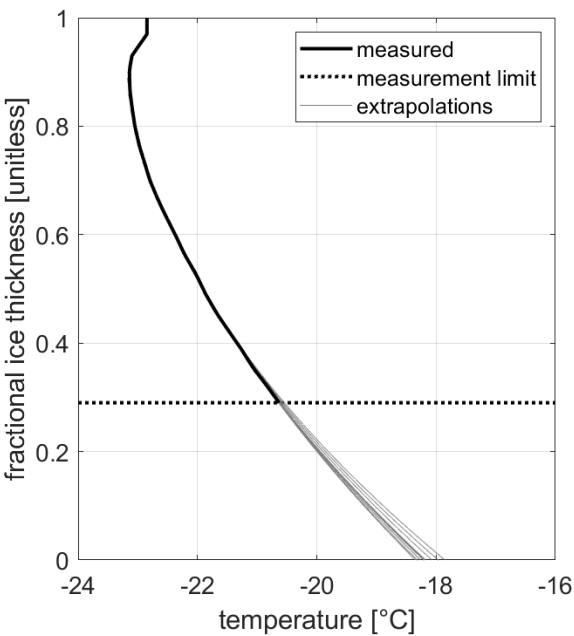
For all subglacial boreholes, thermal conductivity ( $\kappa$ ; in  $\text{W m}^{-1} \text{K}^{-1}$ ) is estimated as a function of ice temperature ( $T_i$ ; in K) based on the following relation (Yen, 1981):

$$\kappa = -0.013T_i + 2.1.$$

(1)

We prescribe the thermal conductivity of ice based on the ice temperature in the bottom 1 % of the borehole.

Four subglacial sites have unique glaciological settings requiring further explanation. Basal melting presently occurs at both NGRIP (North GREENland Ice core Project) and NEEM.



**Figure 2.** Extrapolating temperature gradient at the ice–bed interface in the Devon72 borehole, where ice temperature measurements are not available in the deepest 29 % of the ice column.

This loss of basal ice is evident in local radiostratigraphy and depth–age relations at both sites. NGRIP has substantial basal melting, which prevents application of the basal-temperature-gradient approach described above. For this site, we estimate basal heat flow as  $129 \pm 30 \text{ mW m}^{-2}$ , based on the published range of values based on thermomechanical ice modeling (Dahl-Jensen et al., 2003; Greve, 2005; Buchardt and Dahl-Jensen, 2007). At NEEM, where there is evidence of trace basal melting, the basal-temperature-gradient approach is still valid, but we add a  $+5 \pm 5 \text{ mW m}^{-2}$  correction to account for an estimated  $0.5 \text{ mm yr}^{-1}$  of basal melt there (Rasmussen et al., 2013). At Tuto\_D-11, the ice thickness is



only 48 m. We therefore use a simple linear relation to constrain the temperature gradient in the lowermost 33 m of the ice, within a generous uncertainty from a digitized version of the published temperature profile (Davis, 1967).

Finally, at HansTausen\_Hare, the borehole is drilled in a relatively high-deformation setting, where the ice surface speed is  $\sim 20 \text{ m yr}^{-1}$  with an ice thickness of 289 m (Clausen et al., 2001; Reeh et al., 2001). This situation implies that deformational heating is a significant heat source in the deepest part of the ice column there. Assuming  $n = 3$  deformational and isothermal ice flow, the bottom 1 % (3 m) of the HansTausen\_Hare borehole has an along-flow strain rate of  $0.27 \text{ yr}^{-1}$  (Hooke, 2019). Assuming an overburden confining stress of 2.6 MPa ( $286 \text{ m} \times 917 \text{ kg m}^{-3} \times 9.8 \text{ m s}^{-2}$ ), this calculation yields  $66 \text{ mW m}^{-2}$  of deformational heating ( $3 \text{ m} \times 0.27 \text{ yr}^{-1} \times 2.6 \text{ MPa}$ ) in the bottom 3 m of the HansTausen\_Hare borehole (Marshall et al., 2005). We therefore apply a  $-66 \pm 33 \text{ mW m}^{-2}$  correction to account for significant basal deformational heating at HansTausen\_Hare, but we note that the geothermal gradient is likely also influenced by deformational heating higher in the ice temperature profile. Comparison with the upstream HansTausen\_Dome borehole, which is unaffected by deformational heating, suggests that a deformation heating correction of closer to  $-100 \text{ mW m}^{-2}$  may be appropriate at HansTausen\_Hare.

### 2.2.3 Type 2 – no heat flow (uncertain data)

Our database includes 66 entries where at least one borehole has been drilled to  $> 100 \text{ m}$  depth but for which we are presently uncertain of the availability of borehole temperature data (Fig. 1). While no heat flow values are presently available for these sites, the possibility exists that sufficient data may become available to assess heat flow values in future studies. Where multiple deep boreholes are known to exist at a single site, the number of boreholes is noted in the ‘Comment’ field of the database. For all these Type 2 entries, the database includes metadata showing the site or borehole name, type of site, characteristic drill year, and decimal degree and EPSG:3413 positional and elevation data.

These sites were identified from the Greenland National Petroleum Data Repository (GNPDR) (<http://greenpetrodata.gl>, last access: 8 February 2022; formerly the ‘GEUS Oil & Gas’ database), the Greenland Mineral (GreenMin) database (<http://www.greenmin.gl>, last access: 8 February 2022; formerly the ‘GEUS SAMBA’ database), and also the International Ocean Discovery Program (IODP) drilling database (<https://iodp.tamu.edu/scienceops/maps.html>, last access: 8 February 2022). For several GNPDR and GreenMin sites, a comprehensive evaluation of non-digitized hard-copy reports would likely yield temperature gradients omitted from our preliminary evaluation of digitized reports. We appeal to persons with site-specific knowledge of the availability of temperature profiles at these Type 2 sites to contact our team

to help us parse these sites as either Type 1 or Type 3 in future versions of this database.

### 2.2.4 Type 3 – no heat flow (insufficient data)

Finally, our database also includes entries where a borehole has been drilled, and a heat flow cannot be calculated with presently available data and methods. Our database contains 74 of these Type 3 sites (Fig. 1). Thirty-three of these sites appear in IHFC 2018 but have no associated heat flow value or primary gradient and conductivity values from which to calculate heat flow. All but one of these sites is submarine. At a combination of 27 subaerial and submarine sites, we ascertain through end-of-well reports or personal communications that no temperature profile was collected in the borehole. Finally, at 14 subglacial sites, an ice temperature profile has been collected, but it is not possible to use this profile to calculate heat flow for one or two reasons. First, if the temperature profile is not deep enough to make a reasonable extrapolation of the temperature gradient at the ice–bed interface (i.e., Site II), then heat flow cannot be estimated from the temperature gradient approach. Second, if the basal ice is at the pressure melting point additional glaciological data are needed to characterize basal frictional heating, which warms ice temperatures, and/or basal melting, which cools ice temperatures (Karlsson et al., 2021). At temperate sites, where frictional heating or basal melting is not constrained, heat flow cannot be estimated using the standard basal-temperature-gradient approach due to complex thermodynamics associated with ice–water phase changes (i.e., Jakobshavn\_A).

Despite these limitations, we consider it important to inventory Type 3 subaerial and subglacial boreholes as they may be useful for resolving future heat flow values. Where subaerial boreholes have been capped with metal sealers, as opposed to filled with concrete, it may be possible to reopen them and insert thermistor strings (Balling and Brooks, 1991). Instrumenting abandoned boreholes drilled into consolidated bedrock is unconventional but could be an inexpensive opportunity to rapidly expand the number of reliable Greenlandic heat flow measurements. Any frozen drilling fluid in these boreholes, while extremely dirty in comparison to glacier ice, should be significantly easier to penetrate in comparison to surrounding bedrock. Perhaps similarly, while it is not presently possible to resolve a geothermal heat flow estimate from subglacial temperature profiles near the pressure melting point using the basal-temperature-gradient approach, future methodological improvements may allow heat flow to be inferred where complex water–ice phase changes are present (Colebeck and Gow, 1979; Iken et al., 1993; Lüthi et al., 2015; Doyle et al., 2018).

### 2.3 Topographic correction

The database includes an explicit correction for the influence of topography on geothermal heat flow for all heat flow measurement sites (Types 0 and 1). This topographic correction accounts for elevated heat flow in valleys and correspondingly diminished heat flow along ridge lines (Lees, 1910). We interpolate this site-specific correction from the geostatistical product of Colgan et al. (2021), which is based on the BedMachine v3 digital elevation model (Morlighem et al., 2017). As the BedMachine domain covers only part of the larger Greenland domain of this study, this topographic correction is only available for ca. 34 % of heat flow measurement sites. However, this subregion does include all subaerial and subglacial sites in Greenland and most submarine sites on the Greenlandic continental shelf. Most of the sites for which this systematic topographic correction is not available may be considered abyssal submarine sites, meaning beyond the continental shelf, where topographic variation is generally more subdued compared to subaerial or subglacial sites. Within our Greenland domain, this topographic correction ranges from a minimum of  $-21 \pm 5$  % at Hole 38 in Ilímaussaq, southern Greenland (Sass et al., 1972), to  $+55 \pm 10$  % at the Dybet site in Young Sound, East Greenland (Rysgaard et al., 2018). These end-member sites highlight the potential importance of acknowledging topographic influence on local heat flow when interpreting heat flow measurements. Critically, however, the database only provides these topographically corrected heat flow values as a supplement to measured heat flow values. The machine learning analysis that we perform to interpolate a regularly spaced heat flow map across our Greenland domain (Sect. 2.4) uses uncorrected heat flow measurements. We discuss other heat flow corrections of concern in the Greenland context in Sect. 4.1 and 4.2.

### 2.4 Greenland heat flow map

We derive a spatial heat flow map across our Greenland domain using a machine learning approach that combines the heat flow measurements described above with other geophysical datasets. We employ the machine learning approach to estimate geothermal heat flow at the lithospheric surface, meaning the subaerial, subglacial, or submarine plane. This approach was initially presented for Greenland by Rezvanbehbahani et al. (2017) and subsequently revised for Antarctica by Lösing and Ebbing (2021) (<https://github.com/MareenLoesing/GHF-Antarctica-MachineLearning>, last access: 8 February 2022). Lösing and Ebbing (2021) enhanced the machine learning algorithm by using an advanced and more regularized gradient boosting regression and provided more detailed evaluation of the influence of regional and global geophysical datasets. This evaluation showed the added value of applying well-constrained regional data as global datasets often have a high uncertainty in polar re-

gions. In total, 12 features are used for the machine learning algorithm (Table A2), 3 of which are boundary layers: the topography, the crust–mantle boundary (Moho depth), and the lithosphere–asthenosphere boundary (LAB). A magnetic susceptibility model includes crustal constraints. Seismological information is added to the model as a tectonic regionalization, calculated from a tomography model. In addition, the vertical magnetic field and the gravity field, the latter represented by its mean curvature, are included. Finally, the predicted geothermal heat flow also depends on the distance to five major tectonic elements (trenches, ridges, transform faults, young rifts, and volcanoes). More technical descriptions on the method can be found in Rezvanbehbahani et al. (2017) and Lösing and Ebbing (2021), and a graphical overview of these datasets is provided in the Appendix of this study.

