1. Research field
   1. Research questions
   2. Research aim and objectives
   3. Take home message
2. Research design
   1. Methods
   2. Modeling code
   3. Case studies
      1. Bilston Glen
      2. Dawdon
3. Chapters
   1. Chapter 1:  heat extraction from borehole heat exchanger and shallow recharge potential
      1. GSHP technology
      2. Heat exchanger
   2. Chapter 2: Heat sources and distribution in coal mines
      1. Conceptual models
      2. Hydrology
      3. Heat flow
   3. Chapter 3: Mine modelling – importance of geometric features from mine plans
      1. Mine geometry (type of mining / mining methods)
   4. Conclusion: geothermal potential and sustainability opportunities
4. Time plan (Gantt chart)
5. Resources
   1. Budget (ressources available / needed)
   2. Data management plan
      1. Data repositories and version control
      2. Codes
   3. Training
   4. Conferences
   5. Teaching experiences
6. Supervisory arrangements and collaborations
7. Ethical and Health and safety considerations
8. References

In addition, the large voids provide significant potential for heat exchange between the host rock and the mine-water over the whole depth range spanned by the worked coal seams (Banks et al., 2003).

First of all, an inventory of all the existing data ill be made based on available literature and database, and kept updated throughout the PhD lifetime to account for new acquisitions and publications.

Temperature data will be analyzed and classified according to their geographical/geological location, the measurement types, context of the measurements (i.e. flooding conditions, gravity-driven discharge, pumped water) and timing of the acquisition. In areas where flooding is still occurring, changes in fluid temperature or geochemical composition through time might highlight the existence of significant interconnections at specific mine levels. Interconnections between the mine workings/shafts and recharge/discharge zones at the surface or with surrounding aquifers would also be supported by a comparison with structural/geological maps and/or mine abandonment plans. This preliminary analysis will be done to identify the possible impact or contribution of local heat sources / recharge areas on the measured temperature values (i.e. presence of fault, type of rocks and volume of rock type around the monitoring point) or exclude biased data from the analysis. A database will be created to facilitate the access, storage, management and reference to the data throughout the PhD research work and for future access.

This include a good definition of the hydraulic system (i.e.recharge/discharge areas) and of the heat sources and heat transfer mechanisms (i.e. natural heat flow, solar flux, radioactive decay, and mine-related heat sources). Mining activities are indeed likely to have caused perturbations of natural thermal regime of the mine systems (i.e. change in groundwater flow) and participated to the development of specific processes responsible for anomalous temperature distribution (Jessop 1995).

In the Carrington-1 well, situated in the central part of the coalfield, the total thickness for the Coal Measures, Passage Formation, Upper Limestone, Limestone Coal and Lower Limestone Formation are about 250, 210, 165, 175 and 85 m, respectively. Carboniferous strata from the West Lothian Oil shale and Gullane Formations are expected to extend further down by at least 500 m (Monaghan and Brown, 2014). A 70 m thick interval from the Clyde Plateau Volcanic Formation, which represents a volcanic sequence within the Gullane Formation, has been crossed by the Midlothian-1 well, situated further west.

Long-term climatic effects might therefore have caused transient changes to the near-surface part of the geothermal gradient and impacted the subsurface temperature down to to ~2 km depth. Assuming linear geothermal gradient based on the average surface temperature would therefore lead to an underestimation of the true background "steady state" conductive heat fluxes and thus the size of geothermal resource at depth (e.g. Westaway and Younger, 2013; Busby et al., 2015)

Renz et al., (2009) showed that for hydraulic conductivity contrasts between the mine void $K\_{mine}$ and the rock ${K\_rock} < 1 x 10^5 m/s$, the higher the $K\_{mine}$, the smaller the decrease in extraction temperature. For higher contrasts, the higher the $K\_{mine}$, the greater/quicker is the decrease in extraction temperature. For low K, the cold water mainly flows within the lowermost galleries and is separated from the warm water above due to density effects. From $K = 1 x 10^4 m/s$, mixing of warm and cold water occur as a result of the establishment of free convection within the mine voids, leading to faster transport of cold water toward the extraction well (layering of cold/warm water is therefore only possible for a certain range of hydraulic conductivity values).

\section{Benchmarks, mathematical and analytical expressions}

Here we need to describe:

- convection and conduction of heat through the porous medium

- what is the proportion of heat absorbed / released by the porous media

- how much heat is diffused from surrounding rocks (source term??)

