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# Reflection Seismic Thermometry

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## Abstract

The North Viking Graben (NVG) is part of the mature North Sea Basin petroleum province and designated as a major carbon storage basin for NW Europe. It has been extensively drilled over five decades with an abundance of well and seismic data in the public domain. As such it serves as an excellent setting to demonstrate the efficacy of a reflection seismic data led approach to predicting subsurface temperatures using a state-of-the-art full waveform inversion velocity model covering the entire NVG. In a forward modelling approach, an empirical velocity to thermal conductivity transform is used in conjunction with predefined heatflow to predict subsurface temperature. The predefined heatflow parameters are set based on the range of values from previous studies in the area. Abundant well data with bottom hole temperature (BHT) records provide calibration of results. In the second step of inverse modelling, BHT's as well as the velocity derived thermal conductivity are used to evaluate a 1D steady state approximation of Fourier's Law for heatflow. In this way heatflow is estimated over the 12000 km$^2$ model area at a km scale (lateral) resolution, highlighting lateral variability in comparison to the traditional point-based heatflow datasets. This heatflow is used to condition a final iterative loop of forward modelling to produce a temperature model that is best representative of the subsurface temperature. Calibration against 139 exploration wells indicate that the predicted temperatures are on average only 0.6 °C warmer than the recorded values, with a root mean squared error of 5 °C. BHT for the recently completed Northern Lights carbon capture and sequestration (CCS) well 31/5-7 (Eos) has been modelled to be 97 °C, which is 6 °C below the recorded BHT. This highlights the applicability of this workflow not only towards enhancing petroleum systems modelling work but also for use in the energy transition and for fundamental scientific purposes.

# Highlights

- Estimating subsurface temperature using seismic reflection and velocity data
- Empirical velocity thermal conductivity transform
- Tested in the mature northern North Sea, calibrating results against bottom hole temperatures (BHT) from 139 wells
- Inverse modelling allows derivation of laterally varying heat flow at resolution of tens of km scale
- Blind tested workflow to estimate BHT from Northern Lights CCS exploration borehole 31/5-7 to within 6 °C of reported BHT

**Keywords:**

*Seismic; velocity modelling; subsurface temperature; heatflow*

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## Data Availability Statement

The data that support the findings of this study are available from CGG Services (Norway) AS. Restrictions apply to the availability of these data, which were used under license for this study. Data are available from the author(s) with the permission of CGG Services (Norway) AS.

# 1. Introduction

The importance of understanding the subsurface temperature conditions is manifold. An understanding of the isotherms may help delineate important temperature driven diagenetic boundaries (Bjørlykke et al., 1989). Similarly, it is useful for petroleum exploration as it informs the maturity and types of hydrocarbons to be expected from a given source rock (Allen & Allen, 2013). Unlike frontier areas where the effectiveness of source rocks may be in doubt, this is not a concern in the North Sea. However, understanding of the present-day subsurface temperature can still prove useful in such a mature basin. An example could include identification of the so called “Golden Zone” thermal window in reservoirs (Nadeau, 2011). Similarly, an improved understanding of subsurface temperature could assist in enhanced oil recovery (EOR) such as when $CO_2$ saturation is used to aid recovery of heavy oils by reducing the density of the latter (Davarpanah & Mirshekari, 2020). Investigations into geothermal energy also benefit from improved understanding of subsurface temperature conditions (Bonté et al., 2012; Fuchs & Balling, 2016). The North Sea is one of the world’s most prolific and extensively studied petroleum provinces (Copestake et al., 2003; Cornford, 1998; Leadholm et al., 1985; Rüpke et al., 2008). Over the course of the past sixty years numerous oil and gas fields have been discovered, many of which have served as the testing ground for new technologies such as time lapse 4D seismic or EOR (Awan et al., 2006; Landrø et al., 1999). With this history of developing and applying cutting edge methodologies, it is a fitting setting to test the workflow proposed here. This paper proposes and tests a means of utilising reflection seismic data to predict subsurface temperatures. Previous work has demonstrated that a transform based on empirical velocity and thermal conductivity data may be utilised to convert seismic velocities to thermal conductivities (Sarkar, 2020; Sarkar &

Huuse, 2018). The derived thermal conductivities may be used in conjunction with heatflow data, either from existing open-source data or through modelling of heatflow, to determine subsurface temperatures from Fourier's Law under a steady state condition. There have been numerous studies of the thermal conductivities and heatflow of sediments in the northern North Sea (see Table A.1) (Andrews-Speed et al., 1984; Brigaud et al., 1992; Cornelius, 1975; Evans, 1977; Evans & Coleman, 1974; Houbolt & Wells, 1980; Leadholm et al., 1985). However, there seems to be a hiatus in such studies in recent times. Projects such as the Northern Lights carbon capture and sequestration (CCS) project (Cozier, 2019) would benefit greatly from current knowledge of the subsurface thermal domain for full scale CCS.

## *1.1 Thermal model fundamentals*

When considering heat in the shallow subsurface it is important to understand the sources of heat. That the interior of the Earth is considerably hotter is known, with surface heatflow distribution closely linked to the upper mantle (Pollack et al., 1993). However 40 % of outward heatflow originates in the thin outer crust (Beardsmore & Cull, 2001). Primarily this crustal heat generation is from the radioactive decay of unstable isotopes of elements such as uranium, thorium and potassium (in order of heat production) (Mareschal & Jaupart, 2013). This radioactive heat component is non uniform in the crust due to the variable distribution of radioisotope bearing lithologies within the crust. Oceanic crust for example lacks acidic rocks, which tend to have the highest radioactive heat production, and thus oceanic heatflow has a lower internal heat component (Hasterok et al., 2011). To a lesser extent crustal heating might arise from the frictional heating at faults or from metamorphic processes, which can be either endo- or exothermic in nature (Beardsmore & Cull, 2001).

Fluid movement in the subsurface can lead to heatflow perturbations, particularly in young oceanic crust (Lister, 1972). This might be mitigated by either having sufficient sediment cover or distance from the nearest seamount (from where such hydrothermal fluid systems are likely to originate) (Hasterok et al., 2011). Sediment cover and high rates of sedimentation might also impact heatflow conditions in terms of sediment compaction and thermal rebound with corresponding lower heatflow and suppressed thermal gradient seen in locations of high sediment discharge such as the Bay of Bengal (Hasterok et al., 2011).

To understand how subsurface temperature might be modelled in the present day it is important to establish the key thermal boundary conditions and properties. These are heatflow, seabed temperature, thermal conductivity, the resultant geothermal gradient, and subsurface temperature. The link between heatflow, thermal conductivity and geothermal gradient can be represented by the 1D approximation of Fourier's Law (Eq. 1).

**Equation 1:** $q = k \times \frac{dT}{dZ}$

Where $q$ is heatflow (mW $m^{-2}$); $k$ is thermal conductivity (W $m^{-1}$ $K^{-1}$) and $dT/dZ$ is geothermal gradient (°C $km^{-1}$) where Z is positive downwards. To estimate the temperature at a certain depth in the subsurface, the rate of change of temperature with depth, i.e. the geothermal gradient is important. It becomes apparent then that by rearranging Eq. 1, the input parameters necessary to estimate this are heatflow and thermal conductivity.

**Equation 2:** $\frac{dT}{dZ} = \frac{q}{k}$

Thermal conductivity is a measure of the ease with which heat may be conducted through a material (Popov et al., 2003) and is thus critical to estimating the thermal structure of the subsurface. Thermal conductivity data from direct measurements are made either in situ through well logs or via direct measurements on recovered samples (Andrews-Speed et al., 1984; Jorand et al., 2015; Schön, 2015a). Measuring tools include either a needle probe inserted into a sample or a divided bar apparatus (Evans, 1977; Horai, 1982). However, these measurements may suffer from issues that affect both their ease of acquisition and reliability. For example thermal conductivity probes may have poor contact with borehole walls, or the needle probe reading might be affected by the alignment of the mineral fabric (anisotropy) in relation to the needle orientation (Lucazeau et al., 2004; Pribnow et al., 2000). In the case of the divided bar apparatus, sample preparation is a more involved process, with the application of axial load equivalent to hundreds of metres of overburden on samples having the effect of increasing measurements on both dry and saturated samples (Galson et al., 1987). A range of downhole measuring tools may be used to measure thermal conductivity within the borehole however these are discontinuous measurements that are uneconomic (Mielke et al., 2017). As a result there has been considerable thought put towards deriving thermal conductivity from other more easily measured physical properties such as bulk density, porosity or compressional sound wave velocity (Boulanouar et al., 2013; Esteban et al., 2015; Grevemeyer & Villinger, 2001; Gu et al., 2017; Hartmann et al., 2005; Horai, 1982; Jorand et al., 2015). In the case of velocity, it is found to have similar sensitivity to properties as thermal conductivity (Houbolt & Wells, 1980). That is, thermal conductivity is primarily affected by the mineral composition, porosity and presence of fractures (Mielke et al., 2017; Pimienta et al., 2018; Zamora et al., 1993). Temperature and pressure also impact thermal conductivity though not as much as the other factors (Leadholm et al., 1985; Lee, 2003).

***Figure 1****: Study area overview*

## 2. Geological history

The North Viking Graben (NVG) (Fig. 1) is located in the Northern North Sea between the East Shetland Platform to the West and the Horda Platform to the east, and part of the north western European cratonic block (Brigaud et al., 1992). It is part of the North Sea Graben system (Fig. 2) (Cornford, 1998). It is a Mesozoic rift system, with the rifting in this area having occurred after the Caledonian orogeny (and extensional orogenic collapse), with there being two primary phases of extensional rifting since the Devonian (Fichler et al., 2011; Rüpke et al., 2008; Ziegler, 1992). Primary rifting in the Permian to Early Triassic was followed by a post rift subsidence period (Nøttvedt et al., 1995). The next phase of rifting was from mid Jurassic to early Cretaceous and was also followed by a post rift subsidence period. The rift axis for the Permo-Triassic rifting is believed to be located under the present Horda Platform with the late Jurassic rift axis below the present day Viking Graben (Christiansson et al., 2000). No major tectonic activity is believed to have occurred post Jurassic though there is some conjecture regarding a Tertiary rifting episode (Rüpke et al., 2008) and mid Miocene inversion (Løseth et al., 2013, 2016). The predominant rift direction in the N to NE striking Viking Graben was west-east and northwest-southeast.

The sedimentary record in this area is Devonian and younger. Sand and shales dominate the Triassic to Jurassic basin fill with carbonates and shales predominant in the Cretaceous (Brigaud et al., 1992). Tertiary lithologies consist of shales, silts and sands, with a brief period of Paleocene volcanism marked by the widespread deposition of volcanic tuffs across the basin (Haaland et al., 2000). Crustal basement rocks in this area have a history exceeding one Wilson cycle and trace back to the junction between the Laurentian and Baltican plates, including the opening of the Iapetus Ocean, island arc development linked to oceanic subduction and the Caledonian orogeny (Fossen et al., 2008; Meert & Torsvik, 2003). The composition of the basement rocks can be seen to vary from granites underneath the East Shetland Platform to low and intermediate grade metamorphic and metamorphosed sediments below the Viking Graben and Horda Platform (Fichler et al., 2011).