Following Lösing and Ebbing (2021), we optimize a global supervised-machine-learning regression approach by incorporating regional datasets best suited for Greenland (Fig. A1). We combine the heat flow measurements described above with the global point dataset used by Lösing and Ebbing (2021). All the global or regional predictive geophysical datasets are similarly interpolated from their native resolution to a common 55 km grid, which defines the fundamental resolution of the final heat flow solution. For some datasets, this means increasing the spatial resolution from coarser native resolutions using bilinear interpolation, while for other datasets this means decreasing the spatial resolution from finer native resolutions using area averaging. The heat flow measurements are similarly binned into 55 km cells. Where multiple heat flow measurements exist within a 55 km single cell, they are averaged. We exclude heat flow measurements  $> 200 \text{ mW m}^{-2}$  as these high values are representing local anomalies rather than regional heat flow at the 55 km scale. We also exclude four onshore measurements where heat flow is strongly influenced by various local processes (DH-GAP01, Marraat-1, Alert\_203-1, HansTausen\_Hare; Sect. 4.2).

To ensure globally consistent results, the machine learning algorithm is trained with global data, and the prediction is also global. The training dataset for this algorithm contains 80 % of all available global heat flow measurements, but we include 100 % of all available heat flow measurements within the Greenland domain. As this study focuses on heat flow in and around Greenland, we present only the Greenland domain of this simulation. The Greenland domain contains both continental and oceanic crust, which are known to have markedly different thermal characteristics and geological histories (Dawes, 2009). We therefore run the algorithm separately for each crust type to optimize heat flow prediction. For the continental simulation, we assign the measurements associated with the present-day Iceland plume to the remaining 20 % of the testing dataset. This removes regions of active volcanism from the training data but maintains regions of paleoplume activity within the training data.

We also employ a jackknife resampling method with the 60 available continental measurements. This means we calculate 60 individual simulations, and in each simulation one of the Greenland onshore heat flow points is left out from the machine learning training dataset. From this simulation ensemble, minimum, mean, and maximum heat flow estimates are calculated (Fig. 3). This jackknifing indicates that the magnitude and spatial distribution of continental heat flow are disproportionately sensitive to the inclusion or exclusion of the relatively high and uncertain heat flow measurement at NGRIP. No other point influences the predicted heat flow results to such a degree.

The jackknifing ensembles suggest that the NGRIP measurement is not representative of the regional background lithospheric heat flow being simulated by the machine learning approach. Simply put, there is no plausible source for high heat flow at NGRIP in the 12 input geophysical datasets provided to the machine learning algorithm (Fig. A1). However, NGRIP and the presence of the Northeast Greenland Ice Stream strongly suggest elevated heat flow in North Greenland. Given appreciable community interest in the NGRIP anomaly, we follow Rezvanbehhahani et al. (2017) and run the machine learning algorithm with training datasets that both include and exclude NGRIP. We make both these “with” and “without” NGRIP heat flow maps available in the database and describe the influence of the NGRIP anomaly on the machine learning approach in Sect. 3.2. Generally, while the location of the NGRIP measurement is unique within Greenland for spatial interpolations, from a machine learning perspective, the relations between observations and geophysical fields are more important than spatial relations. In this sense, our machine learning algorithm does not consider NGRIP a spatial outlier but rather a geophysical outlier; the heat flow at NGRIP is not consistent with other observations of heat flow in similar settings. This suggests that the NGRIP heat flow anomaly is due to local processes that are not captured in the 12 input geophysical datasets.

For the regional overview considered here, precise placement of the continent–ocean transition and type of transitional crust is not critical, so we use the 1000 m bathymetry contour as a simple proxy to delineate oceanic and continental domains. This is a reasonable first-order approximation everywhere except along the Greenland–Iceland Ridge, where the boundary is complex. For the oceanic portion of the domain, where spatial variations in submarine topography and ocean-bottom temperatures influencing heat flow are generally more subtle than on land, we do not employ a jackknife resampling method and instead provide a single heat flow estimate. Although the continental and oceanic simulations were run separately, no edge effects are apparent along the continent–ocean transition.

Figure 4 shows the importance of the individual input variables to the machine learning algorithm for the continental and the oceanic domains of our model. The importance parameter evaluates the relative importance of each input

dataset for predicting the results. More details about the calculation and theory of the importance parameter can be found in Lösing and Ebbing (2021). For the continental simulation domain, the distance to volcanoes is the most important feature, followed by the Moho depth and the tectonic regionalization. The mean curvature of the gravity data, the susceptibility model, and the vertical magnetic field component ( $B_z$ ) are the least important features. For the oceanic simulation domain, the distance to the nearest ridge, the lithosphere–asthenosphere boundary depth, and the tectonic regionalization are the most important features. Differences in feature importance between the continental and oceanic domains support the idea that machine learning can be more precise when each domain is modeled individually. The inclusion or exclusion of NGRIP from the training data does not fundamentally shift this importance ranking of the input geophysical datasets.

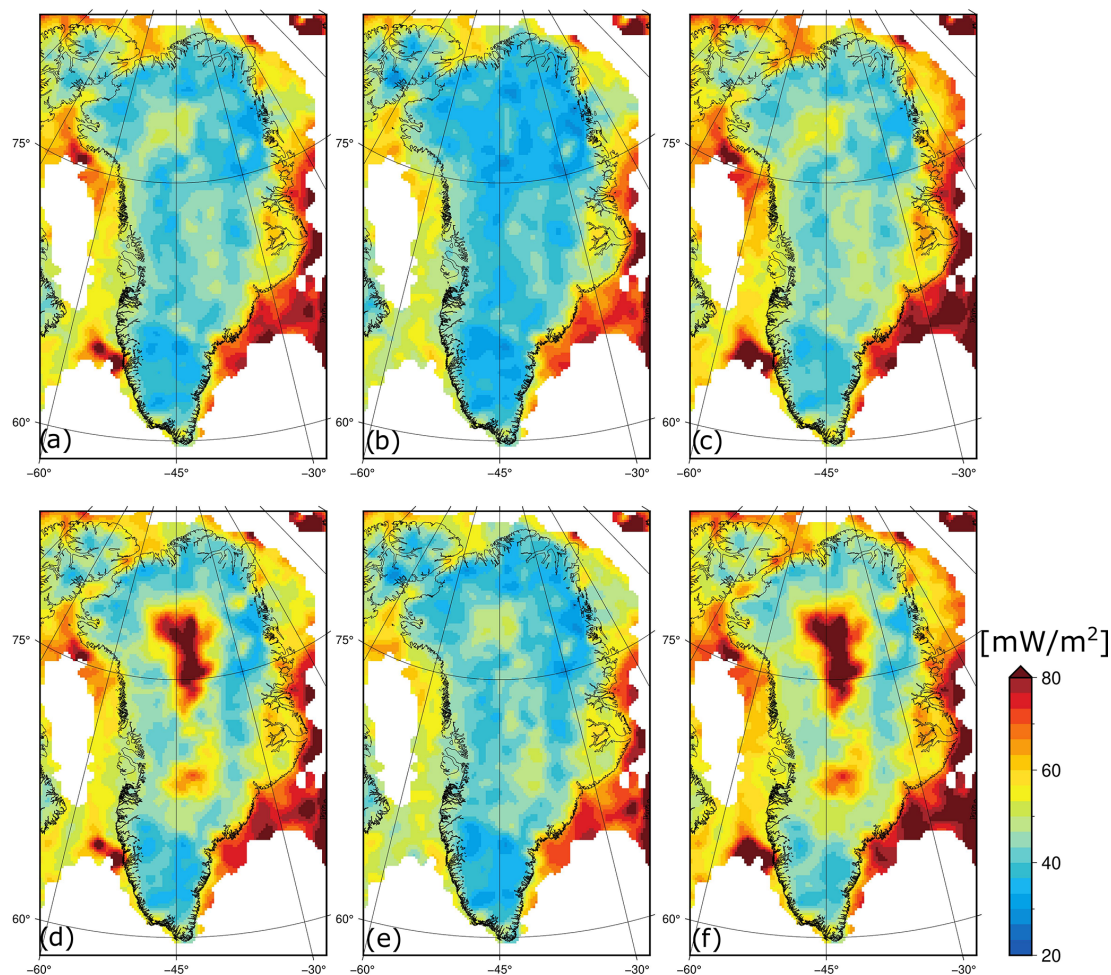
### 3 Data products

#### 3.1 Heat flow measurement database

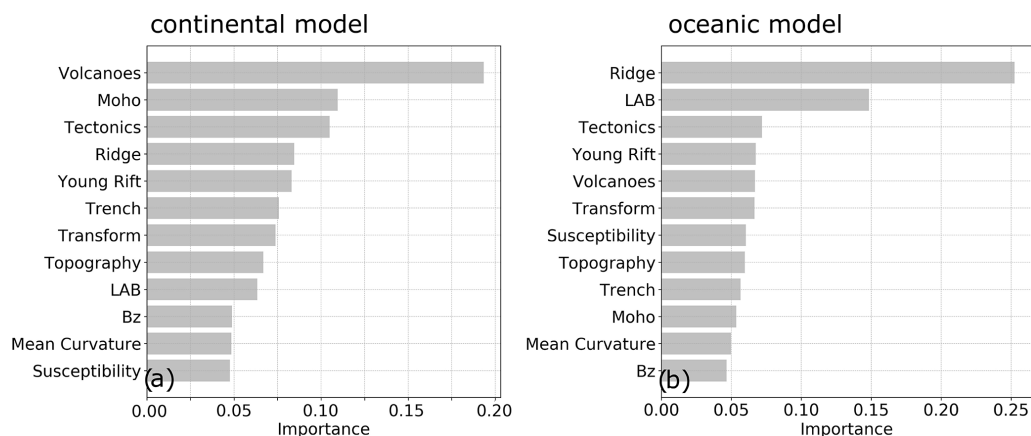
The heat flow measurement database presented here contains 290 measurements that are carried forward from IHFC 2018. Eleven of these measurements have been reassessed with new heat flow values (Table 2). The database contains a further 129 measurements that did not appear in IHFC 2018. The majority of these measurements have not been previously published elsewhere. They consist of 88 offshore measurements and 41 onshore measurements, of which 24 are subglacial. Perhaps most notably, these new measurements provide the first comprehensive sampling of heat flow in Davis Strait and Baffin Bay. The mean distance between a new measurement (Type 1) and an existing IHFC 2018 measurement (Type 0) is 251 km, with distance ranging from < 1 km (HF4-9 in Davis Strait) to 645 km (DANA06-HF93\_01 in Baffin Bay). The Greenland domain that we employ has an area of  $6.4 \times 10^6 \text{ km}^2$ , which yields a characteristic measurement density of one measurement per  $\sim 15\,000 \text{ km}^2$ .