Benchmarks:

The equation for a constant boundary condition:

\[

(erfc(\frac{x-vt}{\sqrt{4\alpha t}})+\exp({\frac{xv}{\alpha}})\*erfc(\frac{x+v t}{\sqrt{4\alpha t}})\*\frac{T\_0}{2}

\]

The solution for a pulse equation:

\[

(erfc(\frac{x-v(t+\Delta t)}{\sqrt{4\alpha (t+\Delta t)}})+\exp({\frac{xv}{\alpha}})\*erfc(\frac{x+v (t+\Delta t)}{\sqrt{4\alpha (t+\Delta t)}})\*\frac{T\_0}{2}

\]

\subsection{Dual-continuum approach}

Heat extraction has generally been simulated either in constant extraction mode (i.e.Hecht-Mendez et al. ) or cyclical annual extraction loads (i.e. Zanchini et al., Angelotti et al., ). This is generally modelled using an imposed heat exchange rate as a source/sink term at the line source (i.e. Christodoulides et al), which assumed that the heat transfer through the tube walls is instantaneous. Angelotti et al., however suggested to use the results from the interaction between the BHE and the surrounding porous medium as input to the BHE heat load in order better assess the BHE energy performance of the BHE by coupling it with its thermal impact in the soil. Similarly, Bozzoli et al. and Brunetti et al. (2017) modeled the heat transfer between the ﬂuid inside the tube and the solid phase (e.g., tube or grout) using a Robin boundary condition corresponding to the heat transfer coeﬃcient that depend on the ﬂuid parameters.

Brunetti et al., 2017 developed a computationally eﬃcient pseudo-3D model numerical model accounting for ground heterogeneities and groundwater flow, combining 1D heat transport in the BHE and horizontal heat transfer in the surrounding subsurface soil, represented as a pile of 2D plans. The authors assumed that heat transfer in the soil domain occurs only in the 2D horizontal plans, as well as in the vertical 1D BHE. 1D equations for the BHE temperature were solved analytically using a semi-implicit approach (i.e. python script) considering heat advection in the tube and convective heat transfer between tube and fluid, by applying periodic Dirichlet BC on tubes. In each horizontal layer (which have their own thermal/hydraulic properties), 2 nodes representing the BHE were attributed the Robin BC representing the convective heat transfer between the ﬂuid and the solid (i.e. function of the inner fluid and tube wall temperatures). This was set up as a source term, thus ignoring the convection effect in the BHE and other vertical fluxes that might enhance heat exchanges. There, both Cauchy- and Neumann-type BC were specified on the remaining boundaries of the model, to account for the presence or not of incoming groundwater flow. A Python script was used to simultaneously solve for the 1D advective heat transport in the tube and its interaction with the 2D numerical model.

Similarly, Biglarian et al., (2017) extended the thermal resistance-capacity model to account for the fluid transport through the U-tube, vertical conduction in the ground, grout and pipe and the variations in heat loading with depth, along the borehole, to solve for temperature distribution along the borehole walls and around the BHE in both the short and long term. The governing equations of the two regions (i.e. borehole and outside region) were solved iterative in each time step. The results showed that the temperature distribution (vertical profile) along the borehole wall and heat loading profiles are not distributed uniformly along the borehole. This model assumed a heat loading from the borehole wall surface (Neumann BC) and neglected heat flux in far field boundaries in radial and vertical direction (ground undisturbed temperature). \\

\[

\Delta T(x,y,z, t) = \frac{q}{8 \pi t \sqrt{\lamda\_x \lambda\_y}} exp(-\frac{(x-u\_T t)^2}{\frac{4 \lamda\_x t}{\rho\_m \c\_m } } - \frac{y^2}{\frac{4 \lamda\_y t}{\rho\_m \c\_m }}) \times (erf(\frac{z}{\frac{4 \lamda\_z t}{\rho\_m \c\_m }} - erf(\frac{z - H}{\sqrt{\frac{4 \lamda\_z t}{\rho\_m \c\_m }}} )

\]

Signorelli et al showed similar far-field temperature change were obtained after a one year operation period for different production scenarios, each totaling 300 kWh/month for a total 1800h operation time per year. Both studies suggested that the system reaches a sustainable operation level after 12 years (Rivera et al., 2015).

Analytical solutions for heat transport can generally describe on mechanism at the time, and heat transfers in complex system generally requires to be solve using numerical approximations. This section summarized different approaches used in the literature to describe heat transfers induces by heat injection/extraction at borehole heat exchangers (BHE) in porous media.