Source rocks in the area are predominantly Kimmeridge Clay (shales) that were deposited in the Late Jurassic (Davison & Underhill, 2012; Gautier, 2005). Abundant reservoir rocks are available in the NVG all of which exist with a variety of trapping mechanisms. These pre rift reservoirs are usually found in tilted fault blocks where fine-grained post rift sedimentary sequences act as seals.

***Figure 2**: Structural transect*

# 3. Data

The dataset used for the study includes reflection seismic and borehole data.

The North Viking Graben 'mega-merge' was acquired by CGG between 2014 and 2016, covering a total area of 35410 km$^2$. This broadband 3D seismic reflection survey of BroadSeis™ and BroadSource™ configuration covers the northern North Sea basin and was shot in a north-south direction, recorded in TWT down to 9 seconds with an acquisition sample interval of 2 ms (Purvis et al., 2018) though for this work the pre-stack depth migrated (PSDM) volume in the depth domain was utilised. The volume has an inline and crossline bin spacing of 6.25 x 18.75 m. A (flip flop) shot point interval of 18.75 m and a source separation of 37.5 m gives a nominal common-mid-point (CMP) fold of 106. Twelve streamers were used in total, each 7950 m long, with 636 channels towed at depths of 7 – 50 m (BroadSeis™ profile).

Multiple algorithms were used to remove noise and multiples. Both manual picking and time tomographic inversion (TOMOT) was used to generate velocities with imaging done using

Kirchhoff PSTM. A proprietary CGG methodology of continuous automatic bi-spectral velocity picking helped generate the final stacking velocities (Purvis et al., 2018). The processed PSTM and PSDM cubes underpin the modelling work conducted in this study.

The full waveform inversion (FWI) technique aims to produce a high-fidelity subsurface representation of velocity, as the velocity model minimises differences between observed and modelled seismic waveforms within the original raw data (Warner et al., 2013). In making the FWI velocity product, a best guess starting model based on seismic processing velocities is iteratively improved using a sequence of linearized local inversions (Warner et al., 2013). For the CGG NVG survey the fast track velocity product was made from the Dix conversion of root mean squared (RMS) stacking velocities, followed by Kirchhoff depth migration to residual move out (RMO) velocities (CGG, 2019). During the FWI process, the model was subdivided into smaller areas targeting regions of key geology, allowing verification to be conducted (CGG, 2019). Three production runs at ever increasing seismic frequency (4; 5.5 & 8 Hz) were conducted as part of the FWI model build (CGG, 2019). The final 8 Hz update produced the velocity model that best follows geological structure and can characterise small scale features such as injectites (CGG, 2019). The FWI model was calibrated using 101 wells with QC checks completed against sonic log data (CGG, 2019).

Borehole data for this study were primarily sourced from the Norwegian Petroleum Directorate (NPD) website. This gave access to well reports, mud logs, geological reports, and wireline logging data. Because the FWI volume was already calibrated against downhole sonic velocity data and provided in the depth domain, the primary data of interest for this study were the corrected bottom hole temperatures (BHT) recorded for each well within the thermal model area. Where drill stem test temperature readings are available these have been used in the NPD dataset. Publicly available heatflow data from the International

Heatflow Commission (IHFC) database were used to provide constraint on the heatflow parameter (Gosnold & Panda, 2002). There is a scarcity of data points covering the model area as seen in Fig. 1b with the nearest offshore data point too distant to confidently interpolate from (Ritter et al., 2004).

# 4. Methods

The modelling work has been conducted using Schlumberger's Petrel software used for 3D seismic interpretation and to create and manipulate the property volumes and thermal models in this work. Standard seismic interpretation techniques (Cox et al., 2020; Posamentier, 2004), including horizon mapping, surface map creation and seismic attribute extractions were conducted for a structural interpretation of two reference horizons, the seafloor and the Base Cretaceous Unconformity (BCU). The BCU follows the regional stratigraphic framework and serves as a reference horizon, upon which model outputs are overlain as attributes, thus giving a regional context to the results (Evans, 2003). The seafloor is the ceiling for the thermal models, separating the hydrothermal and the geothermal domains. Both horizons were mapped in depth using the 3D reflection seismic data. For the purposes of this work, no other structural interpretation was necessary. Further interpretation or import of grids can of course be done to observe the predicted temperature at desired stratigraphic levels.

The workflow utilised in this project is summarised in Fig. 3. It can be broken down into two main components: the forward modelling approach uses the seismic velocity data as the input parameter to model the thermal conductivity and subsurface temperature using constant surface heatflows, which in turn is calibrated against available BHT data; the inverse approach calculates heatflow from the observed BHTs and the thermal conductivity structure. As heatflow is an important input parameter to model temperature, it becomes possible to update the temperature forward model with the inverse modelling results, and thus validate the final temperature model results, in a manner akin to tomographic update of velocity models (Jones, 2018; Prada et al., 2019).

***Figure 3**: Reflection seismic thermometry workflow*

## *4.1 Forward modelling problem – present day subsurface temperature*

The reflection seismic dataset is a modern broadband seismic survey covering the North Viking Graben (NVG) spanning across parts of the UK and Norwegian continental shelves (UKCS and NCS, respectively). The full waveform inversion (FWI) final velocity model is the key reflection seismic data product underpinning the modelling work.

### *4.1.1 Thermal conductivity structure*

This study uses a high-resolution 3D seismic velocity volume and experimental data relating velocity and conductivity with an empirical relationship (Fig. 4). As this project outlines a remote sensing method, direct thermal conductivity measurement is not possible and instead it must be indirectly determined. If the rocks of the subsurface are considered as a multi component system comprised of minerals, texture (of grains, such as their shape and size), porosity and the fluid content, it becomes possible in an ideal scenario to determine the composite effective thermal conductivity from the contribution of each component part using a suitable mixing law or effective medium model (Duffaut et al., 2018; Hartmann et al., 2005; Schön, 2015b). Normally mud logging data (available here from the NPD database) would be utilised to determine the volumetric fraction of each mineral (Brigaud et al., 1990). We, however, are presenting a remote sensing scenario where only the temperature data from wells is being used to correlate the predicted temperatures with.

The elastic properties of the subsurface are well constrained from seismic data (Mavko et al., 2009) and it provides the avenue to the desired thermal structure. To link thermal properties (thermal conductivity) to elastic properties (acoustic velocity), knowledge of their primary controls becomes necessary. Experimental work has shown that these controls include mineral composition, texture, porosity, the presence of fractures and fluid fill (Gegenhuber &

Schoen, 2012). It helps to think of porosity as the crucial intermediary in the proposed empirical relationship as there have been studies examining the link between acoustic velocity and porosity (Eberhart-Phillips et al., 1989; Lee, 2003; Velde, 1996); and similarly the relationship between porosity and thermal conductivity (Fuchs & Förster, 2013; Jorand et al., 2015). By making the direct leap it must be noted that there are inherent assumptions in such an approach. One concern that may arise is the extent to which the variation in the velocity signal solely corresponding to a thermal conductivity variation (as desired) or is it in fact influenced by external factors (for example fluid overpressure). In such an instance, this issue may be obviated by restricting application to regions of hydrostatic fluid pressure only, however it might be the case that slight overpressure would only correspond to a minor increase in velocity (Lee, 2003). While porosity is a crucial intermediary, mineral composition has a dominant impact on thermal conductivity. Argillaceous rocks have lower conductivities than non-argillaceous (quartz rich) rocks for example (Mavko et al., 2009). The more monomineralic a rock, the greater the correlation between velocity and thermal conductivity (Esteban et al., 2015), which alludes to the impact of anisotropy of thermal conductivity. When considering clay rich rocks, they display poorer correlations for thermal conductivity arising from the platy nature of clay grains (versus more rounded sand grains). At greater compaction however, parallel alignment of clay grains could contribute towards lower effective porosities and thus a stronger correlation with both velocity and thermal conductivity (Velde, 1996). With use of this proposed empirical transform expected for both shallow and deep intervals, it is hoped that the variability of clay rich sediment at shallow levels is balanced at deeper levels.

It must be noted that even the best thermal conductivity models are subject to caveats, either in the form of the inherent assumptions or the specific circumstances where direct relationships might not be as strong.

By including a wide range of studies, covering a wide array of lithologies and settings, it is hoped that the resulting empirical relationship can serve as a robust first order estimate for the varying porosities encountered within any projected study area. Lithologies include clastic sediments (sandstones of varying grain size from fine to coarse; varying mineral content including nearly isotropic clean quartz rich sandstones & clay rich sandstones); carbonates (limestone; dolomite; marl; etc); mudstones (clay- & siltstones); and volcanic & granitic rocks (granite; basalt; gabbro; etc). The sample dataset is limited to wet samples only and measurements taken using transient measurement apparatus such as the optical scanning method (Popov et al., 1999), in order to maintain applicability to fluid filled rocks in the subsurface and parity between data points respectively. Though the study area is dominated by shales, the approach adopted with the empirical relationship is intended to be globally applicable and thus incorporates samples not proportionally dominant in the NVG study. This approach has previously been applied in passive margin settings offshore Namibia and offshore USA (Sarkar & Huuse, 2018, 2021). Future studies could aim to utilise an edited sample list reflective of the dominant lithologies in the target area. The best fit regression (Fig. 4) through the subset of points is as follows:

**Equation 3:** $k_V = (6 \times 10^{-5}) {V_p}^{1.3279}$

Where $k_V$ is thermal conductivity from velocity (W $m^{-1}$ $K^{-1}$) and $V_P$ is P wave velocity (m $s^{-1}$)

***Figure 4**: Empirical transform for velocity and thermal conductivity*

With a highly detailed FWI velocity volume representing the P-wave velocity of the subsurface and a function relating velocity and thermal conductivity, it is possible to convert the FWI velocity volume into a volume of thermal conductivity varying with depth (using Eq. 3). To facilitate the workflow the FWI interval velocity volume was first converted to an average velocity volume in depth below the seabed (as the geotherm starts at seabed) and this in turn allowed the estimation of instantaneous average thermal conductivity as a function of depth in the entire volume.

### *4.1.2 Heatflow input scenarios*

In order to determine the geothermal gradient (Eq. 2), information regarding the heatflow in the study area is required alongside thermal conductivity.