Within the Greenland domain, the median of all heat flow measurements ( $n = 419$ ) is  $79 \text{ mW m}^{-2}$ , with a standard deviation of  $53 \text{ mW m}^{-2}$  (Fig. 5). The highest heat flow measurement is  $377 \text{ mW m}^{-2}$  at the RK2105 site on the Mid-Atlantic Ridge just north of Iceland. The lowest heat flow measurement is  $3 \text{ mW m}^{-2}$  at the DH-GAP01 site in a permafrost talik in southwestern Greenland. While there is a large range of both onshore and offshore heat flow values, a two-tailed  $t$  test, assuming two samples with unequal variance, suggests that the population of offshore heat flows (median  $85 \text{ mW m}^{-2}$  and standard deviation  $52 \text{ mW m}^{-2}$ ) is significantly ( $p < 0.05$ ) warmer than the population of onshore heat flows (mean  $58 \text{ mW m}^{-2}$  and standard deviation  $55 \text{ mW m}^{-2}$ ). This difference can be attributed to the more intensively sampled elevated heat flow in the vicinity of the





**Figure 3.** Machine learning results highlighting spatial variability due to the jackknifing approach within the continental portion of the domain. The mean (**a**, **d**), minimum (**b**, **e**), and maximum (**c**, **f**) geothermal heat flow calculated from  $n = 59$  jackknifing simulations without NGRIP (**a**, **b**, **c**) and  $n = 60$  jackknifing simulations with NGRIP (**d**, **e**, **f**).



**Figure 4.** The importance of the 12 input variables used in the machine learning (**a**) for the continental model domain and (**b**) for the oceanic model domain. “ $B_z$ ” denotes the vertical magnetic field component.

Mid-Atlantic Ridge, in the eastern portion of our Greenland domain. Of the  $n = 419$  heat flow measurements, 53 % have a measurement uncertainty of  $< 5 \text{ mW m}^{-2}$ , 40 % have a measurement uncertainty of between 5 and  $20 \text{ mW m}^{-2}$ , and 7 % have a measurement uncertainty of  $> 20 \text{ mW m}^{-2}$  (Fig. 6).

### 3.2 Greenland heat flow map

Combining multiple simulations from the machine learning algorithm provides a seamless heat flow map across both continental and oceanic areas around Greenland (Fig. 7). This seamless heat flow map represents the mean, or “best estimate”, of geothermal heat flow across the domain. Our “without NGRIP” heat flow simulation suggests a mean onshore heat flow across Greenland of  $44 \text{ mW m}^{-2}$ , ranging from  $28$  to  $76 \text{ mW m}^{-2}$  across the country. Aside from a heat flow anomaly of up to  $100 \text{ mW m}^{-2}$  in central North Greenland, our “with NGRIP” heat flow simulation is broadly similar to the “without NGRIP” simulation. The presence of the NGRIP anomaly, however, increases the mean onshore heat flow across Greenland to  $48 \text{ mW m}^{-2}$ . Generally, the range between maximum and minimum heat flow simulations is  $< 20 \text{ mW m}^{-2}$  for most continental areas. The heat flow anomaly around NGRIP is caused by the machine learning algorithm classifying the anomaly region as broadly similar to NGRIP, based on the 12 input geophysical datasets.

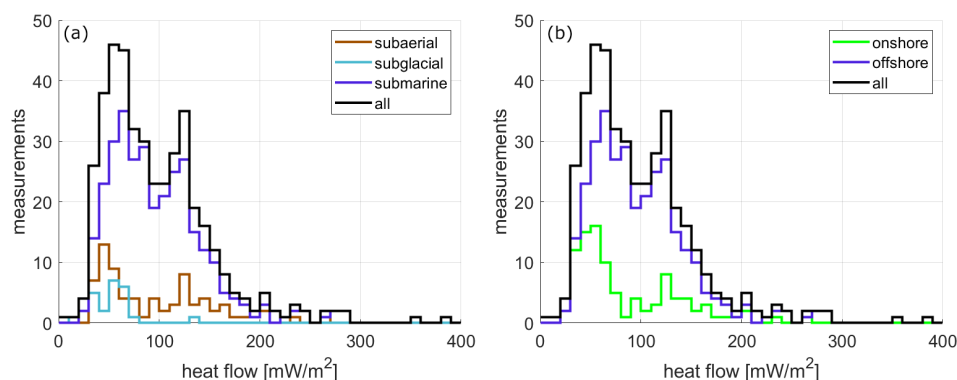
Aside from the NGRIP anomaly, the most distinctive onshore heat flow feature is the relatively low heat flow within the North Atlantic Craton of southern Greenland. As the North Atlantic Craton is an old Archaean block, it is expected to have an average surface heat flow significantly less than younger continental terrains (Goes et al., 2020). Apart from the NGRIP anomaly, the highest onshore heat flow is in east-central Greenland. This positive anomaly, or warm bias, in heat flow is attributable to proximity to the Mid-Atlantic Ridge. Offshore, there is a clear asymmetry to the east and west of Greenland. West of Greenland, in Davis Strait and Baffin Bay, heat flow is generally similar to continental values with some indications of near-shore warm anomalies. However, these warm anomalies may be due to our approximate delineation between continental and oceanic crust types. East of Greenland, in the North Atlantic and Greenland Sea, there is a pronounced high heat flow along the Mid-Atlantic Ridge. Offshore heat flow is enhanced throughout the Irminger Basin off southeastern Greenland.

We compare our modeled heat flow map with the measured heat flow values (Fig. 8). Generally, the residuals between measured and modeled heat flow are  $< 20 \text{ mW m}^{-2}$  in the continental portion of the domain. NGRIP, however, is a clear outlier in this  $n = 419$  site comparison. In the oceanic portion of the domain, the residuals are typically larger. This asymmetry may be attributable to our coarse resolution of the continent–ocean crustal boundary or differences in the accuracy and resolution of geophysical datasets in the onshore

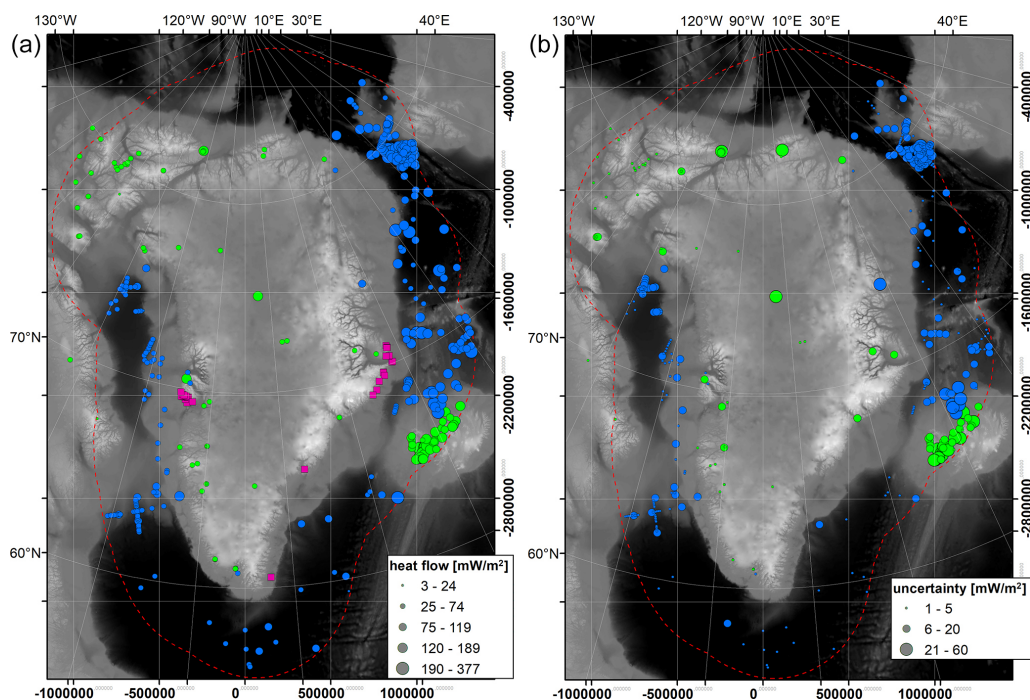
and offshore portions of the domain. Our relatively poor delineation of the continent–ocean crustal boundary inevitably results in some oceanic measurements lumped into the continental simulations, and vice versa. These larger residuals suggest that available geophysical datasets do not capture local processes offshore, especially along the Mid-Atlantic Ridge, as well as in some onshore areas. In other words, the global or regional datasets upon which the machine learning algorithm depends have insufficient resolution to capture the variety of local processes reflected in heat flow measurements (Sect. 4.2). This is a fundamental limitation of the 55 km spatial resolution we adopted to ensure a globally consistent heat flow simulation. We expect these residuals to decrease with finer spatial resolution and the inclusion of more geophysical datasets.

The magnitude and spatial distribution of heat flow are clearly very sensitive to the inclusion of the relatively high and uncertain NGRIP heat flow measurement ( $129 \pm 30 \text{ mW m}^{-2}$ ). This machine learning outcome is similar to that highlighted by Rezvanbehbahani et al. (2017). The inclusion of NGRIP in the training data introduces substantial ensemble uncertainty in central North Greenland as the machine learning algorithm cannot reconcile the NGRIP anomaly with available input geophysical datasets. The magnitude and spatial distributions of the ensemble uncertainty ranges of the “with” and “without” NGRIP simulations clearly show that, of all onshore measurements, the machine learning algorithm is most sensitive to the inclusion or exclusion of NGRIP (Fig. 9). Our recommendation to exclude NGRIP from the training data is on the basis that it is not representative of the regional background lithospheric heat flow. The heat flow measured at NGRIP is  $\sim 86 \text{ mW m}^{-2}$  greater than the heat flow we simulate at NGRIP:  $43 \text{ mW m}^{-2}$ , with a range of 35 to  $45 \text{ mW m}^{-2}$ . However, strictly speaking, the NGRIP ice borehole measurement reflects elevated heat flow in the basal layers of the ice sheet and is not necessarily representative of thermal conditions in the underlying bedrock. A substantial portion of this  $\sim 86 \text{ mW m}^{-2}$  discrepancy between observed basal and simulated geothermal heat flows may be attributable to local subglacial hydrological processes (Gooch et al., 2016; Bons et al., 2021; Smith-Johnsen et al., 2021). For example, subglacial water flow or hot springs are much finer-scale processes in comparison to the 55 km resolution of our machine learning algorithm. We note that future additional measurements of elevated heat flows in central North Greenland may yet render NGRIP statistically representative of the broader region. We therefore make both the “with” and “without” NGRIP heat flow maps available in the database.





**Figure 5.** Histograms of heat flow measurements in the subaerial ( $n = 77$ ), subglacial ( $n = 25$ ), and submarine ( $n = 317$ ) populations (a) and the onshore ( $n = 102$ ) and offshore ( $n = 317$ ) populations (b).