The transient heat diffusion equation can be written as follow. The first term on the right side represents the advective term, with the mass flow a characterized by the product of the water density $\rho\_v$ and velocity $v$, and the second term is the conductive heat transfer.

\[

<\rho\_m \beta\_m> \frac{dT}{dt} = - \rho\_w \beta\_w v\_w \frac{dT}{dx} + \nabla(\lambda \nabla T)

\]

The matrix heat conductivity and capacity can be defined as the product of the porous matrix and water properties as followed.

\[ k\_m = \phi\_e \lambda\_w + (1-\phi\_e) \lambda\_r \]

\[ <\rho\_m \beta\_m> = \phi\_e \rho\_w \beta\_w + (1- \phi\_e) \rho\_r \beta\_r \]

The heat dispersion coefficient is defined as follow:

\[ \alpha = \frac{k\_m}{\rho \beta} \]

Raymond and Therrien (2008) estimated the total geothermal energy content of rock from the sum of the energy associated with each ﬂooded underground section:

\[E\_r = V z \alpha <\rho\_m \beta\_m> \]

with V the volume of water in a mine section, z its average depth, $\alpha$ the measured geothermal gradient and $E\_r$ the energy of a given section. The rate $E\_hp$ at which geothermal energy can be extracted using heat pumps was calculated as foillowed:

\[E\_hp = Q(Tp-T\_r)\rho\_w \beta\_w \],

with Q the average flow rate, Tp water temperature and Tr the exchanger return temperature. In a porous rock, the total energy extractable from cooling rocks can be calculated as follow:

\[E = \Delta T V <\rho\_m \beta\_m> \]

The power exchange of water in a system with colder fluid reinjection can thus be written as followed:

\[P= \frac{C \phi \rho \beta}{t} (T\_{in} - T\_{system}) \]

\subsection{Flow in fractures or porous media}

The Lauwerier-Pruess-Bodvarsson model (Lauwerier, 1955; Pruess and Bodvarsson, 1983) set up a formula to predict the temperature of water $T\_s$ in a fracture of aperture 2b at time t and distance x, where water injection commences at x= 0 and t =0.

\[

T\_s - T\_r = (T\_e - T\_r) erfc (\frac{c}{2b\rho\_x c\_w v\_w}\sqrt{\frac{\lambda\rho c}{(t-\frac{xA\rho\_w}{q})}}) U(\theta-\epsilon)

\]

with $T\_e$ the injected water temperature and $T\_r$ the initial rock temperature. With flow velocity v and injection flow rate q, the equation can be rewritten for a generalized flow channel:

\[

T\_s - T\_r = (T\_e - T\_r) erfc (\frac{xP}{2qc\_w}\sqrt{\frac{\lambda\rho c}{(t-\frac{x}{v\_w})}}) U(\theta-\epsilon)

\]

The Rodriguez and Diaz model (2009) expressed the heat power gain and water exit temperature as a succession of time increments.

\subsection{Temperature change due to periodic heating}

The analytical solution for temperature change in a semi-infinite half space under periodic heating / seasonal change in temperature $T\_s = T\_0 + \Delta T cos(w t)$ (Turcotte and Schuber, XXX) is:

\[

T = T\_0 + \Delta T + \exp{(-y \sqrt{\frac{w}{2 \alpha}})} cos(wt - y \sqrt{\frac{w}{2 \alpha})}

\]

with $w = \frac{2 \pi}{\Theta}$.\\

For a instantaneous cooling / heating of a semi-infinite half-space, where a constant temperature $T\_0$ is applied on the top boundary condition of a domain of initial temperature $T\_1$, with $T\_1 \textless T\_0$, the temperature profile or temperature change can be calculated at a time $t$ or distance $z$ from the boundary using the following equation (Dirichlet heat conduction problem, with q = 0 on the lower boundary):

\[

T = T\_1 + (T\_0 - T\_1) erfc(\frac{z}{\sqrt{4 \alpha t}})

\]

with $erfc$ the complementary error function and $\alpha = \frac{\lambda}{c \rho}$ the thermal diffusivity (m2/s). \\

In case of a heat flux boundary condition, the change in temperature on the surface $T\_s$ and at a distance $z$, $T\_z$ in the half space can be defined as follow (using $T \xrightarrow{}T\_1 $ as $ y \xrightarrow{ } \infty$):

\[

T\_y = T\_1 + \frac{2q\_0}{k} (\sqrt{\frac{\alpha t}{\pi}} \exp{(\frac{-z^2}{4 \alpha t})} - \frac{z}{2} erfc \frac{z}{ \sqrt{4 \alpha t}})