A priori heatflow values were defined based on a combination of existing literature and data from the IHFC. As seen on Fig. 1b, there is a paucity of IHFC data points in the study area. This leads to examining the published record for maps of heatflow covering the North Sea, and these tend to exist in the form of heatflow estimate grids at regional or global scale. Examining these grids such as Davies (2013) or Lucazeau (2019), it becomes apparent that the heatflow varies greatly at basin scale. This seems to depend not just on the size of the grid squares over which the authors have applied their interpolation, but the exact technique used to interpolate and the input parameters they have used will lead to this variability. Consequently it was decided that analysis of the range of heatflow values that are observed in the literature for this area (Andrews-Speed et al., 1984; Cornelius, 1975; Davies, 2013; Evans & Coleman, 1974; Harper, 1971; Leadholm et al., 1985; Lucazeau, 2019; Ritter et al., 2004) will be used to define predetermined starting conditions for heatflow. By using the conclusions of these existing studies, previous modelling accounting for factors such as the impact of the Curie depth in the area or the impact of basal heatflow, is incorporated. This gives a low-, mid- and high- case a priori surface heatflow of 60, 70 and 80 mW $m^{-2}$ respectively.

### *4.1.3 Temperature grids & calibration*

Having determined thermal conductivity and established heatflow scenarios it becomes possible to calculate three geothermal gradient scenarios for the model area. By convolving this with the subsurface depth and incorporating the bottom water temperature (BWT) (i.e., seabed temperature) an estimate of present-day subsurface temperature can be made.

**Equation 4:** $T = T_{SEABED} + (\frac{dT}{dZ} \times Z_{SUBSURFACE})$

where $T$ is predicted temperature (°C); $T_{SEABED}$ is the temperature at seabed (°C); $dT/dZ$ is the instantaneous geothermal gradient (°C $km^{-1}$) (Eq. 2); and $Z_{SUBSURFACE}$ is the subsurface depth (km).

Early studies into the geothermal aspects of the North Sea would set temperature at the seafloor to a constant, for example 10 °C (Cornelius, 1975; Evans & Coleman, 1974; Harper, 1971). Here however seafloor temperature was assigned through decadal averages from the World Ocean Atlas (WOA) dataset (Boyer et al., 2014; Locarnini et al., 2013). Using the nearest measurement node from the 0.25-degree grid of the WOA dataset, the temperature profile (hydrothermal gradient) for this was used to interpolate seafloor temperatures across the seafloor within the model area. Temperatures are found to be in the range of 6 to 9 °C with seafloor temperatures in the shallower water of the UKCS averaging 7.3 °C, and in the relatively deeper waters to the east, these temperatures average 8.7 °C. By basing the seafloor temperatures on a decadal average, the variability of seasonal bottom water conditions and longer scale variability will be negated. This gives a baseline temperature from which the well readings can be seen in context to, particularly with regard to determining the geothermal gradient.

Thermograms, or temperature profiles, depicting temperature change with depth are one way of displaying the model results for each well site (Cornelius, 1975). Petrel allows for the creation of synthetic logs from a reservoir model or pillar grid. This requires resampling the temperature models into a pillar grid coincident with the extent of the area of interest. The dimensions of the individual cells of the grid, and subsequently the total number of cells constituting the entire grid are important with respect to the total compute power. The modelling results displayed here have a lateral resolution of 50*50 m (XY orientation), with the top of the model coinciding with mean sea level (MSL) and the base being set at 5.5 km depth. It is important to establish the base for the thermal model as this will define the interval over which geothermal gradient is calculated. This basal depth encompasses the maximum vertical depth of the wells used for calibration. Cell height was set to 20 m, in accordance with the vertical resolution of the velocity data. For these dimensions, the entire gridded model comprises about 1.4 billion voxels.

With all the input and derived seismic attributes resampled as properties in the pillar grid, synthetic logs for each well are generated. These synthetic logs are made for the three temperature prediction volumes coincident with the three heatflow input scenarios. The estimated temperature at the bottom of each well from the synthetic logs is then plotted against the recorded corrected BHT (Fig. 7). Theoretically, the misfit between observed and modelled temperature for each well can be used to calculate how much the heatflow input for each well must be adjusted for there to be no misfit. In this way, the most representative heatflow value for the area might be calibrated.

## *4.2 Inverse modelling problem – solving for heatflow*

The first stage of subsurface temperature modelling assumed 3 discrete heatflow parameter input scenarios. In this instance, discrete values for heatflow across the model area were used to simulate each temperature scenario. However, heatflow is likely not homogeneous across the model area and there may be lateral variability across individual structures and preferred fluid migration paths. If recorded temperatures in the wells are assumed to be correct and with knowledge of the seafloor temperature, geothermal gradient can be independently computed for each well site. Additionally, if the seismic velocity derived thermal conductivity is considered valid, it becomes possible to use the 1D approximation of Fourier's Law to derive heatflow at each well location. This is the inverse problem.

**Equation 5:** $q = K_V \times \left(\frac{T_{BHT} - T_{SEAFLOOR}}{dZ}\right)$

Where $q$ is heatflow (mW $m^{-2}$); $K_V$ is thermal conductivity from velocity (W $m^{-1}$ $K^{-1}$); $T_{BHT}$ is bottom hole temperature (°C); $T_{SEAFLOOR}$ is temperature at seafloor (°C); and $dZ$ is the vertical depth interval (km).

Interpolating for heatflow between the wells creates a map of inversely modelled laterally varying heatflow. An inverse distance weighting (IDW) function has been used to interpolate heatflow. The values of cells in the vicinity of the processing cell (interpolated gap) are averaged, with the distance of the neighbouring cell having an inverse weighting (Watson & Philip, 1985).

## *4.3 Final temperature model*

Calibrating the results of the temperature forward model with the different heatflow input scenarios, it is likely to show different degrees of agreement. Iteratively updating the heatflow parameter input each time and remodelling for subsurface temperature should theoretically permit the most representative heatflow across the study area to be arrived at ultimately. However, this former approach was not pursued as it was shown previously that inverse modelling (see section 4.2) allows the determination of laterally varying heatflow across the model area. Convolving this heatflow with the seismically derived thermal conductivity (Eq. 2), it should be possible to generate a volume of instantaneous geothermal gradient that has most agreement across all the calibration wells. This thermal gradient is used to generate a final temperature model (Eq. 4) that is resampled into the pillar grid and used to output synthetic temperature logs. Predicted temperature from these logs at hole bottom is extracted and used to calibrate against recorded BHT as before.

The well 31/5-7 is located due southwest of Troll A field and targets the lower Jurassic Cook and Johanssen formations for injection. The targeted interval has ca.173 m of sandstone overlain by 75 m of shale acting as a seal above it. Pressure data indicates no communication in rocks above and below the shale, reinforcing its seal properties. With no results from this well published due to its very recent completion, it was decided to use the available details (well head coordinates and total depth) to simulate the temperature profile for this well using our model.

# 5. Results

The results from the modelling work are presented either as seismic attribute overlays on transects or by means of attribute maps on a gridded horizon. But for the purposes of calibration the primary output is the synthetic temperature log data from the wells. These results are discussed individually in the following section.

The FWI interval velocity volume was converted to an average velocity volume, before applying the velocity to thermal conductivity transform (Fig. 4) to produce a volume of average thermal conductivity. Looking at the internal velocity derived thermal conductivity structure (Fig. 5b), the direct nature of the bulk shift results in the thermal conductivity variation with depth across the volume corresponding to the level of detail seen in the input velocity (Fig. 5a). Cretaceous and younger Cenozoic sediment cover is seen to have relatively low thermal conductivities, with a distinct increase in thermal conductivity seen in the tilted fault blocks of the Viking Graben. The graben corresponds with the greatest amount of overlying sediment cover and lower conductivities relative to the neighbouring rift shoulder.

***Figure 5****: Derived thermal conductivity structure*

The forward modelling results using the predefined heatflow input scenarios are displayed on the west-east transect A-A' (see Fig. 6). As expected, temperatures increase gradually with depth, with no major anomalous zones seen. Cenozoic sediments are coolest with the highest temperatures in the graben itself, which makes sense considering constant heatflow but lower conductivities of the sediments within the graben. Isotherms are more widely spaced as we move towards the Horda Platform in the east where the conductivity is greater. As predicted by Fourier's Law, increasing heatflow leads to increased subsurface temperatures. The direct impact of each heatflow input scenario is better visualised when cross plotting the temperature extracted from each well (Fig. 7).

***Figure 6****: Forward modelling output for heatflow scenarios*

As stated earlier, synthetic thermograms corresponding to the well path of each calibration well were extracted from the predicted temperature models. The temperature at bottom hole in each case has been plotted against the recorded BHT. The best fit regression through each cluster of points and the gradient of this is used as a measure of the degree of agreement between predicted and observed values. The R-squared value, a measure of statistical validity for each regression line, for all three heatflow scenarios is above the 0.7 that is regarded as being the threshold for statistical reliability. Analysing the misfit between predicted BHT and corrected BHT, it is found that on average the low heatflow scenario is 0.64 °C cooler; the mid heatflow scenario has average misfit 15.4 °C higher and the high heatflow scenario has an average misfit of 31.3 °C higher. With increasing heatflow, the corresponding gradient of the regression line through that cluster of points also increases, as does the Y-intercept. The optimum desired gradient that would be expected for best agreement between predicted and observed temperatures would verge on 1, and from this initial forward modelling it is seen that the lowest heatflow scenario is closest to this. However, it is also the case that the optimum regression should be seen to pass through the origin of the cross plot and increasing the heatflow moves each regression further away from this.

***Figure 7****: Forward modelling results calibrated*

When deriving ocean bottom heatflow across the model area through an inverse modelling problem (see 4.2) (Fig. 8), gives mean heatflow of 60.05 mW $m^{-2}$ and a median of 62.82 mW $m^{-2}$. The effective range of heatflow is between 40 to 70 mW $m^{-2}$. Observing the heatflow variation across the area (Fig 8) certain wells are seen to have much lower heatflow than their immediate vicinity. Consequently, during interpolation, these wells are seen in a bulls-eye zone of depressed heatflow. The most prominent of these is observed to the southwest, with two wells (30/5-1 & 30/8-2) having depressed effective heatflow. Referring to Fig. 1 these two wells do not coincide with any field. The other prominent heatflow depression is found to the northeast of the model area at the intersection between Troll-A and Troll-B. The north-westernmost and easternmost wells (29/3-1 & 32/4-1, respectively) (Fig. 8) influence the immediate vicinity by elevating the heatflow interpolated here.

***Figure 8**: Inverse modelling of heatflow and results*

The result of running a final iteration of forward modelling for temperature with the continually laterally varying heatflow derived in Fig. 8 is shown in Fig. 9. We find when cross plotting predicted temperatures from this final model against BHT that the general distribution of points has a far tighter spread than in the previous modelling instances. The average misfit between predicted BHT and measured BHT is 0.58 °C. Furthermore, the best fit regression through this set of points has a gradient nearly equal to 1, and passes nearest to the origin, as is expected for the model best reflecting the actual subsurface temperature. Interestingly, the average of the inverse modelled heatflow as stated earlier is nearly equal to the low case heatflow input scenario used in the early-stage forward modelling. Looking at the distribution of well points (Fig. 9) however it is clear that there is a great disparity in the two approaches. The visual impact of the two approaches is shown in Fig. 10d, e.