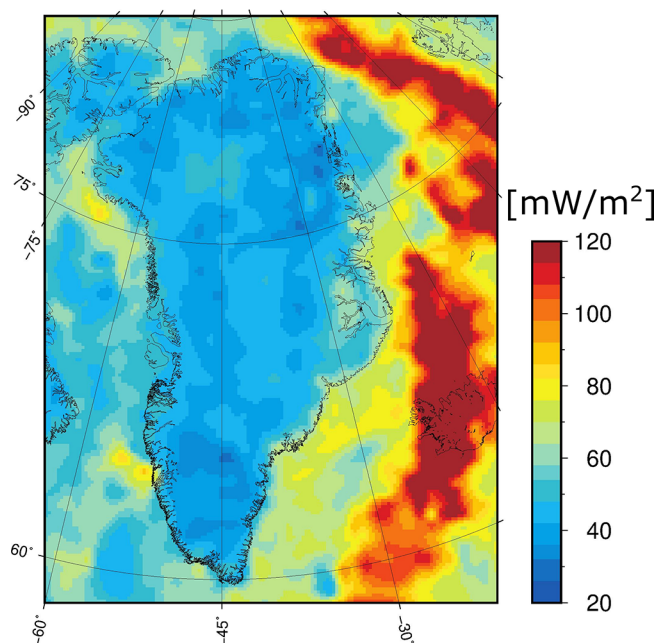
**Figure 6.** (a) Heat flow measurements, scaled by magnitude, overlaid on the ETOPO1 digital elevation model (Amante, and Eakins, 2009). Green dots denote onshore measurements ( $n = 102$ ), and blue dots denote offshore measurements ( $n = 317$ ). Magenta squares denote hot springs inventoried by Hjartarson and Armannsson (2010). (b) The same for uncertainty in heat flow measurement. Projection is EPSG:3413 (bottom  $x$  axis and right  $y$  axis).

## 4 Discussion

### 4.1 Paleoclimate correction

Ground surface temperature is an important boundary condition for geothermal flow. Increases in surface temperature generally decrease the temperature gradient and heat flow, while decreases in surface temperature generally increase the temperature gradient and heat flow. The depth of these heat flow perturbations depends on the magnitude and duration of the paleoclimatic shift. In Arctic settings, the most striking paleoclimatic heat flow perturbations are those associ-

ated with submarine-to-subaerial transition. In areas where crustal uplift causes land to emerge from the ocean, geothermal flow is greatly enhanced by the transition from a relatively warm ocean-bottom boundary temperature to a relatively cold atmospheric boundary temperature. While this heat flow anomaly is limited to low-elevation coastal fringes, it can be surprisingly pronounced. For example, a heat flow of  $148 \text{ mW m}^{-2}$  was measured in the Alert\_203-1 borehole at 5 m elevation, whereas a heat flow of  $72 \text{ mW m}^{-2}$  was measured in the Alert\_202-2 at 77 m elevation < 2 km away (Taylor et al., 2006). Similarly, the Marraat-1 borehole at



**Figure 7.** Machine learning results for the combined continental and oceanic portions of the domain. For the onshore part, the model trained without NGRIP is used. Note the different color scale in comparison to Fig. 3.

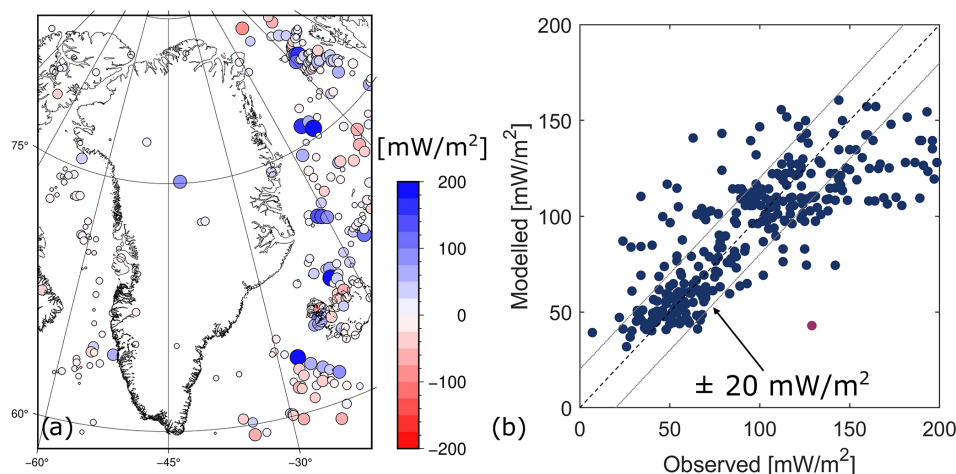
13 m elevation yielded a geothermal flow of  $132 \text{ mW m}^{-2}$  (Dam and Christensen, 1994), which is  $\sim 160\%$  greater than the  $50 \text{ mW m}^{-2}$  measured at G02\_Paakitsoq at higher elevations  $\sim 110 \text{ km}$  away (Van Tatenhove and Olesen, 1994). While the basalt and gneiss rock types are different between the Marraat-1 and G02\_Paakitsoq sites, their thermal properties are not sufficiently different to readily explain this discrepancy in local heat flows.

We demonstrate the magnitude and duration of a submarine-to-subaerial paleoclimate perturbation using a simple 1-D heat flow model. In this model, we prescribe a constant basal heat flow of  $50 \text{ mW m}^{-2}$  applied at 500 m depth and uniform thermal diffusivity of  $0.5 \text{ mm}^2 \text{ s}^{-1}$ . We then assume that the surface boundary condition shifts from a  $+4^\circ\text{C}$  submarine setting to a  $-6^\circ\text{C}$  subaerial setting at model initialization. For this submarine-to-subaerial transition, which is characteristic of the Ilulissat area, it is plausible that geothermal flows of  $> 150 \text{ mW m}^{-2}$  can persist to depths of 100 m for  $> 400$  years as freshly emerged coastal land cools in the atmosphere (Fig. 10). While it is clear that the Alert\_203-1 and Marraat-1 boreholes represent encounters with the elevated coastal heat flow anomaly associated with recent land emergence, this effect may influence other heat flow measurements in the database to a lesser degree. Caution should be exercised when interpreting heat flow measurements from land that has emerged in geologically recent times (i.e., since the Little Ice Age). While it may be difficult to interpret anomalous coastal heat flow measurements, it is

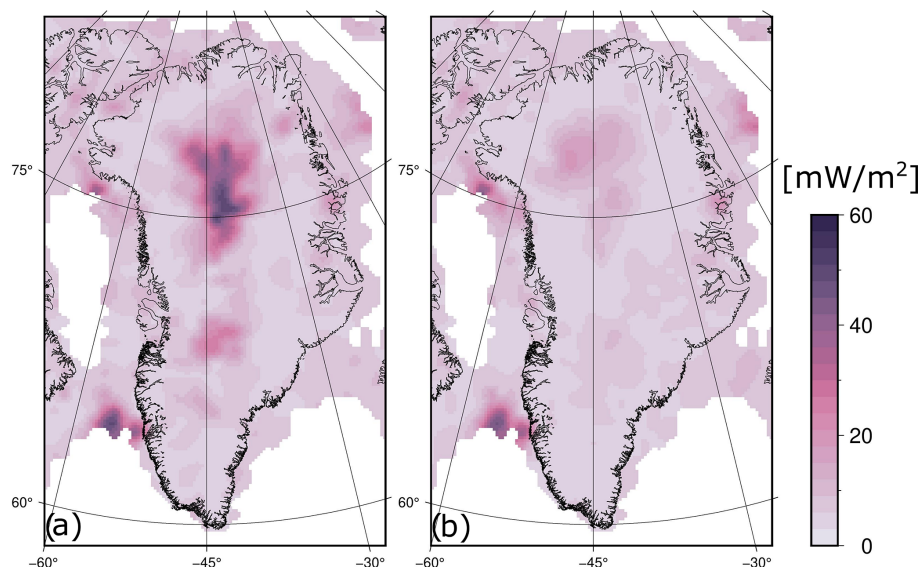
conceivable that this coastal heat flow anomaly could be harnessed as an indirect, or low-temperature, geothermal heat source for coastal settlements around Greenland.

A second paleoclimatic heat flow perturbation is that associated with subglacial-to-subaerial transitions (and vice versa), as well as transitions between cold-based (or frozen) and warm-based (or thawed) subglacial conditions. The effect of such transitions is highlighted by 2D simulations of bedrock temperature and heat flow at the DH-GAP04 borehole in West Greenland (Claesson Liljedahl et al., 2016; Hartikainen et al., 2021). For this site, an uncorrected heat flow of  $28 \text{ mW m}^{-2}$  was calculated over the 280–480 m depth range. The uppermost 280 m of borehole temperatures were discarded to avoid the effect of topography and recent variations in surface climate. Calculation of a longer-term paleoclimatic correction, one that accounts for both ice sheet history and climatic events influencing ground surface temperature, has been performed using a 2D cross-sectional simulation and site-specific data for the past 100 ka (Hartikainen et al., 2021). This period includes a full glacial cycle, during which DH-GAP04 transitioned between ice-covered and ice-free periods, as well as between cold- and warm-based ice sheet conditions. This approach suggests a paleoclimatically corrected heat flow of  $38 \pm 2 \text{ mW m}^{-2}$ , which is  $\sim 36\%$  greater than the present-day measurement. This paleoclimatically corrected heat flow represents the equilibrium heat flow through Earth's crust, unaffected by long-term variations in ground surface temperature over millennial timescales. When interpreting near-surface heat flow in previously glaciated terrain, it is important to use a paleoclimatically corrected heat flow as the deep (or lower) boundary condition, in heat flow simulations.

Figure 11 shows a 2D cross-sectional heat flow simulation at DH-GAP04 that exemplifies both types of glacial transitions described above. First, at around 10 ka (3 kyr before deglaciation), the local basal thermal state of the ice sheet changes from being cold-based (basal ice temperature  $\sim -8^\circ\text{C}$ ) to warm-based (basal ice temperature at the pressure melting point,  $\sim -1^\circ\text{C}$ ; Hartikainen et al., 2021). This results in a very strong decrease in heat flow. At the DH-GAP04 borehole, the heat flow is reduced from  $35 \pm 3$  to  $12 \pm 1 \text{ mW m}^{-2}$  ( $\sim 65\%$  decrease). Subsequently, at the time of deglaciation (around 7 ka), the ground surface cools by  $2\text{--}6^\circ\text{C}$ , compared to the ice-covered warm-based period, as the area becomes subject to Holocene air temperatures. This cooling lithospheric surface boundary condition increases heat flow to  $25 \pm 5 \text{ mW m}^{-2}$ . At this site, the glacial transitions therefore result in complex changes in heat flow, with values ranging from 11 to  $38 \text{ mW m}^{-2}$  over the past 100 ka (Hartikainen et al., 2021). Across its 25 km transect distance, this simulation also highlights considerable spatial variability in the heat flow response to glacial transitions. Thus, similar to submarine-to-subaerial transitions, glacial transitions may result in considerable temporal variations in geothermal heat flow with a complex kilometer-



**Figure 8.** (a) Difference (measured minus modeled heat flow) at measurement sites in comparison to our “without NGRIP” simulation. Blues denotes that the model is “colder” than the measurement. (b) Comparison of modeled and observed heat flow values at 419 sites. The NGRIP measurement is included in this comparison as a purple dot.