\]

and

\[

T\_s = T\_1 + \frac{2q\_0}{k}\sqrt{\frac{\alpha t}{\pi}}

\]

Angelotti et al., modelled in 3D the temperature distribution (i.e. thermal plume) in the porous aquifer, both in the BHE surroundings and downstream to it, under seasonal production and for different groundwater velocity. The authors moreover performed a quantitative assessment of the heat rate increase due to groundwater flow, allowing to get an insight on insight into its role on the yearly thermal balance in the ground. The temperature response of the medium is here expressed as follow:

\[

T(x,y,t) = T\_g0 + \frac{q}{4 \pi \lambda\_m} exp(\frac{u\_x}{2 \alpha\_m}) \int\_{0}^{r^2 / (4 \alpha\_m t)} \frac{1}{\psi} \exp(-\frac{1}{\psi}-\frac{u^2r^2\psi}{16\alpha\_m^2}) d\psi

\]

with $u= v\_x \frac{c\_w}{c\_m}$ the effective water velocity. A steady state temperature might be reached for high simulation time and can be defined as follow:

\[

T(x,y,t) = T\_g0 + \frac{q}{2 \pi \lambda\_m} exp(\frac{u\_x}{2 \alpha\_m}) K\_0 (\frac{u\sqrt{x^2+y^2}}{2\alpha\_m t})

\]

\section{Surface boundary condition}

Hein et al. (2016) built a comprehensive numerical model including hydraulic and thermal heat transport to consider the effects of heterogeneous subsurface properties, groundwater flow, geothermal gradient, heat flux, as well as a varying surface temperature, on the evolution of BHE outflow and soil temperatures. Is was combined with a heat pump model aiming to include the dynamics of heat pump efficiency (i.e. varying heat loads due to specific COP characteristics and building heat demand). The analysis was performed for a typical single family house GSHP in the Leipzig area in Germany. Results showed that the recovery of shallow geothermal energy only accounts for about 89\% of the energy extracted in the first year and then outflow and soil temperature decrease until quasi-steady-state. Both convection on the pipe and dispersion in the grout zone (i.e. with Cauchy BC) were used for the BHE. The inlet temperature for the BHE was calculated iteratively from the prescribe BHE thermal load that is function of the heat flux and flow rate, including the COP of the heat pump. Bortoloni et al. compared the influence of different BC at the ground surface in modeling shallow horizontal ground heat exchangers. They found out that applying the surface temperature (Dirichlet) in contrast to heat flux (Neumann) and mixed (Cauchy) boundary conditions is appropriate with respect to the BHE fluid temperature. In Hein et al. (2016), a stepwise function of monthly mean temperatures was imposed as a Dirichlet type BC on the ground surface, and was compared to situations with ground surface temperature BC only and bottom heat flux boundary condition only. Results showed that after the first year, the ground surface temperature contributes to about 98\% of the energy balance, against only 2\% for the bottom heat flux. It is finally interesting to note that the non linear distribution of the temperature due to circulating fluid in the BHE (i.e. dynamic behavior of the BHE) tends to disappear away from the BHE (Salah Saadi et al., 2017).

A sinusoidal temperature model representing the variations in heat transferred to the soil at the surface was derived by Hillel (1982), using a sinusoidal forcing function and is described as followed (Ozgener et al., 2013). The soil temperature at time t and depth z can be described as follow:

\[

T(z,t) = T\_m + A\_z sin(\frac{2 \pi}{P} (t-t\_0) - \gamma z - \frac{\pi}{2})

\]

with $\gamma = \sqrt{\frac{\pi}{\alpha P}}$ the inverse damping depth, where P is the period of the oscillations and $\alpha$ is the thermal diffusivity. Tm the mean surface temperature, t0 the time lag needed for the surface soil temperature to reach Tm, Az the amplitude of temperature wave at depth z, time t, which decay exponentially with depth. It is defined as follow:

\[

A\_z = A\_0 \exp(- \gamma t)

\]

The effect of undisturbed ground temperature was assessed by Radioti (2017), taking into account daily and annual temperature variations, using 3D models. The undisturbed temperature at depth z and time t was calculated as followed:

\[

T(z,t) = T\_m - A exp(-z \sqrt{\frac{\pi}{365 \alpha}}) cos(\frac{2\pi}{365}(t-t\_{T\_{min}}-\frac{D}{2}\sqrt{\frac{365}{\pi\alpha}})) + T\_g(z)