***Figure 9**: Well calibration including final temperature model*

The result from the final iteration of the temperature model overlain with key structures is shown in Fig. 10a. Running an RMS amplitude extraction for thermal conductivity on a reference horizon (Fig. 10c), in this case the BCU, it is observed that the highest thermal conductivities are observed in the heart of the Viking Graben, where the BCU is deepest (Fig. 10b). On the flanks of the graben and eastward towards the Horda Platform, thermal conductivities are seen to decrease considerably, consistent with the increasingly shallowing BCU surface at that point. Reverting to the temperature model outputs, we see the difference between the initial forward modelling approach (Fig. 10d) and the final inverse modelled heatflow influenced iteration (Fig. 10e). Comparing the temperatures, within the deep graben for example, it is immediately apparent there is greater variation and detail discernible in from this latter approach. Indeed, what is seen is that some anomalous temperature zones are seen in this final iteration at the northern tip of the Troll field that seem to directly correlate with the anomalous heatflow zone from the interpolation (see Fig. 8a).

***Figure 10**: Temperature modelling results*

The results of simulating the thermal profile for the Northern Lights well 31/5-7 (Eos) are shown in Fig. 11, with temperature at bottom hole of 97 °C falling well within the projected 100 °C range published on the project website (see: https://northernlightsccs.com/en/about accessed at 25/07/2020). Data released by Equinor and the Norwegian government in October 2020 with preliminary results indicated temperature at the bottom of 103 °C. The final well report with its BHT record has not yet been made public but preliminary results from the 31/5-7 prediction show a good agreement between model prediction and reality.

***Figure 11**: CCS well 31/5-7 modelled*

# 6. Discussion

There are multiple corollaries of modelling for subsurface temperature in this manner. The seismic led method of deriving subsurface thermal conductivity structure should enable the verification of zones of thermal blanketing (Cercone & Pollack, 1991; Wangen, 1995). Due to the direct relationship between velocity and thermal conductivity as used here, any zones of anomalously high or low velocity will be reflected in thermal conductivity anomalies. The general trend contains lithologies such as sandstone, shale, limestone and crystalline rocks. This excludes lithologies that deviate from our general trend such as halite, which will need to be considered explicitly to represent the temperature above, within and below such anomalous bodies.

The benefit of the highly detailed FWI velocity model available in this area is that the velocity data has been calibrated against wells in the NVG, thus ensuring the velocity model is a good representation of true subsurface properties and conditions. The conversion to thermal conductivity and temperature provides another means of visualising the subsurface. It is of critical importance to both petroleum exploration and carbon sequestration in the area. In other locations such estimates would be highly beneficial to geothermal or gas storage operations. It must be noted that any artefacts in the velocity data will be translated to the derived thermal conductivity, instantaneous geothermal gradient and predicted temperature volumes as a consequence of the direct transitions in the workflow.

## *6.1 Heatflow modelling – impact and implications*

The validation of the inverse modelling of heatflow as an input as opposed to the use of a discrete integer value heatflow input is borne out by the results. The interpolation of heatflow over the 12000 $km^2$ model area (Fig. 8) highlights the variation in heatflow magnitude laterally at improved coverage compared to most existing studies. As shown earlier (Fig. 1), global and regional compilations usually produce heatflow grids that are at the scale of 1-degree grid squares or larger. The lateral resolution is thus many orders of magnitude poorer than when based on BHT and conductivity data. Interpolating heatflow at such fine scale might enable the examination of any trends, if any, in heatflow versus features that may be hydrocarbon fields, or structural trends (for example major fault networks). While there is a paucity of data points in the IHFC grid coinciding with large parts of the Northern and Central North Sea, the many decades of hydrocarbon exploration in the area led to numerous wells, many of which have detailed records of BHT. When combined with the seismic velocity driven model to ascertain thermal conductivity structure, it should allow for the possibility to fill in the large gaps in the global point heatflow database, which in turn should allow for more detailed basin scale heatflow studies.

The thermal anomaly visible in the interpolated ocean bottom surface heatflow map (Fig. 8) can be seen to coincide with the Troll Field (Fig. 1). Records of this thermal anomaly attribute it to transient effects of uplift caused by late Quaternary deglaciation (Cornford, 1998). The method used to interpolate heatflow might have an impact. IDW gives the best results when the sampling is sufficiently dense with respect to the local variation being simulated (Watson & Philip, 1985). Where sampling is sparse or uneven, the interpolated result will insufficiently represent the desired surface (Philip & Watson, 1982). From the density of wells around existing fields in the area, we can be sure that the IDW interpolation will reliably capture the laterally varying heatflow at this local scale. Nonetheless at the edges of the model area there will be some degree of uncertainty associated with the interpolated heatflow, a consequence of the sampling sparseness in these regions. Kriging would be a more statistically rigorous method of interpolating heatflow, but it requires a prior investigation of the spatial behaviour of heatflow in the sample points. This is incumbent on a pre-existing understanding of the factors influencing the modelled parameter. In a blind test the latter would not necessarily be possible, and thus IDW should be satisfactory for a first order interpretation.

While heatflow has been derived from inverse modelling, in sedimentary basins there is a thermal contribution linked to the radiogenic heat production of sediments and crustal material, and the contribution of heat from the earth's deep interior (that is mantle) (Allen & Allen, 2013; Hasterok et al., 2011; Hasterok, 2010; Hokstad et al., 2017). The modelling here estimated present-day subsurface temperatures using a steady state approximation (Eq. 1). As such, a conscious decision was made to introduce as few variables as possible. Radiogenic heat production would be one such variable. Its impact here has been noted but not explicitly modelled. Mantle heat production is usually estimated from the Moho, whose depth in the area averages roughly 30 km (Grad & Tiira, 2009). Estimates of the Curie isotherm at its shallowest in the study area place it at similar average depths of ca.30 km (Fichler et al., 2011; Kubala et al., 2003). Under the graben, Moho depth does become shallow, up to 22 km in places (Licciardi et al., 2020). Referring to Fig. 8a, modelled heatflow is higher towards the graben centre, which suggests some correlation. With the thermal model base set at 5.5 km, and the degree of shallowing of the Moho not exceeding ~22 km at its shallowest point, the impact of mantle heat input from a modelling perspective is considered negligible in this instance. The general Moho trend in the northern North Sea has a gradual shallowing in the north-west, near the Shetland Islands (Licciardi et al., 2020). If conducting thermal modelling over the whole NVG survey area, and with the basal limits for the model set sufficiently deep such that there might be a basal mantle heatflow effect due to proximity to the Moho for the deepest section of the model. Future models seeking to integrate surface heatflow with a basal heatflow flux can rely on the Curie depth for constraint (Blakely, 1988). The Curie depth is often seen to coincide with a compositional boundary reflecting the loss of magnetic minerals in rocks (Rajaram et al., 2009) and thus can be treated as an isotherm, identifiable using magnetic geophysical methods. It should become possible then to contrast modelled surface heatflow against calculated surface heatflow from past studies, allowing determination of whether the sensitivity of NVG thermal anomalies is to basal flux or thermal conductivity effects, or a combination of the two factors.

## *6.2 Relevance*

The applications of this proposed workflow range from immediate usage by the hydrocarbon industry to supporting new subsurface uses aligned with the energy transition. Hydrocarbon explorationists may use the knowledge of the isotherms to help develop their petroleum systems models or it can help production teams better ascertain the distribution of temperature in the reservoir in order to inform reservoir engineering projects to maximise recovery. Knowledge of subsurface thermal structure is important for the nuclear waste disposal industry due to the sensitivity of the waste to thermal perturbations (Brigaud et al., 1992). From a low carbon technology solutions perspective, mapping subsurface isotherms may enable geothermal energy prospecting and the understanding of subsurface temperature will be important for CCS operations in both frontier and mature basins.

Simulating temperatures for the current CCS Northern Lights well 31/5-7 emphasises the real-world applicability of this model, both in terms of its speed of producing an estimate and its relevance in the energy transition for the future. It is important to understand the temperature conditions in $CO_2$ storage reservoirs as the properties of the gas vary with temperature and pressure. Of these only the pressure is routinely estimated based on seismic data (Eiken et al., 2011). At higher temperatures the density of $CO_2$ decreases, theoretically allowing for a greater volume of it to be stored in a reservoir. Studies of $CO_2$ injection into the Utsira formation (part of the Sleipner Project) have shown that reservoir temperature is a source of uncertainty as it can also impact the diffusivity of the gas within the reservoir (Chadwick et al., 2006). By using the subsurface thermal model proposed here, this key uncertainty may be constrained by project planners, both giving more constraint on the volume of $CO_2$ able to be sequestered within a reservoir, but also the ability to ascertain lateral temperature variability would enable more nuanced storage across different parts of a reservoir. Finally understanding the temperature conditions and where the potential of $CO_2$ diffusivity is highest could help mitigate the possibility of $CO_2$ leakage.

It is not just industrial applications for which this methodology may be utilised. Academic end use cases are also envisioned. The study of the microorganisms endemic to the deep subsurface is nascent and opens up the possibility of the crust playing host to potentially great biodiversity and biomass (Basso et al., 2005). Limited studies into the microbial organisms found in oil reservoirs have yielded surprising results. One such study in the Troll field, examining the microbial diversity of produced water, indicated that these microbes were not introduced as contaminants into the reservoir as a by-product of drilling; instead RNA analyses and gene matching has indicated that these are a distinct genera of temperature sensitive microbes that do not match existing known mesophiles or thermophiles (Dahle et al., 2008). Due to the temperature dependence of these novel microbes, and the difficulty with sampling, an understanding of subsurface conditions might help in providing some inclination of the exact genera that can be encountered in a reservoir based on the predicted temperatures from the model. Thus, it is envisioned that the proposed model can assist the microbiological community as well. Bacterial remediation has been studied as a means of clean up for chemical or hydrocarbon contaminated reservoirs or aquifers (Hazen, 1997). Understanding of the temperature field in the subsurface can help determine how conducive the conditions are to the proliferation of such organisms. From a resource perspective microorganisms have been found to impact natural gas, carbon sequestration, hydrocarbons or even interfere with the underground storage of nuclear waste (Christofi & Philip, 1997).

# 7. Conclusion

The work outlines a novel methodology that utilises state of the art velocity model data from a mature basin such as the North Viking Graben to determine present day subsurface temperatures non-invasively. Forward modelling simulations underpinned by the velocity data and utilising an empirical thermal conductivity transform have been calibrated against recorded temperature data from oil field wells in this sector of the North Sea. Existing work using well data allows the computing of the vertical component of heatflow, in the same orientation as the well. Through inverse modelling here it has been demonstrated that heatflow can be computed in such a manner that lateral heatflow variability coverage is improved compared to existing datasets. Using this derived heatflow to iteratively update the forward model produced a temperature model, the calibration results for which indicate the validity of this approach. To prove the real-world efficacy of this work it has been applied to a recently drilled carbon capture and sequestration well, estimating the temperature in the target reservoir to be within a 5 °C margin at ~3 km subsurface depth, highlighting the usability and robustness of this methodology in hydrocarbon exploration and future energy transition projects.