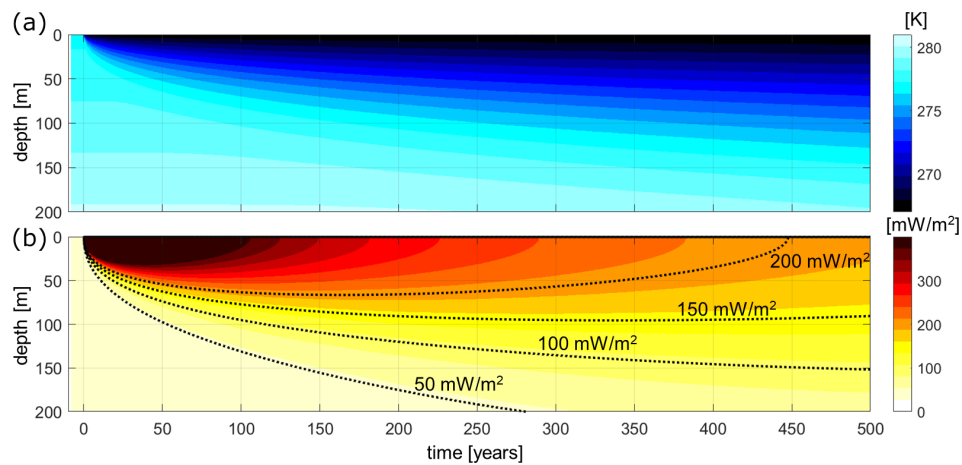


**Figure 9.** The jackknifing ensemble uncertainty range, calculated as the difference between maximum and minimum heat flow, for the (a) “with” and (b) “without” NGRIP simulations. The inclusion of NGRIP introduces substantial ensemble uncertainty in central North Greenland as the machine learning algorithm cannot reconcile the NGRIP anomaly with available input geophysical datasets (Fig. A1).

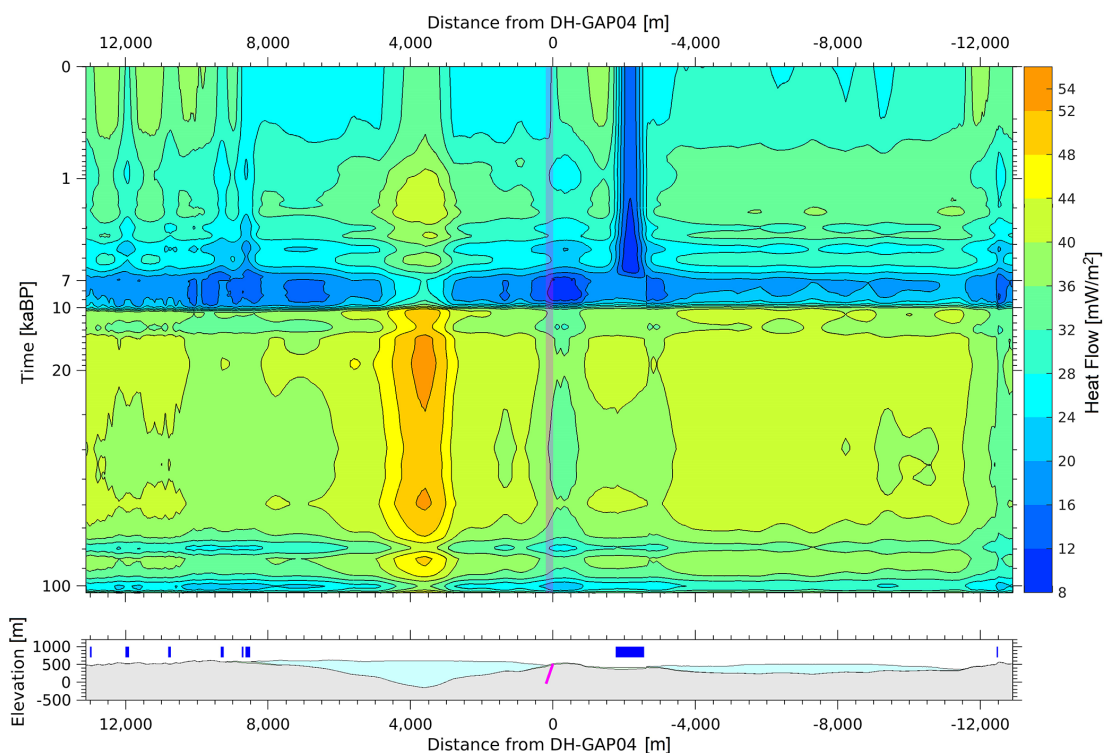
scale spatial pattern. Paleoclimatic corrections for previously glaciated terrain will always have some degree of uncertainty associated with assumed basal ice temperature, near-surface ground temperature, and ice cover histories.

A third paleoclimatic heat flow consideration is the propagation of surface temperature changes through persistent ice sheet cover to the subglacial interface where heat flow is measured (Calov and Hutter, 1997). The propagation of surface temperatures through the ice sheet is controlled by heat conduction and advection, where the latter is caused by a combination of snowfall leading to vertical transport of relatively cold surface snow downwards and ice flow dynam-

ics giving rise to horizontal ice transport from colder interior sites to warmer marginal sites (Hooke, 2019). In the interior of the ice sheet, advection generally dominates in the upper ice column. There, vertical velocity at the ice sheet surface is effectively equivalent to the snowfall rate, which means that downward advection outpaces upward conduction from the bed. Lower in the ice column, where vertical velocities become smaller, conduction becomes increasingly important. This means that – even in the absence of heat sources or sinks – a deep ice temperature profile measured today is typically not representative of present-day climate but is instead a convolution of competing advection and conduction pro-



**Figure 10.** Simulation of ground temperature (a) and geothermal flow (b) in response to a change in boundary temperature from submarine to subaerial settings at year 0. This simulation is parameterized to approximate conditions characteristic of Ilulissat, western Greenland.



**Figure 11.** Top: simulated temporal changes in ground surface heat flow along a 25 km profile crossing the DH-GAP04 borehole (at a distance of 0 m) during the last glacial cycle (104–0 ka). In this simulation, the ice sheet changed from cold-based to warm-based conditions around 10 ka, and the borehole location was deglaciated around 7 ka. Note the logarithmic scale on the y axis. Bottom: surface topography along the 2D cross-sectional model domain. The DH-GAP04 borehole is shown in pink. Present-day outlet glaciers are shown in light blue, and lake locations are shown by dark-blue bars (Hartikainen et al., 2021).

cesses (Calov and Hutter, 1997). Measurements of ice temperatures on the ice divide thus display a cold anomaly in the Last Glacial Period ice temperatures (115–11 ka) and a warm anomaly in the part of the ice column that corresponds in age to the Holocene Climatic Optimum (8–5 ka) (Gundestrup et al., 1994; Dahl-Jensen et al., 2003). Due to con-

duction, the magnitudes of these anomalies moderate over time. Simply put, however, there can be a multi-centennial to multi-millennial time lag, depending on ice thickness, for surface temperature changes to reach the ice–bed interface at the ice divide. In the ice sheet interior, present-day basal ice temperatures still reflect a past cooler climate, and mea-



sured temperature gradient and heat flow will be greater than if Greenland's climate had not changed significantly in the past ca. 100 ka. Indeed, the  $61 \pm 2 \text{ mW m}^{-2}$  present-day heat flow that we estimate at GRIP is  $\sim 20\%$  greater than the  $51 \text{ mW m}^{-2}$  estimated for that site with differing paleoclimatic corrections (Dahl-Jensen et al., 1998; Greve, 2019).

#### 4.2 Other corrections

In addition to correcting local heat flow measurements to account for the effect of non-steady local past climate, there are several other heat flow corrections relevant within the Greenland domain. First, there is local topographic correction to account for elevated heat flow in valleys and correspondingly diminished heat flow along ridge lines (Lees, 1910). This is the only systematic correction currently provided in the database, based on the geostatistical model of Colgan et al. (2021). Within our Greenland domain, this topographic correction ranges from a minimum of  $-21 \pm 5\%$  at Hole 38 in Ilímaussaq, southern Greenland (Sass et al., 1972), to  $+55 \pm 10\%$  at the Dybet site in Young Sound, East Greenland (Rysgaard et al., 2018). These end-member sites highlight the potential importance of acknowledging topographic influence on local heat flow. However, there are several measurement sites where topographic correction is not resolved by the Colgan et al. (2021) product, primarily due to geographic limitations. At Agassiz Ice Cap, for example, four borehole ice temperature profiles have been measured within  $\pm 1100 \text{ m}$  of the same position (Agassiz77, 79A, 79B, and 84; Clarke et al., 1987; Fisher and Koerner, 1994). The ice thicknesses vary between 128 and 341 m across these sites, and assessed heat flows vary between  $51 \pm 6$  and  $58 \pm 5 \text{ mW m}^{-2}$ . Topography likely explains part of the apparent discrepancy among those four closely spaced measurements.

Second, groundwater flow can substantially modify the apparent geothermal heat gradient. The heat advection associated with even relatively small ( $\sim 1 \text{ L s}^{-1}$ ) groundwater flows can be comparable to the heat diffusion associated with the geothermal gradient (Mansure and Reiter, 1979). In Greenland, there are clearly some areas where groundwater flow is substantially suppressing or enhancing the apparent geothermal heat gradient. For example, at the DH-GAP01 site near Kangerlussuaq, in southwestern Greenland, the apparent geothermal gradient in a talik beneath a small lake is almost entirely moderated by groundwater flow. There, measured heat flow is reduced by 90% in comparison to the nearby DH-GAP03 and DH-GAP04 sites (Johansson et al., 2015). In contrast, there are also areas of Greenland with active hot springs where groundwater flow can similarly overwhelm the regional geothermal heat flow value (Persoz et al., 1972). At Unartukavsaq, West Greenland, for example, a subaerial  $15^\circ\text{C}$  hot spring discharge of  $1 \text{ L s}^{-1}$  represents a 63 kW sensible heat source relative to a mean annual air temperature of  $0^\circ\text{C}$  (Hjartarson and Armannsson, 2010). This is equivalent to a heat flow of  $6300 \text{ mW m}^{-2}$  over a char-

acteristic area of  $100 \text{ m}^2$ . These end-member sites highlight the need to account for the differing influences of “cold” and “hot” groundwater flow onshore, including in subglacial settings.