\]

with Tm the mean annual surface temperature, A the amplitude of air temperature oscillations, $t\_{T\_min}$ the day number corresponding to the minimum temperature, $\alpha$ the equivalent ground daily thermal diffusivity and $T\_g$ the product of the geothermal gradient by the testing depth. Results showed that seasonal variations can be detected down to 18m. At that depth, the ground temperature equals the mean annual air temperature and below, it follows the local geothermal gradient. Results for the numerical simulation show that heating at a rate of 150 W/m along the borehole modified the temperature gradient at the location of the boreholes down to a depth of 100 m after 20 years, and reached grater depth after 45 years. Curvature of the temperature profile evolved in time due to increasing amount of heat added to the ground. The depth-average temperature of the ground which is not influenced by the weather conditions (below 18 m) increases at a mean rate of 0.03 °C/year for the first 45 yrs.

Daily soil temperature variations are relatively difficult to model due to short term weather variations, seasonal variations, moisture content of soil, and heterogeneous thermal conductivities. Ozgener et al. (2013) attempted to improve a model predicting daily change in near-surface ground temperature, down to 20 cm depth. In their study, daily soil temperature variations modelled from daily air temperature as sinusoidal function of time and depth. Considering shallow sub-surface, they assumed that surface temperature equals the average annual air temperature. Transient heat flow principles were used with assumptions of one dimensional heat flow, homogeneous soil, and constant thermal diffusivity.

Numerical analysis of ground temperature variation has also been performed by Singh and Sharma (2017) using FDM, based on time variant ambient air temperature and solar radiation. Heat conduction through the ground surface is the results of the combination of convective heat transfer between ground surface (Ty=0) and air (Ta), that depends on the wind velocity, minus the longwave thermal radiations emitted from the ground, that depends on the emissivity of soil ($\epsilon \Delta R$), plus the solar radiation (S) absorbed by the ground surface. This was written in the form of a general convective heat transfer boundary condition, used to calculate the hourly average effective ground surface temperature used as a time-variant first type boundary condition to the numerical model. The bottom BC was set as a constant temperature corresponding to the annual mean effective temperature. Results of the numerical solution were validated against experimental measurement of ground temperature in India.

\[

q = SR - LR - CE - LE

\]

with CE (convective heat flux due to wind velocity) and LE (latent heat due to evaporation) are given by Mihalakakou et al., and LR (longwave radiation emitted from the ground) and SR (absorbed solar radiations) are given by Thiers and Peuportier.

Most of Scotland is underlain by a crystalline basement, formed by weakly to strongly metamorphosed Lower Paleozoic sedimentary rocks. In the Northern Highlands and Grampian Highlands, the basement is characterized by a thick sequence of metamorphosed sandstone and mudstone units. Weakly metamorphosed sediments (i.e. sandstone and shale) disrupted by folding, faulting and tilting can be found in the Southern Uplands. Those are intersected by intrusions of basic ingenious rock, generally in the form of laccoliths, as well as granite intrusions. Highly metamorphosed granitic and basaltic intrusions seen at outcrop in the Northwest Highlands are likely to be similar to rocks found in the deepest part of Scotland.

The Midland Valley of Scotland (MVS) is a 80 km wide, 150 km long WSW-ENE trending Carboniferous graben that developed on the eroded and deformed remnant of the Caledonian Mountains (Cameron and Stephenson, 1985; Underhill, 1998; Read et al., 2002). It is separated from the Grampian Highlands by the Highland Boundary Fault, and from the Southern Uplands by the Southern Upland Fault. Both of those SW-NE striking fault formed major lineaments during the Caledonian Orogeny (Cameron and Stephenson, 1985). The MVS contains a complex arrangement of several Upper Palaeozoic sedimentary basins and Lower Palaeozoic metamorphic rocks (Cameron and Stephenson, 1985; Trewin et al., 2002; Underhill et al., 2007). It only has a few small granite intrusions at outcrop but contains numerous minor intrusions, mainly of basaltic and andesitic composition. The metamorphic basement only outcrop along the main faults together with the Devonian Old Red Sandstone sedimentary rocks. According to geophysical data, the basement-cover boundary is situated at about 8 km depth in the central part of the MVS.