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# 9. Figures

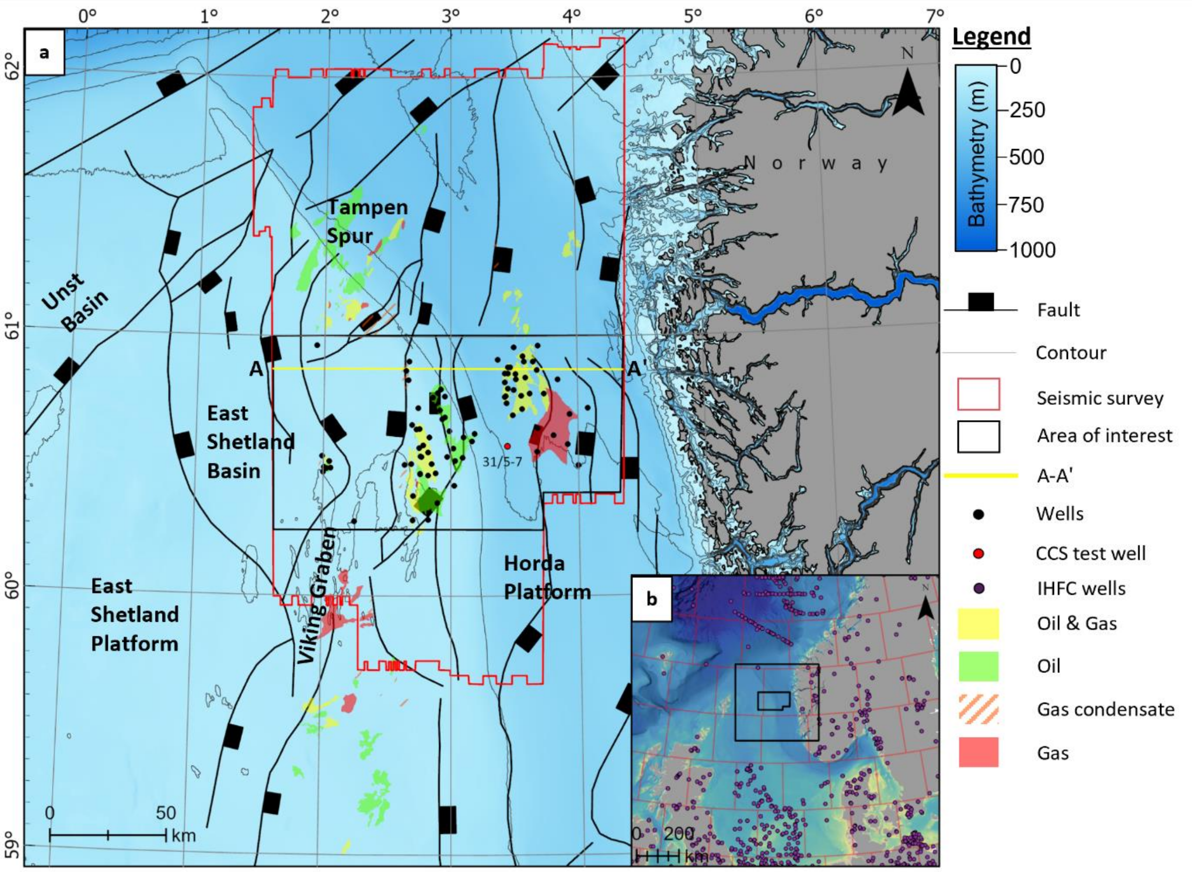


*Figure 1: (a) Study area in the northern North Sea outlining the extent of the CGG NVG survey (displayed in red) offshore the Norwegian continental shelf with structural features displayed from* (Færseth, 1996). *Transect A-A' based on NVG type section (Copestake et al., 2003). Exploration wells displayed are used for calibration. Northern Lights CCS test well 31/5-7 also displayed (red circle) (between Brage and Troll fields). Bathymetry from Generalised Bathymetric Chart of the Oceans (GEBCO)* (Becker et al., 2009; "The GEBCO_2019 Grid - a continuous terrain model of the global oceans and land.," 2019)*. (b) Existing heatflow data from the IHFC (purple points) shows a scarcity of data in the model area (Gosnold & Panda, 2002). Relying on published heatflow grids such as the Davies (2013) shown above (red grid) demonstrates the coarseness of the data when compared to the scale of the model area (Davies, 2013).*

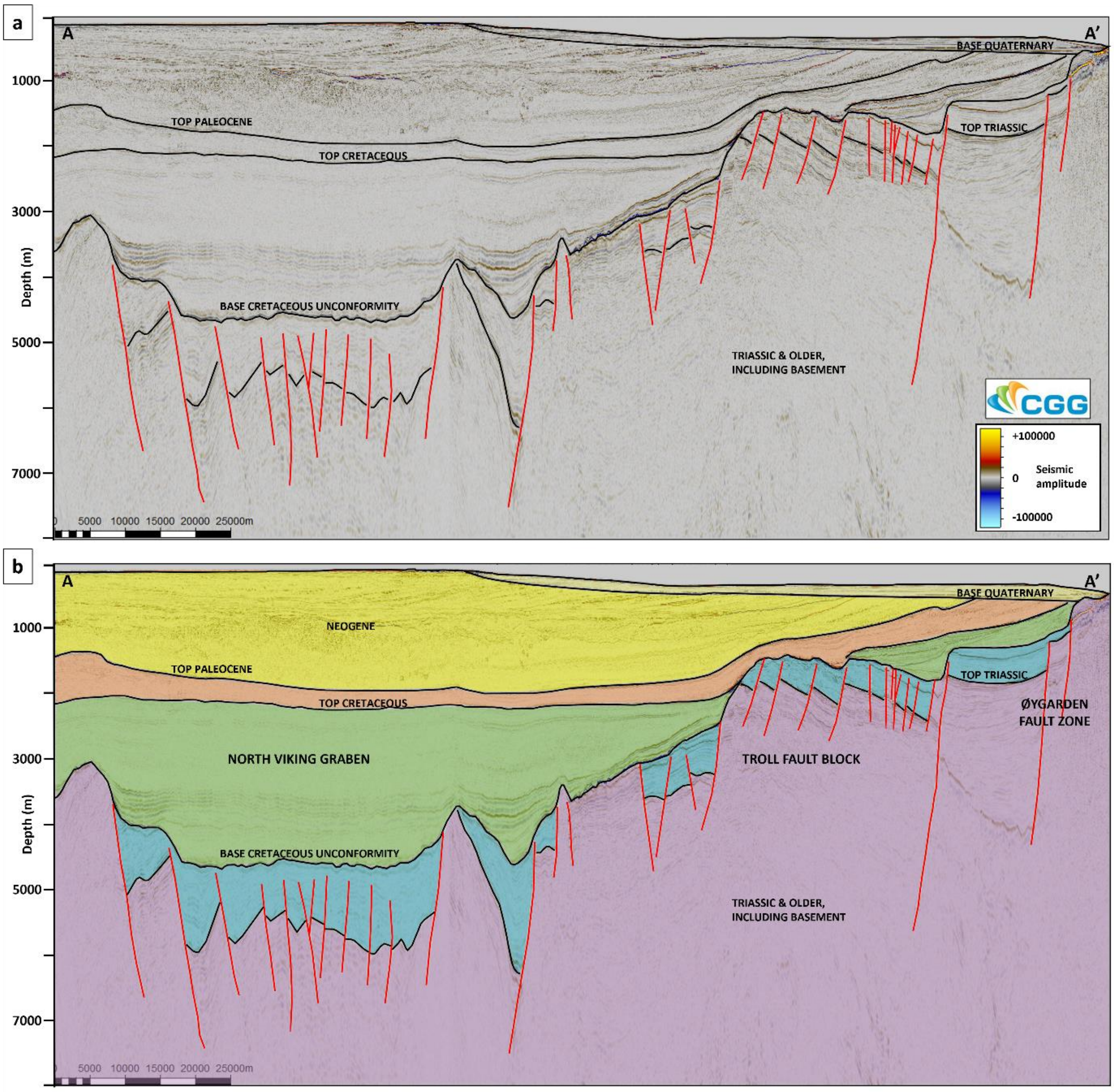


*Figure 2: (a) East West transect A-A' displaying reflection seismic data, annotated with major chronostratigraphic surfaces and structures of note. (b) Overlay of major intervals highlighting the geometry of the NVG in the west of the model area, with tilted fault blocks apparent. The study area is bounded to the west by the East Shetland Basin, with the eastern limits coinciding with the Horda platform. Adapted from (Copestake et al., 2003).*