Third, while approximately half of Earth's contemporary heat flow is ultimately derived from radioactive decay, primarily within the mantle, heat production from near-surface radioactive sources can influence the apparent magnitude and spatial distribution of the deeper geothermal gradient (Lees, 1910). The temperature profiles of the Ilímaussaq boreholes, drilled for the Kvanefjeld uranium prospect in southern Greenland, are substantially influenced by near-surface radioactive heat production. Their measured heat flows ( $34$  to  $39 \text{ mW m}^{-2}$ ) have been estimated to reflect a near-surface radioactive heat production of  $\sim 8 \mu\text{W m}^{-3}$  over limited horizontal ( $< 10 \text{ km}$ ) and vertical ( $< 1 \text{ km}$ ) extents. Based on heat flow measurements in analogous Precambrian Shield settings in Canada, Australia, and the United States, Sass et al. (1972) suggested that the measured Ilímaussaq heat flows are 26% to 44% higher than the  $\sim 27 \text{ mW m}^{-2}$  heat flow that would otherwise be expected for the Precambrian Shield. The Ilímaussaq boreholes highlight the potentially non-trivial influence of near-surface radioactive heat production for altering local geothermal gradients in Greenland. Given the diverse subaerial geology of Greenland and the vast area of poorly resolved subglacial geology beneath the ice sheet (Dawes, 2009), it is likely that there are areas in the Greenland domain beyond Ilímaussaq and Kvanefjeld where heat flow may be similarly influenced by non-trivial heat production associated with near-surface geology.

Finally, lateral contrasts in the thermal conductivity of rock types can result in local heat flow refraction (Jaeger, 1965; Lachenbruch, 1968). In these settings, geotherms close to the geological boundary are influenced by the contrast between relatively high- and low-conductivity rock types, with heat flow diverted from the low-conductivity rock type into the high-conductivity rock type. Sharp spatial contrasts in thermal conductivity associated with rock type may modify local heat flow by  $\pm 15\%$  in subglacial settings (Willcocks et al., 2021). Within our Greenland domain, the effect of lateral conductivity contrasts on heat flow has been qualitatively described at the Isua site. There, Colbeck and Gow (1979) suggest that the heat flow measured in ice sheet boreholes may be biased low as subglacial heat flow is potentially being diverted into the adjacent, and highly conductive, subaerial iron ore formation. With the recent availability of digital maps of subaerial Greenland geology available on open platforms, such as the Escher and Pulvertaft (1995) geological map that now appears in QGreenland (<https://qgreenland.org>, last access: 8 February 2022), and growing knowledge of variations in thermal conductivity with rock type, it appears feasible to begin systematically quantifying the effect of lateral conductivity contrasts on heat flow across ice-free areas of Greenland.



**Table 4.** Number of heat flow observations and geophysical datasets used by this study and five previous studies within the Greenland domain that we consider. These values may be regarded as best estimates. The interpolation method of each study is also listed.

Model	Number of Greenland heat flow measurements		Number of geophysical datasets	Interpolation method	
	Onshore				Offshore
	Subglacial	Subaerial			
This study	25	77	317	12 Machine learning	
Rezvanbehbahani et al. (2017)	5	4	0	20 Machine learning	
Artemieva (2019)	1	60	229	8 Thermal isostasy model	
Martos et al. (2018)	6	2	0	5 Forward model	
Greve (2019)	5	3	0	3 Paleoclimate and ice flow model	
Lucazeau (2019)	4	62	248	14 Geostatistical model	

4.3 Comparison of heat flow models

We statistically compare our heat flow model to five previously published Greenland heat flow models, whose inputs and methods are summarized and compared with ours in Table 4. With a mean Greenland heat flow of  $44 \text{ mW m}^{-2}$ , our “without NGRIP” simulation infers the lowest mean Greenland heat flow across all models to date (Table 5). Even when including NGRIP in the continental jackknifing simulations, our mean “with NGRIP” heat flow only increases to  $48 \text{ mW m}^{-2}$ . When comparing all seven heat flow simulations with the  $n = 21$  grid cells of binned heat flow measurements (including NGRIP) within the Greenland land area common to all seven models, we find that our model, which includes NGRIP in the training dataset, has the smallest bias and root mean square error (RMSE). Our model is followed by the model of Rezvanbehbahani et al. (2017), which has the second-smallest bias, and the model of Greve (2019), which has the second-lowest RMSE (Fig. 12). If we exclude the NGRIP measurement from the comparison our simulation trained without NGRIP yields the smallest bias ( $0 \text{ mW m}^{-2}$ ) and RMSE ( $11 \text{ mW m}^{-2}$ ). As we exclude the relatively high and uncertain NGRIP heat flow measurement from our training dataset, our model clearly does not reproduce this heat flow anomaly, unlike models that do include the NGRIP measurement. However, even our model that is trained with NGRIP predicts a heat flow value of only  $90 \text{ mW m}^{-2}$  at NGRIP, which is  $39 \text{ mW m}^{-2}$  less than the measured value ( $129 \text{ mW m}^{-2}$ ). This difference of  $39 \text{ mW m}^{-2}$  between measured and predicted values remains the highest residual among all onshore sites (Fig. 12).

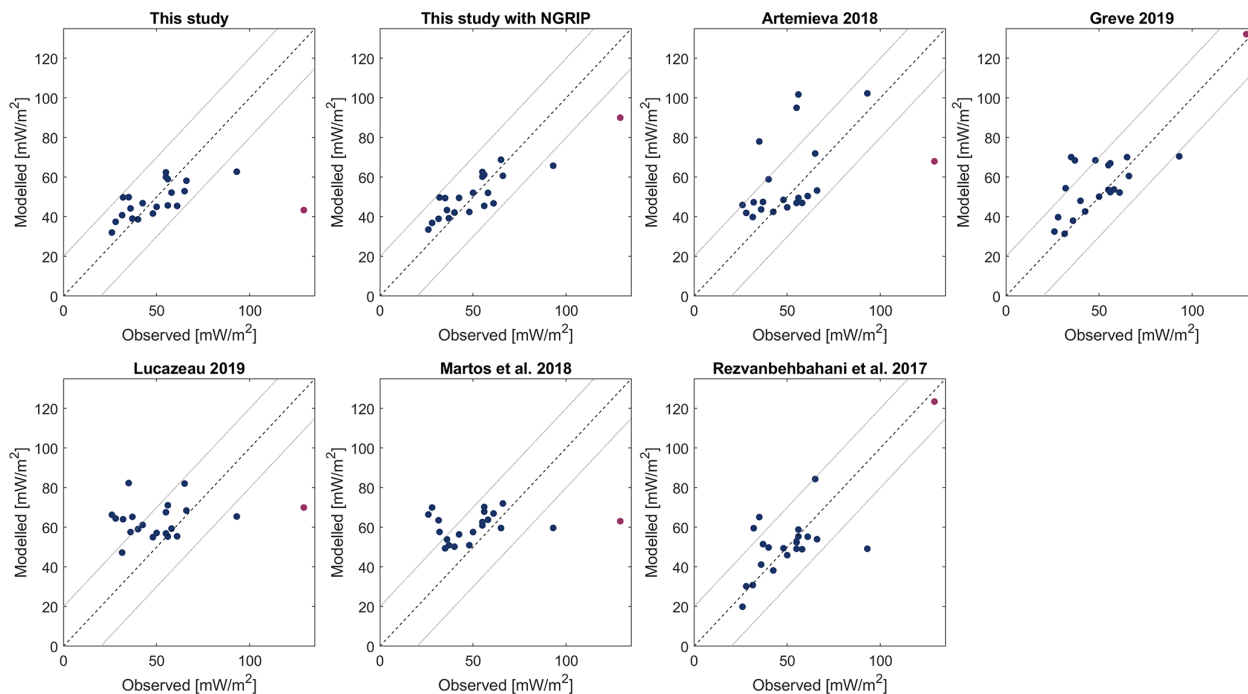
We also qualitatively compare the magnitude and spatial distribution of our geothermal heat flow map without NGRIP with these five previous studies (Fig. 13). In comparison to the most methodologically similar model, the machine learning model of Rezvanbehbahani et al. (2017), we infer a significantly cooler central North Greenland. This pronounced difference can be explained partly by our choice of excluding the NGRIP measurement, but also the model which is trained

with NGRIP is colder in central North Greenland than Rezvanbehbahani et al. (2017). We also employed more Greenland regional datasets, which are generally finer-resolution and better suited to our study area than global datasets, and trained our model with a significantly larger number of in situ measurements.

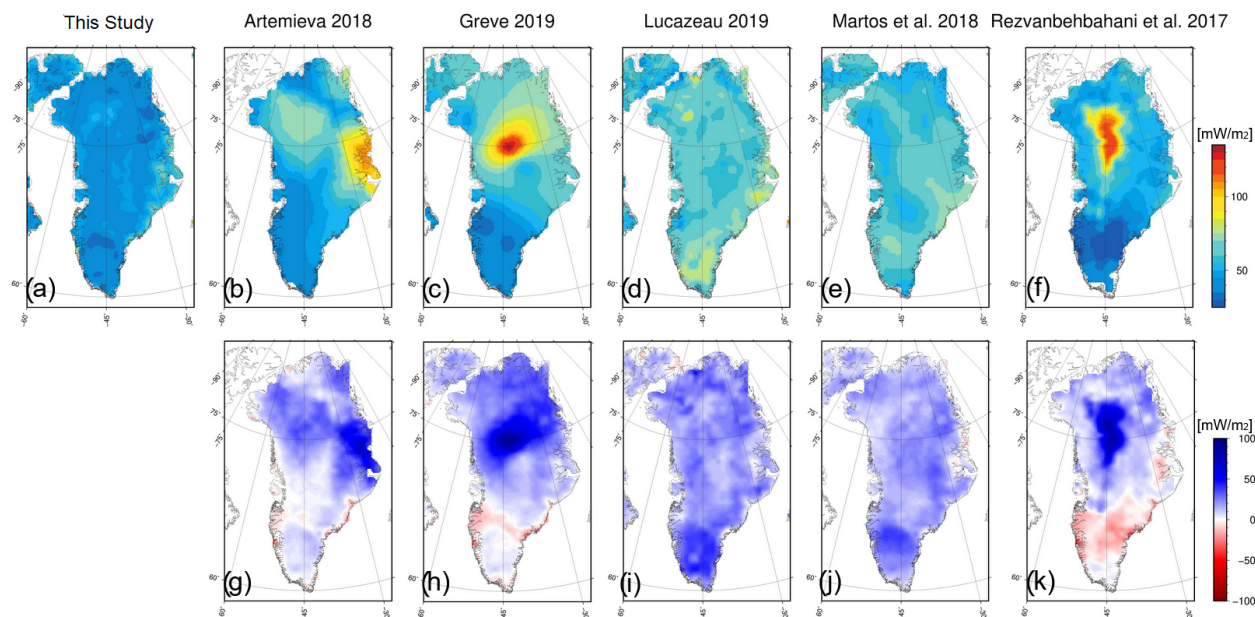
The Lucazeau (2019) heat flow model is also geostatistical as it relies on empirical correlations between global geophysical datasets and heat flow measurements. While the Lucazeau (2019) model shares similar length scales of spatial variability ( $\sim 50 \text{ km}$ ), its patterns are quite different from our model. For example, they infer a relatively warm North Atlantic Craton. These differences can be attributed to our explicit preference for more detailed regional datasets instead of global datasets where possible. An additional source of deviation from the Lucazeau (2019) model may be their manual weighting of included geophysical datasets, whereas our machine learning algorithm includes geophysical datasets without a priori weights.