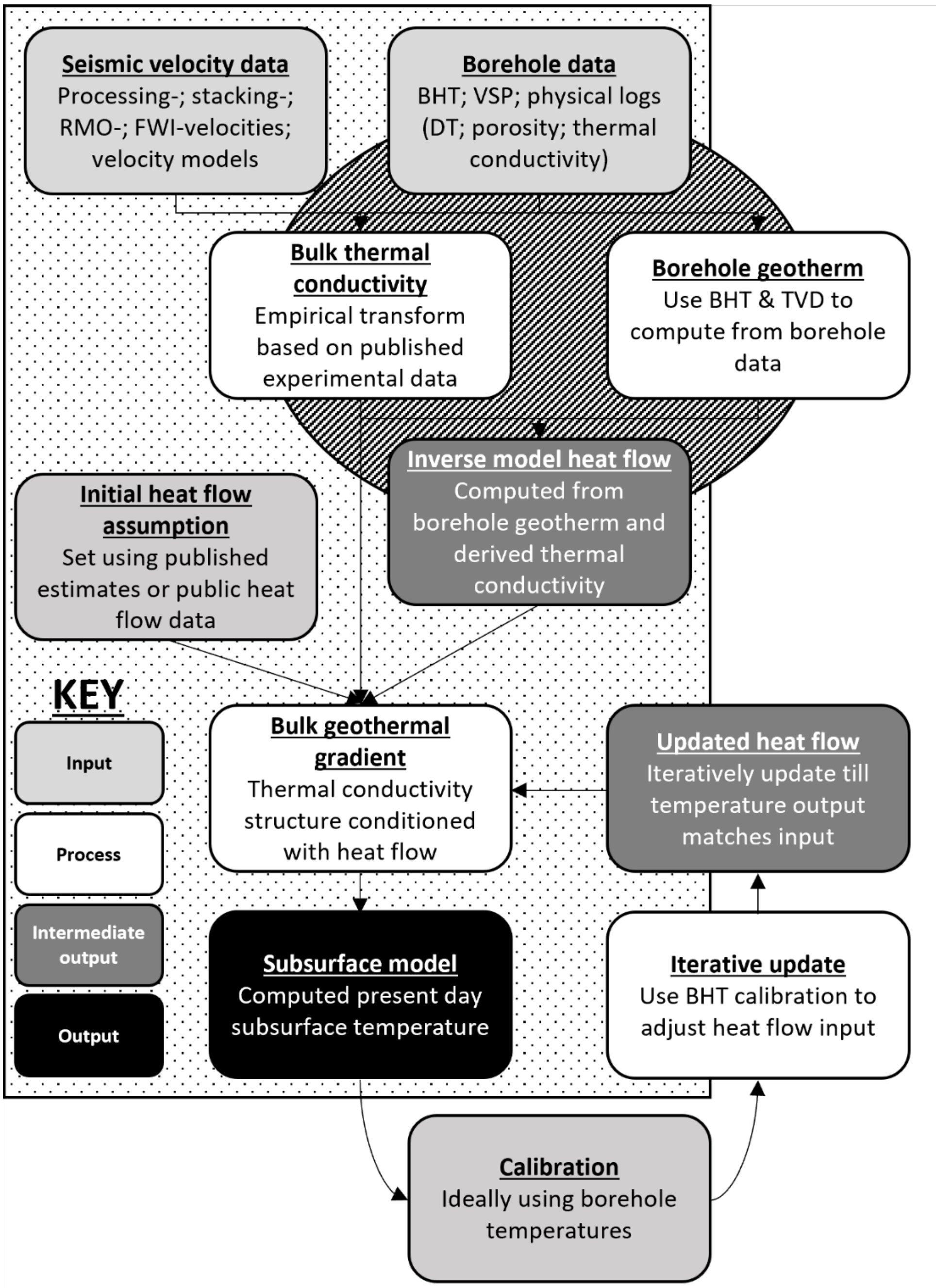


*Figure 3: Model building workflow displayed in terms of input, processing steps and outputs/deliverables. There are two main pathways, a forward modelling pathway (demarcated with the dotted background polygon) where seismic data is used to simulate BHTs; and an inverse modelling pathway (demarcated with the hashed background polygon) where BHTs are used to determine the heatflow conditions needed for it. This allows for an iterative final forward modelling pathway utilising the derived heatflow to arrive at a subsurface temperature model representative of present-day conditions.*

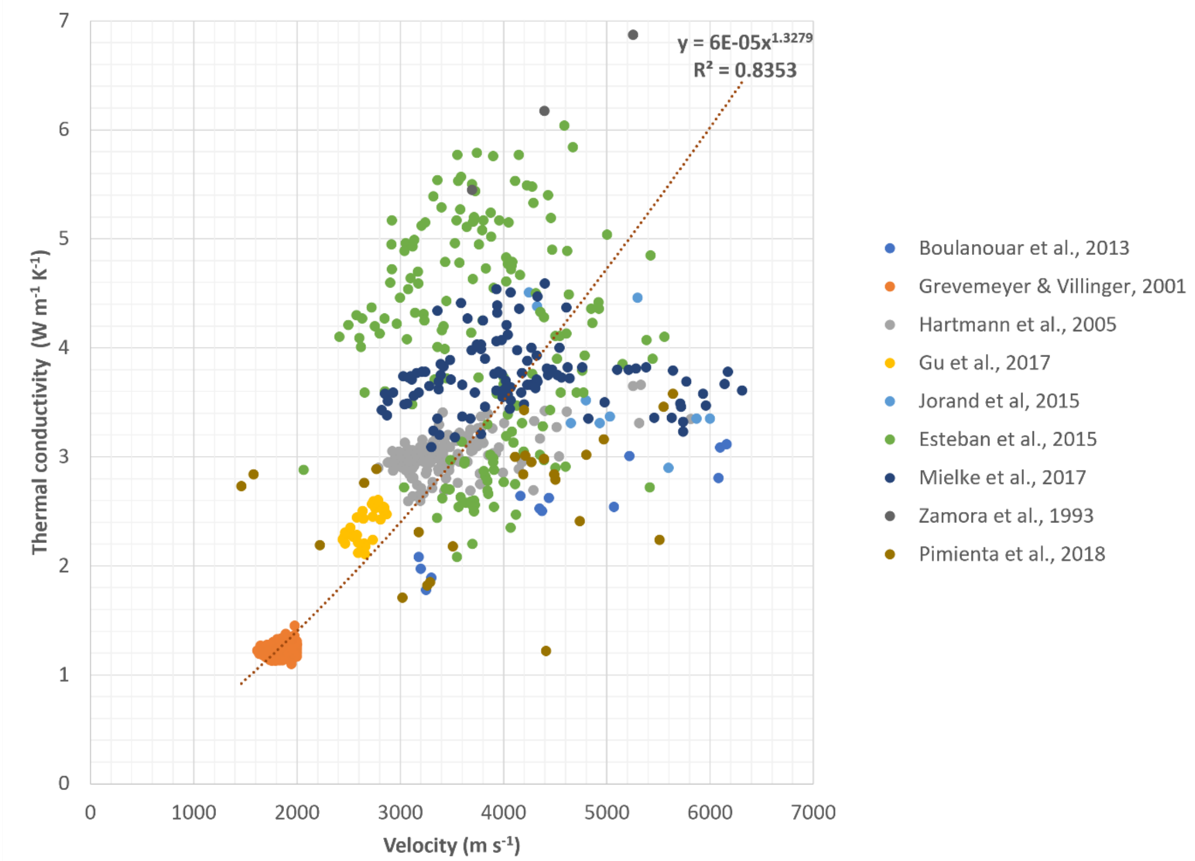


*Figure 4: Bulk shift transform from velocity to thermal conductivity derived from experimental data published in the literature. All points displayed are wet samples with laboratory measurements of both velocity and thermal conductivity having been done with similar tools. This is to both reflect the presence of fluids in the subsurface and to also reduce the variables between displayed data respectively.*

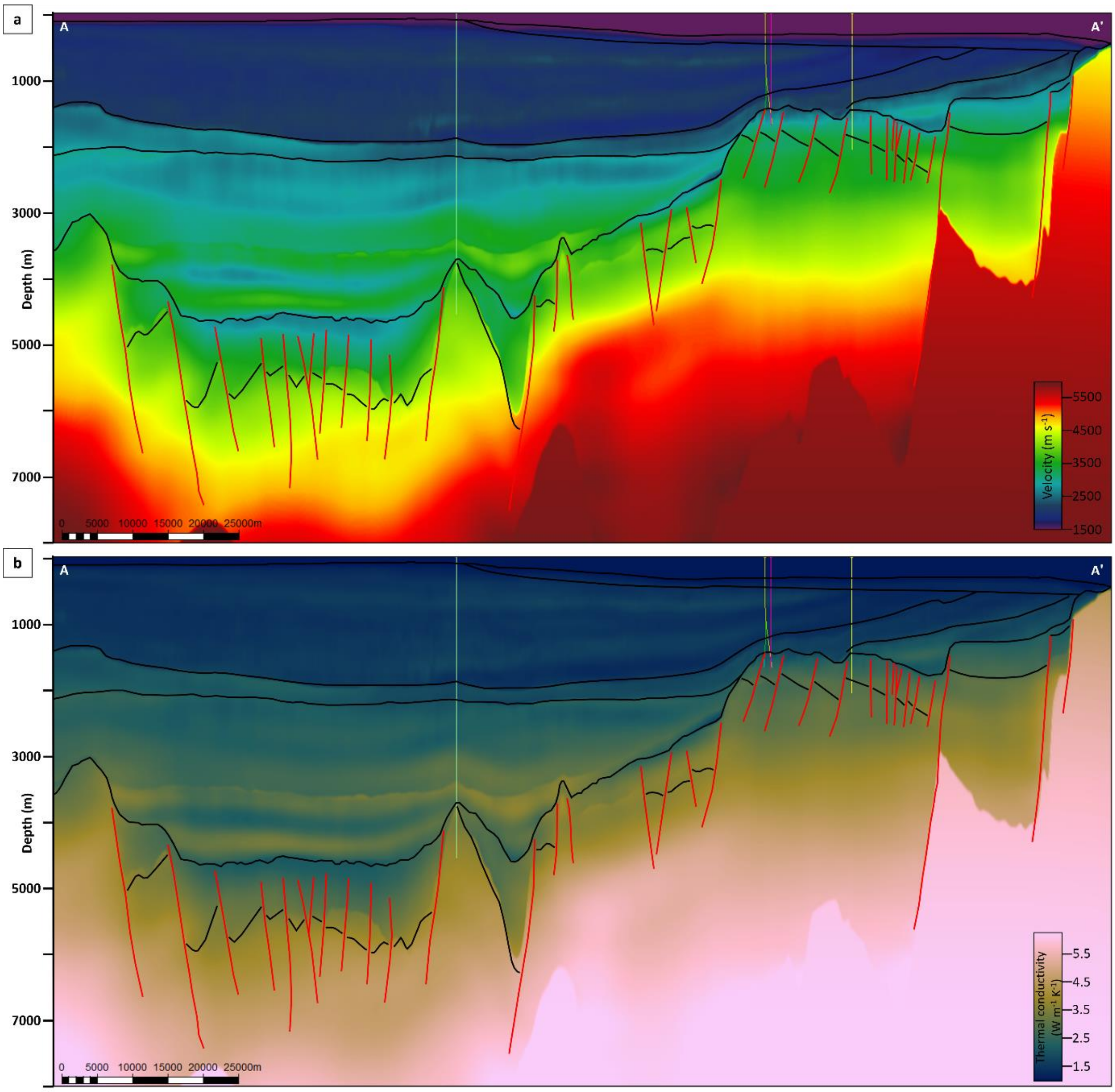


*Figure 5: (a) Interval velocities with wells displayed. (b) Instantaneous thermal conductivity from interval velocities using scientific colour bar* (Crameri et al., 2020).

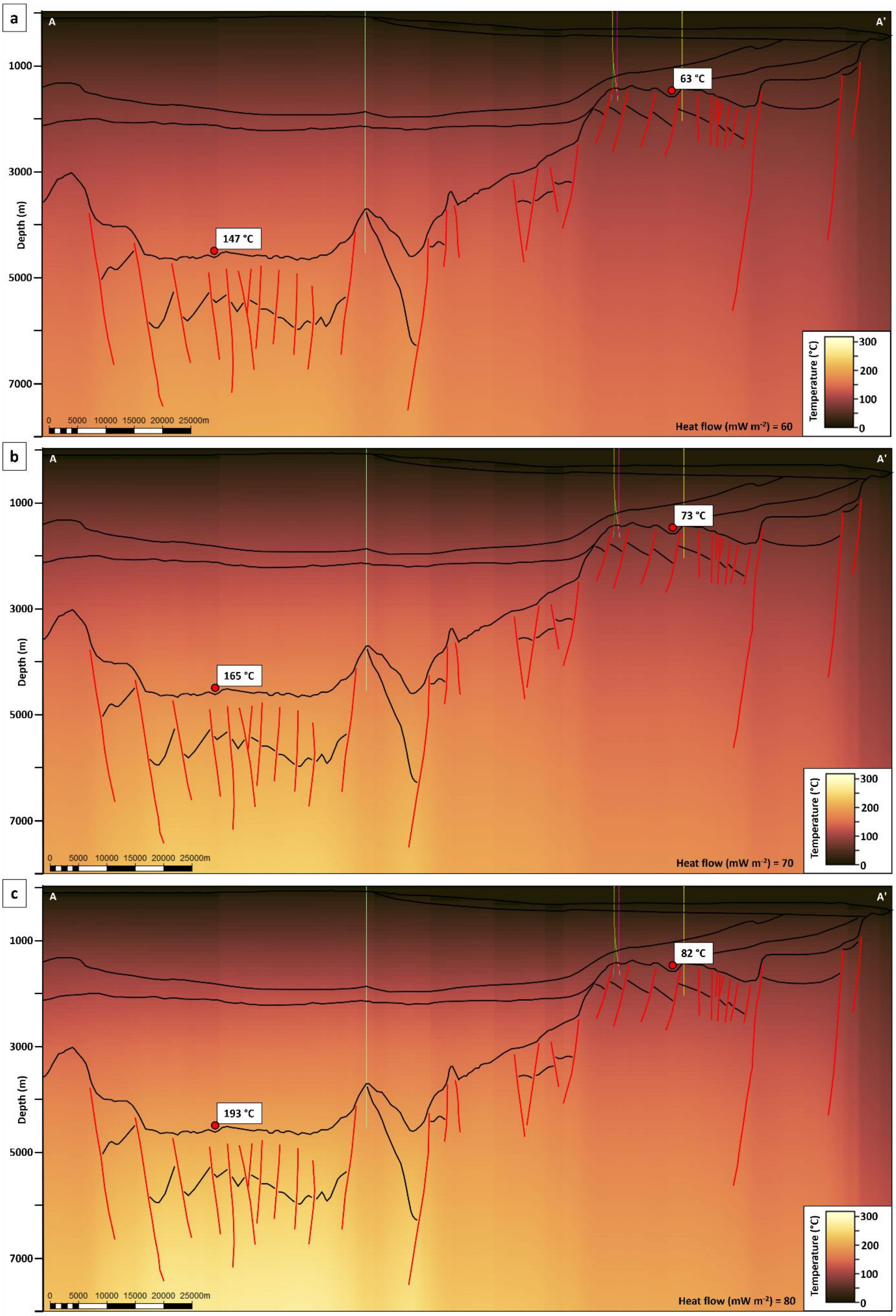


*Figure 6: Forward modelling temperature prediction profiles with transect A-A' overlain for (a) low case (60 mW m$^{-2}$); (b) mid case (70 mW m$^{-2}$); & (c) high case (80 mW m$^{-2}$) heatflow*

*scenarios respectively. Temperature readings in the graben centre and on the graben flank are shown for reference. Highest temperatures are observed in the heart of the graben. Block like appearance is an artefact of the dimensions of the individual voxels comprising the thermal model pillar grid used to represent the subsurface.*

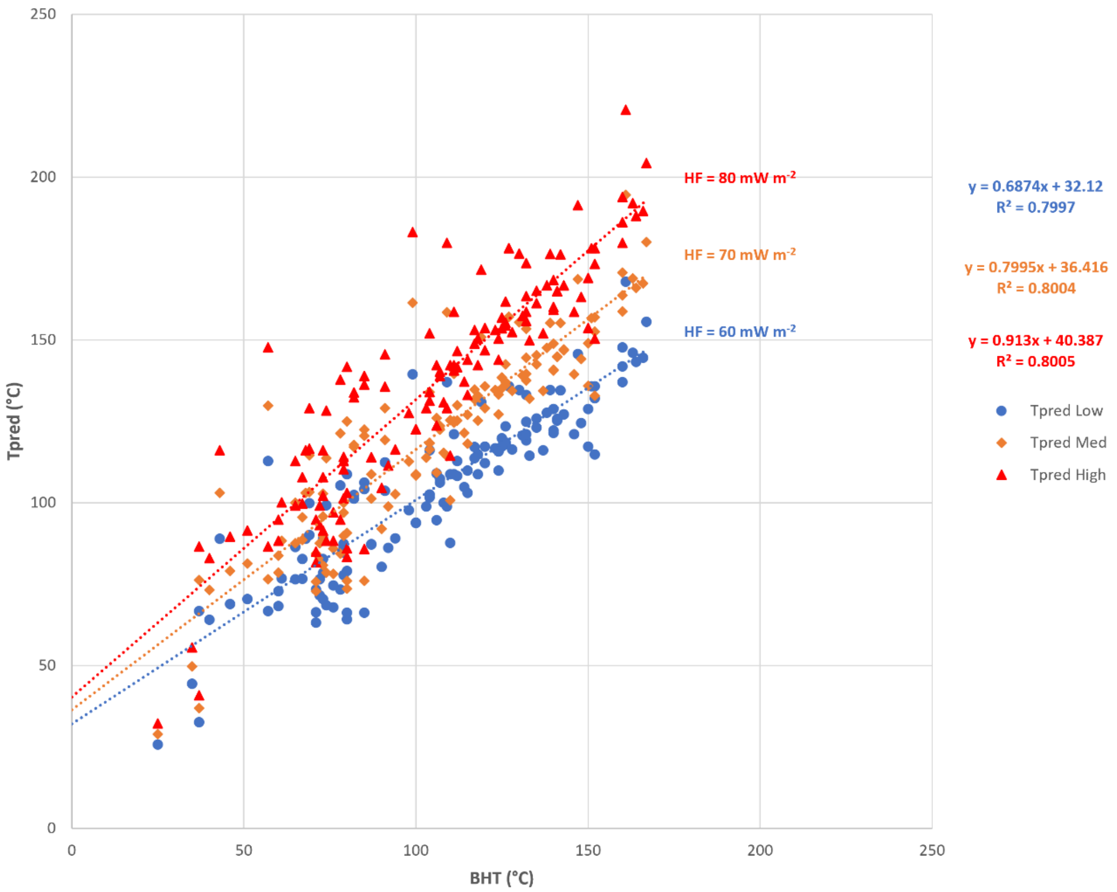


*Figure 7: Cross plot of BHT against predicted temperatures (Tpred) for each of the three forward modelling starting conditions for heatflow (low; mid and high case corresponding to 60; 70 and 80 mW m$^{-2}$ respectively). With increasing input heatflow a corresponding increase is seen in the gradient of the regression line through that set of points.*

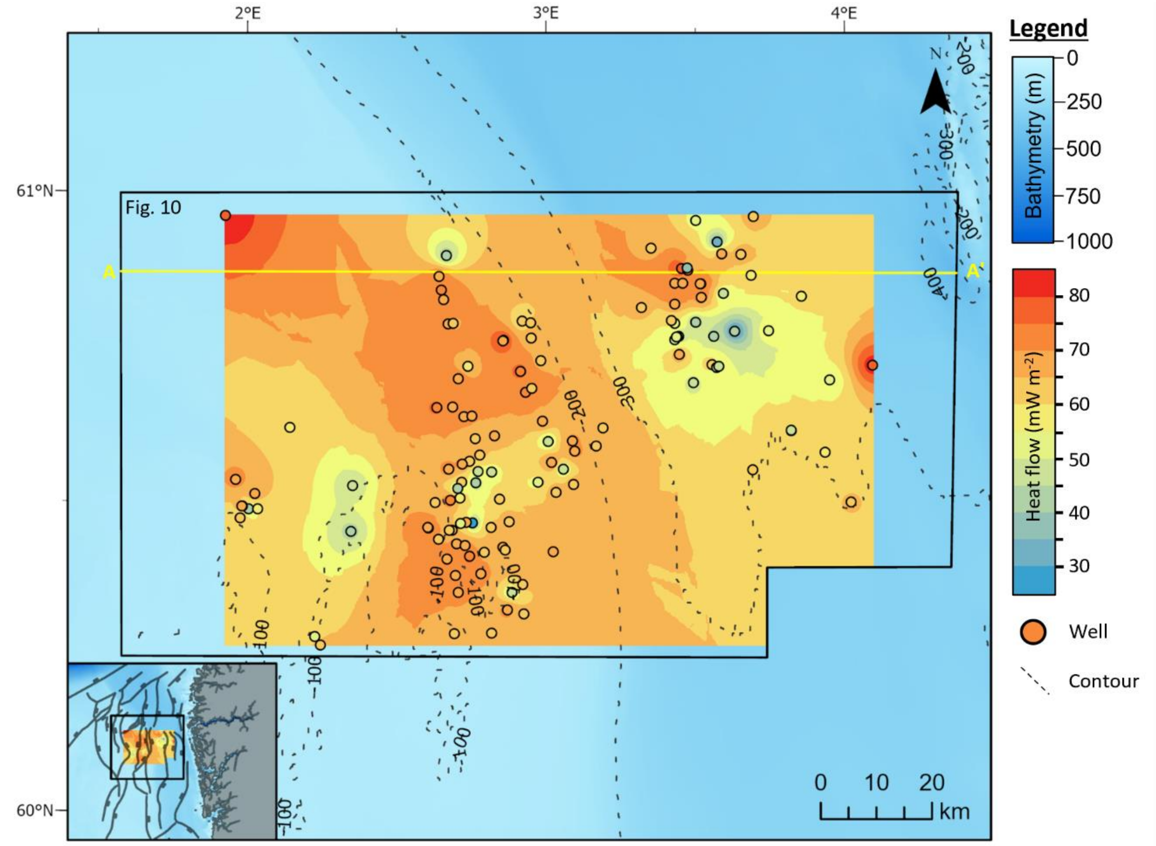


*Figure 8: Inverse modelling of heatflow at each well location from BHT and TVD. This data is used to interpolate heatflow across the model area. Interpolated heatflow shows lateral variability at much higher resolution than published global grids.*