Martos et al. (2018) based their heat flow model on identifying deep magnetic sources, assuming that the deepest sources coincide with the Curie isotherm and constructing a regional thermal model of heat flow. Their regional thermal model infers a band of enhanced heat flow anomaly from central East Greenland to northwestern Greenland. Our model does not infer this enhanced heat flow anomaly and also finds substantially lower heat flows in the North Atlantic Craton. We attribute these differences to the choice of input data, including availability of heat flow measurements and differing geostatistical and thermal modeling approaches.

Artemieva (2019) proposed a thermal isostasy method to calculate upper-mantle temperatures, lithosphere thickness, and geothermal heat flow using bedrock topography, ice thickness, and Moho depth from seismic data. This approach is sensitive to the choice of surface-wave-tomography models and reference values. Our model has the largest discrepancy with the Artemieva (2019) model along the coast of central East Greenland, where our model infers a substantially lower heat flow over a broad region. All four other models,



**Figure 12.** Comparison of simulated and measured geothermal heat flow in binned 55 km grid cells within the Greenland land area common to all seven models ( $n = 21$ ; Table 4). RMSE and bias for each model listed in Table 5. The NGRIP measurement is included in this comparison as a purple dot.



**Figure 13.** Comparison of the heat flow model of this study without NGRIP (a) to previously published models: (b) thermal isostasy from Artemieva (2019) model 1, (c) fitting to available heat flow data from Greve (2019), (d) global similarity study from Lucazeau (2019), (e) magnetic data from Martos et al. (2018), and (f) machine learning by Rezvanbehbahani et al. (2017) with NGRIP =  $135 \text{ mW m}^{-2}$ . Anomaly plots are shown in (g) to (k), where blue denotes this study as “colder”. Mean geothermal heat flow of each model summarized in Table 6.

however, also predict lower heat flows than Artemieva (2019) in this region, which suggests that this anomaly is unique to the data or methodology of Artemieva (2019).

Finally, the Greve (2019) heat flow model uses a numerical ice sheet model to infer the heat flow required to match observed basal ice temperatures at ice core locations, treating the Pollack et al. (1993) heat flow model as an a priori guess. Similar to the difference field with Rezvanbehbahani et al. (2017), our difference field with Greve (2019) highlights the NGRIP measurement site, where we predict a much lower regional heat flow. The spatial patterns in the Greve (2019) model suggest it is sensitive to a relatively small number of input heat flow measurements ( $n = 8$ ). Part of the difference from our model may therefore be attributable to differing availability of heat flow measurements between studies.

#### 4.4 Icelandic plume track

There is substantial interest in understanding the paleo-Icelandic plume track beneath Greenland and its ice sheet. Generally, there is consensus among plume track models that the Icelandic plume was located in the vicinity of present-day Kangersertuaq Fjord in central East Greenland at  $\sim 50$  Ma (Morgen, 1983; Forsyth et al., 1986; Müller et al., 1993; Doubrovine et al., 2012; Martos et al., 2018). In addition to reconstructing past tectonic motion, this paleoplume fix is supported by the presence of basalts up to 7 km thick from the North Atlantic Igneous Province dated to  $\sim 56$  Ma along the Blossville Coast (Storey et al., 2007) and the presence of a relatively low upper-mantle viscosity there today (Khan et al., 2016). However, there is less consensus as to the plume position prior to this time, with analogous west coast plume positions previously hypothesized from Disko Island ( $70^\circ$  N) to Petermann Glacier ( $81^\circ$  N). It has also been recently suggested that Greenland bears the imprint of not just one, but two plume tracks with differing histories of activity (Toyokuni et al., 2020).

Our model does not show an area of elevated heat flow that might be interpreted as a remnant of the Icelandic plume track. This pattern is consistent with Rezvanbehbahani et al. (2017), Greve (2019), and Lucazeau (2019) but differs from Martos et al. (2018) and Artemieva (2019), which both infer plume tracks. An earlier model from Rogozhina et al. (2016) can also be interpreted in terms of west-to-east passage of a plume. There is debate over the potential influence of paleoplume activity on present-day heat flow beneath the Greenland Ice Sheet (Smith-Johnsen et al., 2020; Bons et al., 2021). It is plausible that enhanced heat flow of  $\sim 10 \text{ mW m}^{-2}$  may exist along the most recent 50 Ma of the plume track, but heat flow anomalies likely become difficult to observe after this time as heat flow returns to balance (Martos et al., 2018). The main observable impacts of paleoplume activity are now likely limited to the base of the lithosphere, meaning that plume tracks derived from geophysical

modeling may not reflect near-surface heat flow but rather an imprint of underlying lithospheric architecture.

Interpretations of paleoplume tracks are further complicated by the presence of very young volcanic and geothermal processes. Young igneous formations in East Greenland, dated to 14 Ma, suggest very recent widespread volcanism well after Greenland moved off the Icelandic plume (Storey et al., 2004). Similarly, the presence of coastal geothermal hot springs in both central West and central East Greenland suggests appreciable contemporary heat sources far from the contemporary location of the Icelandic plume (Hjartarson and Armannsson, 2010; Fig. 6). Storey et al. (2004) suggested that metasomatic processes and emplacement of volatiles from the plume into the shallow mantle may have had long-lived influence on the region. It is conceivable that small pressure perturbations resulting from loading and unloading of ice sheets and/or tectonic stresses could result in local volcanism in such a mantle (Jull and McKenzie, 1996). A north–south transect of onshore heat flow measurements in central East Greenland, either subaerial, subglacial, or a combination of both, could improve understanding of the residual heat flow anomaly associated with the paleoplume as well as offer insight on elevated heat flow potential associated with young secondary processes.

#### 5 Data availability

Digital assets for the Greenland Geothermal Heat Flow Database and Map (Version 1) are now available open-access on the GEUS Dataverse at <https://doi.org/10.22008/FK2/F9P03L> (Colgan and Wansing, 2021). These include (1) heat flow measurement database as .xlsx and .shp, (2) heat flow maps at 10 and 55 km resolutions as .xyz and .nc for the model with and without NGRIP, (3) hot spring database as .xlsx, (4) regional coast+500 km domain boundary as .shp, and (5) extrapolation code for subglacial temperature gradients as .m. The primary temperature and stratigraphy profiles of the Hellefisk-1, Ikermiut-1, Kangâmiut-1, Nukik-1/2, Alpha-1, Atammik2-1 (AT2-1), Atammik7-1 (AT7-1), Delta-1, Gamma-1, LadyFranklin7-1 (LF7-1), Qulleq-1, T4-1, and T8-1 submarine boreholes are owned by the Geological Survey of Denmark and Greenland. The derived temperature gradients and heat flows included here are published with special permission. The primary temperature and stratigraphy profiles of the Alcoa\_Site7e-I, Alcoa\_Site7e-P, and Alcoa\_Site6g-P subaerial boreholes are jointly owned by the Greenland Government and Alcoa. The derived temperature gradients and heat flows included here are published with special permission.

#### 6 Summary remarks

Here, we document the first version of the Greenland Geothermal Heat Flow Database and Map. While we have

**Table 5.** Minimum, mean, and maximum heat flow over Greenland onshore area from this study and five previous studies. Root mean square error (RMSE) and simulated minus measured bias relative to heat flow measurements within the common land area domain also shown (Fig. 12). Here, the number of onshore heat flow measurements (sample) reflects binned 55 km grid cells. See Table 4 for a comparison of the heat flow measurements used in each of these studies.

Model	Min [mW m <sup>-2</sup> ]	Mean [mW m <sup>-2</sup> ]	Max [mW m <sup>-2</sup> ]	RMSE [mW m <sup>-2</sup> ]	Onshore bias [mW m <sup>-2</sup> ]	Sample size
This study (without NGRIP)	28	44	76	11	0	20
				22	-4	21
This study (with NGRIP)	29	48	102	13	0	21
Rezvanbehbahani et al. (2017)	20	54	124	15	+1	21
Artemieva (2019)	40	58	108	23	+6	21
Martos et al. (2018)	50	60	75	24	+8	21
Greve (2019)	32	62	130	14	+6	21
Lucazeau (2019)	46	64	83	25	+11	21

increased the available heat flow measurements by 44 % (290 to 419) within the Greenland domain, onshore measurements remain disproportionately scarce. Greenland represents  $\sim 1.5$  % of global land area, but – once our regional database is merged into the IHFC database – Greenland will only represent 0.08 % of IHFC onshore measurements. We anticipate updating this measurement database as new measurements and corrections become available. We will continue efforts to parse the Type 2 sites (uncertain temperature data) into either Type 1 (new measurement) or Type 3 (no temperature data) classes. Topographic correction is presently the only systematic correction available within the measurement database, but there is clearly a need for more systematic corrections. Most notably, the application of a standardized, yet site-specific paleoclimatic correction to all measurement sites would ensure that heat flow measurements are interpreted in the correct climatic context.

We have used machine learning to create a heat flow map that is self-consistent with this measurement database. While there has been a proliferation of Greenland heat flow maps in the past decade, compilations of in situ heat flow measurements have generally lagged these models. The heat flow map we present may be the first of a new generation of heat flow maps that are self-consistent with a larger and better documented inventory of in situ heat flow measurements. Clearly, a fundamental question in mapping Greenland heat flow is whether or not to include NGRIP as a regionally representative heat flow measurement. This touches on a larger issue of how to not only understand the effect of local processes in measurements but also how to represent the influence of these local processes in large-scale models. For the NGRIP measurement, this may mean resolving potential differences between “basal” and “geothermal” heat flows. In regional models, this may mean representing the cumulative effect of local processes like widespread hot spring activity.

There is a present possibility of using unconventional sources to increase the inventory of available in situ Greenland heat flow measurements. Abandoned prospecting wells

can be repurposed for heat flow measurements by inserting temperature strings to measure local geothermal gradients. While this requires knowledge of well closure and present conditions, it has been successful in the past. Local temperature gradients can also be resolved by combining knowledge of seismic- or radar-derived permafrost depth with mean annual ground temperature. This is most feasible in valley bottoms, where the permafrost is contained within a thick sediment package, and there is free water below the permafrost limit. Finally, while hot spring temperatures do not serve as reliable indicators of regional geothermal heat flow, chemical analyses of their water can resolve deep temperatures that are characteristic of regional phenomena and heat flow. It is likely that many uncharted hot springs – both subaerial and subglacial – exist in Greenland.

There are presently at least four grand challenges that motivate further improvement of our scientific understanding of Greenland’s geothermal heat flow. First, geothermal heat flow is a critical basal boundary condition for the thermo-mechanical ice flow models being used to project Greenland’s future sea level rise contribution. Second, geothermal heat flow is similarly a critical basal boundary condition influencing the stability of periglacial processes and landforms, including both onshore permafrost and shallow offshore gas hydrates, that can deteriorate under future climate change. Third, resolving how the present-day magnitude and spatial distribution of geothermal heat flow may reflect the pre-50 Ma influence of the Icelandic plume track may be key to resolving Greenland’s geologic history. Finally, from a social perspective, there appears to be potential for low-temperature geothermal heating to play a role in Greenland’s green-energy transition.