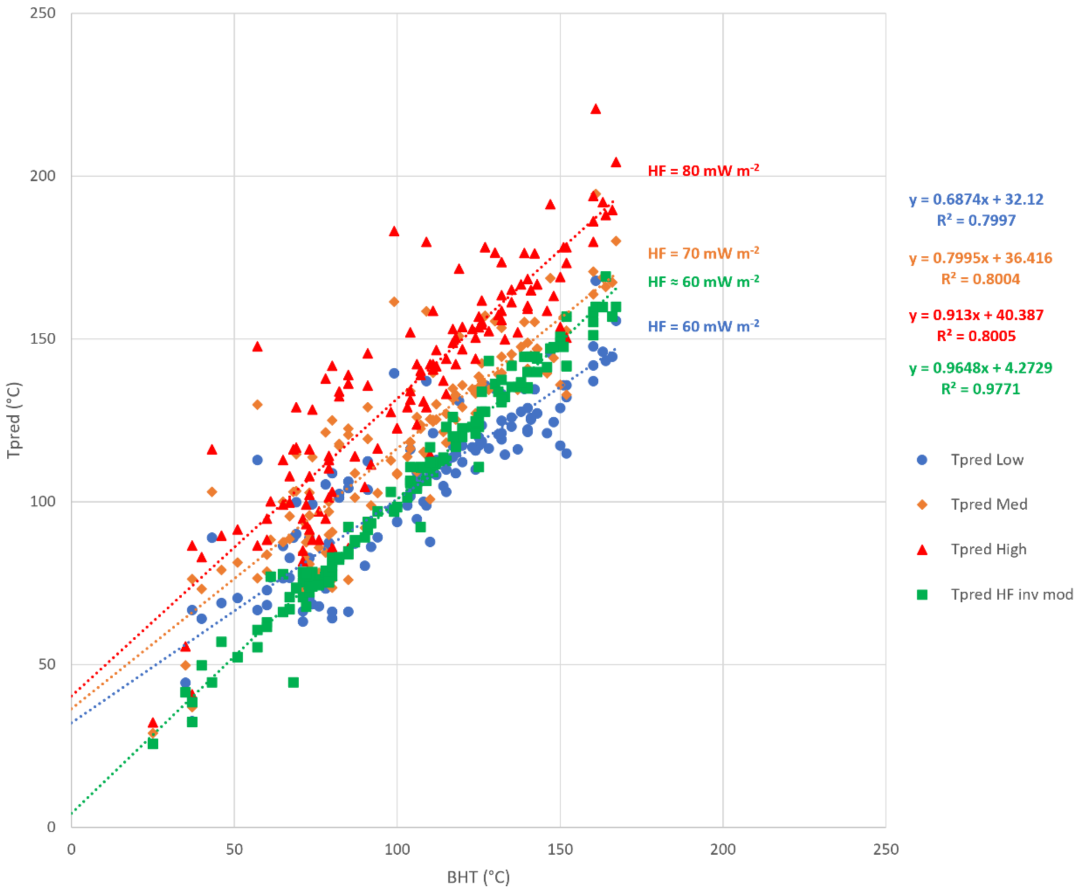


*Figure 9: Cross plot of BHT against predicted temperatures (Tpred) with model results using inverse modelling of heatflow across the area displayed (green squares). R-squared for this regression suggests a good statistical fit, the results having a gradient verging on 1:1 as would be expected for results best reflecting the actual subsurface temperature conditions. Furthermore, the spread of points is much narrower for this modelling outcome.*

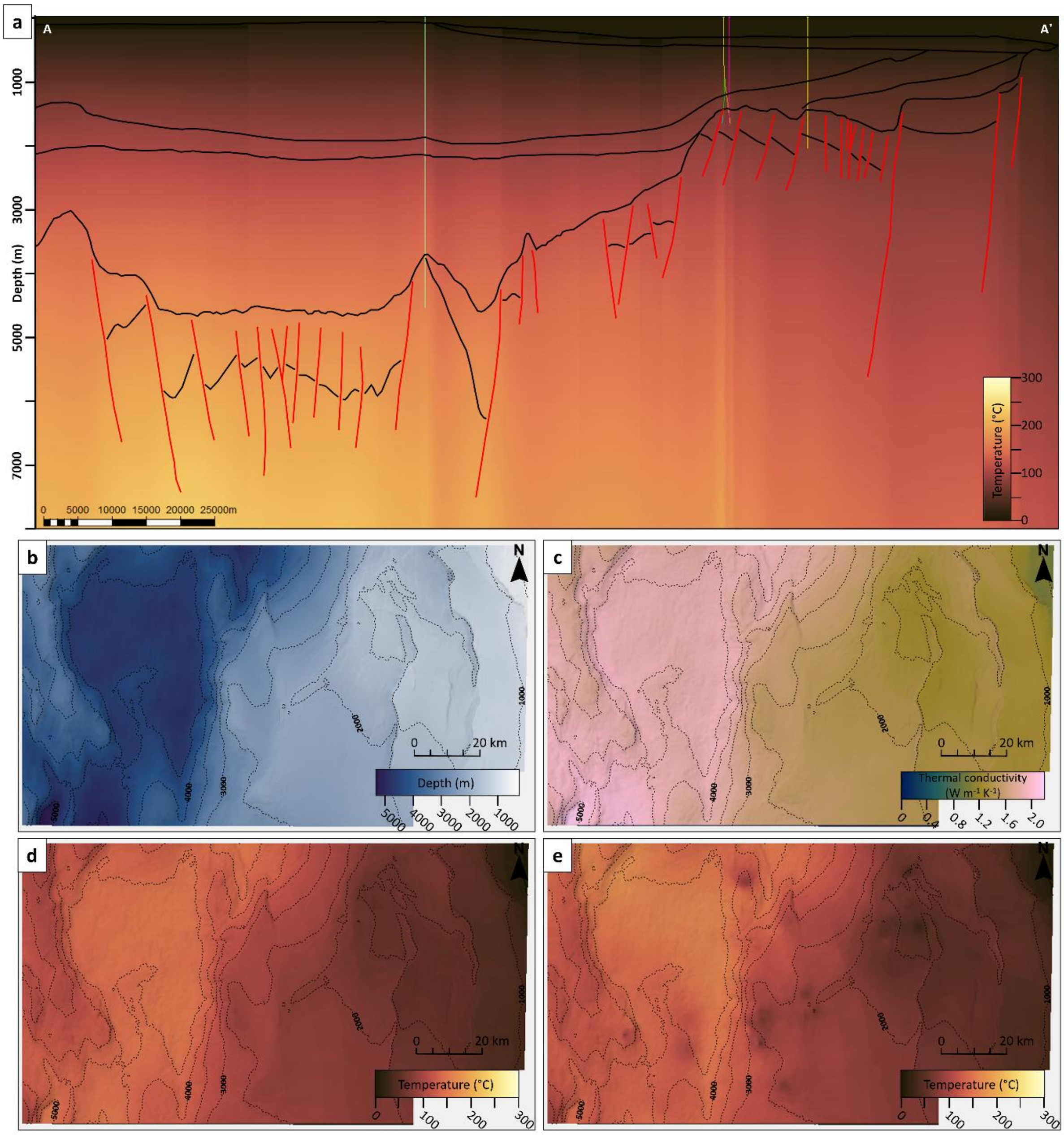


*Figure 10: (a) Final temperature model produced using inverse modelling of heatflow overlain on transect A-A' (with well paths and BCU displayed). (b) BCU in depth with 500 m interval contours shown. (c) RMS amplitude extraction of derived thermal conductivity at BCU. Where BCU is interpreted to shallow towards the northeast, there are correspondingly low conductivities that reflect Quaternary sediments in this region instead. (d) Low case prediction of temperature along BCU. (e) Final temperature prediction using inverse modelled heatflow along BCU. Comparing with (d) some differences are apparent. Bulls eye like temperature anomalies in the northeast are likely the translation of the interpolated heatflow (see Fig. 8).*

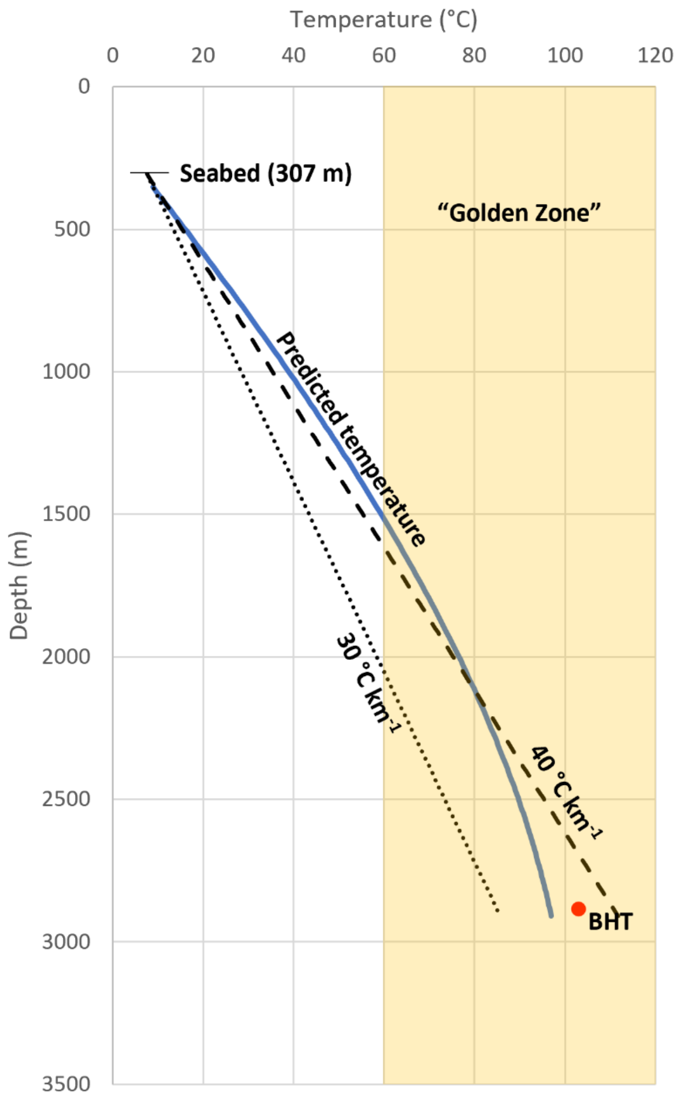


*Figure 11: Modelled subsurface temperature at well 31/5-7 (see Fig. 1) as part of the Northern Lights CCS project. "Golden zone" for sandstone reservoirs is also displayed for context. BHT displayed is from October 2020 data release of preliminary results, indicating that prediction is in close agreement with what was discovered downhole (prediction is offset by 6 °C to reported BHT). Also shown are the temperature profiles taken from seabed assuming a constant linear geothermal gradient. Typically used geothermal gradients in basin modelling are 30 °C km$^{-1}$ (dotted line) and 40 °C km$^{-1}$ (dashed line). These are displayed to show how much subsurface temperature predictions may vary using standard processes, particularly at bottom hole (up to ±13 °C).*

# A.1 Appendix

| Source | Geothermal gradient (°C/km) | Heatflow (mW $m^{-2}$) |
|---|---|---|
| **(Harper, 1971)** | 29.7 | 49.8 – 62.0 |
| **(Evans, 1977)** | | 63 |
| **(Brigaud et al., 1992)** | 31.8 – 36.3 | 50 – 65 |
| **(Leadholm et al., 1985)** | 30 – 35 | 58.6 – 67 |
| **(Justwan et al., 2006)** | | 52.3 |
| **(Cornford, 1998)** | | 60 – 82 |
| **(Lucazeau & Le Douaran, 1985)** | | 65 |
| **(Goff, 1983)** | 32 | 57 – 65 |
| **(Rüpke et al., 2008)** | 30 – 40 | |
| **(Ritter et al., 2004)** | | 65 |

*Table A.1: Some examples of reported geothermal gradient and heatflow for the NVG and surrounding basins from the literature.*