## Appendix A

**Table A1.** Comparison of the data fields in the IHFC 1976 and 2021 database structures (Jessop et al., 1976; Fuchs et al., 2021a) and this study. “X” denotes an included field. Red outlines and bold text denote the 18 mandatory data fields of the IHFC 2021 structure, and red shading denotes the 8 mandatory data fields that are not available in our new database. Here, “T” denotes temperature, and “TC” denotes thermal conductivity. Full IHFC naming convention described in Fuchs et al. (2021a).

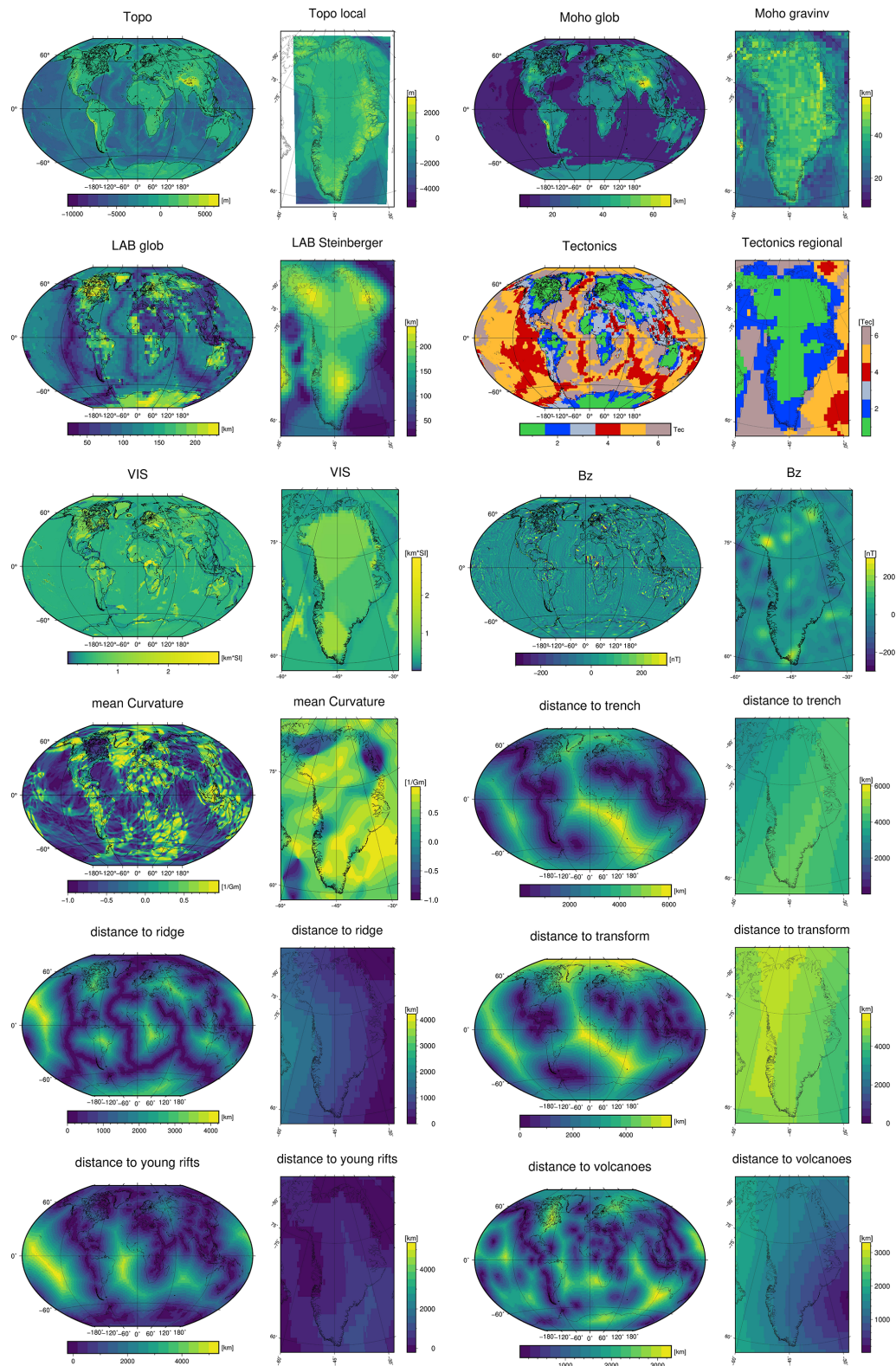
	IHFC 1976	IHFC 2021	This Study
<b>Heat-flow</b>	X	X	X
Heat-flow uncertainty		X	X
<b>Site name</b>	X	X	X
<b>Geographical latitude</b>	X	X	X
<b>Geographical longitude</b>	X	X	X
Geographical position uncertainty			X
Geographical elevation	X	X	X
Geographical elevation uncertainty			X
EPSG:3413 Easting			X
EPSG:3413 Northing			X
EPSG:3413 Uncertainty			X
<b>Geographical environment</b>		X	
Measurement type			X
Type of exploration method		X	
Original exploration purpose		X	
Flag heat production		X	
Bottom-water temperature		X	
General comments	X	X	X
<b>Heat-flow method</b>		X	
<b>Heat-flow interval top</b>	X	X	X
<b>Heat-flow interval bottom</b>	X	X	X
Penetration depth		X	
Probe type		X	
Probe length		X	
Heat-flow transfer mechanism		X	
Additional references		X	
Date of acquisition		X	
<b>Relevant child</b>		X	
Flag in-situ properties		X	
Flag temp correction		X	
Flag sedimentation effect		X	
Flag erosion effect		X	
Flag topographic effect		X	X
Topographic correction			X
Topographic correction uncertainty			X
Topographic corrected heat flow			X
Topographic corr. heat flow uncertainty			X
Flag transient climate effect		X	
Flag convection effect		X	
Flag bottom-water effect		X	
Flag heat refraction effect		X	
Lithology		X	
Stratigraphic age		X	
<b>Temperature gradient</b>	X	X	X
Gradient uncertainty		X	
Average gradient corrected		X	
Gradient cor. uncertainty		X	
<b>T method (top)</b>		X	
<b>T method (bottom)</b>		X	
Shut-in time (top)		X	
Shut-in time (bottom)		X	
T correction method (top)		X	
T correction method (bottom)		X	
Number T recordings	X	X	
Probe tilt		X	
<b>TC mean</b>	X	X	X
TC uncertainty		X	
<b>TC source</b>		X	
TC method		X	
<b>TC saturation</b>		X	
<b>TC pT conditions</b>		X	
TC pT assumed function		X	
TC number		X	
TC averaging method		X	
Heat Production	X		
Number of heat production measurements	X		
Number of sites	X		
<b>Primary reference</b>	X	X	X
Measurement date		X	X
Publication year	X	X	
Source individual			X
International Geo Sample Number (IGSN)		X	
Identification number	X	X	X
IHFC status		X	X



**Table A2.** Geophysical datasets and features provided to the machine learning algorithm to spatially interpret point measurements of heat flow. Data extent is global unless otherwise stated. A graphical overview of these input datasets is shown in Fig. A1.

Feature	Dataset or method	Reference
Topography		
– global	– Earth 2014	Hirt and Rexer (2015)
– regional	– BedMachine v3	Morlighem et al. (2017)
Moho depth		
– global	– Kriging interpolation	Szwillus et al. (2019)
– regional	– Satellite gravity gradient inversion	following the method of Haas et al. (2020)
Lithosphere–asthenosphere boundary		
– global	– LithoRef2018	Afonso et al. (2019)
– regional	– AMISvArc	Steinberger et al. (2019)
Tectonic regionalization		
– global	– SL2013sv	Schaeffer and Lebedev (2015)
– regional	– NAT2021	Celli et al. (2021) following the method of Schaeffer and Lebedev (2015)
Magnetic susceptibility	VIS model	Hemant and Maus (2005)
Vertical magnetic field	LCS-1	Olsen et al. (2017)
Mean curvature	GOCE gravity gradient	Ebbing et al. (2018)
Distance to		
– trench	– UTIG (plates project)	Goutorbe et al. (2011)
– ridge	– UTIG (plates project)	Goutorbe et al. (2011)
– transform faults	– UTIG (Plates project)	Goutorbe et al. (2011)
– young rifts	– n/a	Şengör and Natal'in (2001)
– volcanoes	– n/a	Siebert et al. (2010)

n/a – not available



**Figure A1.** Graphical overview of the 12 geophysical datasets and features provided to the machine learning algorithm (Table A2). Each variable is shown in both global and local (Greenland) extents.

**Author contributions.** Author contribution is captured following the CRediT system. Conceptualization by WC, AG, KM, RSF, AB. Data curation by KM, KL, FGC, TIN, LCL, AH, SB, SF, JH, JL, DDJ, JON, NB, TF, DP, UG, GD, TN. Formal analysis by WC, AG, ML, KL, JE, FGC, TIN, LCL, DDJ, FM, TF. Funding acquisition by WC, JE, RSF. Investigation by WC, AG, KM, JH, JE, TIN, JAM, NBK, RSF, AB, FM, YM, KKK. Interpretation by WC, AG, KM, ML, JH, JE, FGC, LCL, JAM, AH, SB, NBK, JH, JL, DDJ, AB, JON, FM, YM, KKK, SAK, AL. Methodology by WC, AG, KM, ML, JE, SF, DDJ, KKK. Project administration by WC, AG, KM, RSF. Software: WC, AG, ML, JE. Supervision by WC, JE, SAK. Validation by WC, AG, ML, JE. Visualization by WC, AG, JH. Writing (pre-submission) by WC, AG, KM, ML, JH, JE, LCL, JAM, NBK, SF, JH, JL, JON, YM, KKK, AL. Writing (revisions) by WC, AG, JAM.

**Competing interests.** The contact author has declared that neither they nor their co-authors have any competing interests.

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**Acknowledgements.** The authors thank Ralf Greve (Hokkaido University) and the anonymous referee for their helpful reviews of this work as well as Kirsten Elger (German Research Centre for Geosciences) for serving as scientific editor.

**Financial support.** This publication is made possible by the award “HOTROD: Prototype for Rapid Sampling of Ice-Sheet Basal Temperatures” supported by the Experiment Programme of the Vilum Foundation (grant no. 00022885). This work has also been supported by the Programme for Monitoring of the Greenland Ice Sheet (<http://www.promice.dk>, last access: 8 February 2022), funded by the Danish Ministry of Climate, Energy and Utilities. Additional support was given by the European Space Agency through the STSE 3D Earth and the German Research Council (DFG) to the projects “GreenCrust” and “Antarctic Heat”, the latter in the framework of the priority program SPP 1158: Antarctic Research with Comparable Investigations in Arctic Sea Ice Areas. Observational and modeled data for the DH-GAP04 drill hole were financed by the Swedish Nuclear Fuel and Waste Management Company (SKB).

**Review statement.** This paper was edited by Kirsten Elger and reviewed by Ralf Greve and one anonymous referee

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