Magmatic activity in the MVS extended from the Carboniferous to Permian for a period of about 90 Ma (Upton et al. 2004). It is suggested that volcanism led to an increase in the heat flow during the Carboniferous, especially in the active eastern region. Early magmatic activity occurred in the Pentland Hills during the Lower Devonian, in the form of lava flows and ash layers of basaltic, andesitic, trachytic and rhyolitic composition. During the Visean (Lower Carboniferous), basaltic to rhyolitic lavas and intrusions were formed from short-lived volcanoes that erupted in flood plain environment, releasing a vast amount of tuff materials (e.g. Garleton Hills, Arthur’s Seat, Clyde Plateau Volcanic Fm). Volcanic rocks from the Arthur's Seat Volcanic Formation in West Lothian, Edinburgh, and at the southern end of the Midlothian Syncline mostly consist of ankaramite to mugearite lavas, tuffs and volcaniclastic sedimentary rocks (Monaghan et al. XXX). Later Visean and Namurian activity (e.g. Kinghorn and Bathgate Hills volcanic Fms) was dominated by basaltic lavas and tuffs (Smedley 1986, Stephenson et al. 2003, Upton et al., 2004). In the late Carboniferous, short-lived episodes of tholeiitic magmatism extended across the MVS (Timmerman 2004, Monaghan and Parrish 2006). Those moslty occured in the form of dolerite intrusive sills (e.g. Salisbury Crags) and ENE-to ESE-trending tholeiitic dykes emplaced along extensional faults, cross-cutting the shortened and inverted Carboniferous basins (Cameron & Stephenson 1985, Rippon et al. 1996). From the Latest Carboniferous to Permian, alkaline magmatism (i.e. Mauchlin Volcanic Formation) occurred in response to the broader post-orogenic extension associated to the opening of the Altantic (Neumann et al. 2004; Wallis 1989), accompanied by the intrusion of NNE- to NW-trending dykes (Anderson et al. 1995, Upton et al. 2004). The more widespread thermal effect around tholeiitic sills/dykes compared to alkaline ones has been suggested by Murchison & Raymond (1989), as the result of their emplacement in dry and compacted sediments. Those likely exerted a major influence on hydrocarbons maturity within the lower Limestone Coal, Lower Limestone and upper West Lothian Oil-Shale formations, in the Central Coalfield and Midlothian-Leven basins (Monaghan and Brown, 2014).

Table XXX summarizes the proportion of different lithologies (i.e. sandstone, siltstone, limestone, coal, mudstone) for the Carboniferous formations of the Midland Valley of Scotland. Data are based on the analysis of geological cores and samples from the Glasgow Observatory (Geothermal Energy Research Field Site), published by Vincent et al. (XXX) and Entwisle (2019).

\begin{table}[h!]

\centering

\begin{tabular}{||l l l l l l||}

\hline

lithology & UCMS & MCMS & LCMS & LSC - PGP & BELOW \\

\hline\hline

Sandstone & 40 & \textgreater 40 & \textgreater 50 & 40 & 30 \\

\hline

Mudstone & 60 & 30 & 30 & 40 & 50 \\

\hline

Siltstone & & 24 & 14 & & \\

\hline

Coal & & 6 & 4 & & \\

\hline

Ironstone & & & \textless 1 & & \\

\hline

Limestone & & & & 20 & 20 \\

\hline

\end{tabular}

\caption{Percentage of lithologies recorded in existing borehole data. Data for the Lower (LCMS) and Middle (MCMS) Coal Measures are from the BGS Open Report OR19019 (Entwisle, 2019). Data for the Upper Coal Measures (UCMS), Limestone Coal (LSC) to Passage (PGP) Formations and underlying Carboniferous strata (BELOW) were determined from LOGS from 9 boreholes in the MVS (from Vincent et al., 2010) }

\label{table:1}

\end{table}

In mines in the Carboniferous succession of the MVS, the presence of many low conductivity "cap" rocks (i.e. coal, mudstone strata) in the sedimentary cover above the crystalline basement likely creates situations where the vertical upflow of heat is trapped at depth, generating positive heat anomalies. Such blanketting effects has been described by Busby et al. (2011) for the NE England, where low conductivity Carboniferous rocks are present above the Weardale granite, which has a high heat production (Downing and Gray, 1986a,b).

The average thermal conductivity for the formations of the Carboniferous succession in the MVS is shown in Table XXX together with values for common rock types reported by Lee et al. (1984) and Wheildon et al. (1984). 23 borehole measurements of rock thermal conductivity and mean thermal conductivities for specific lithologies for Scotland are also provided by Rollin (1987) and reported in Banks (2008). Browne et al. (1985) noted considerable variations both in the proportion of rock types in each formation in the MVS and in the conductivity for each lithology in the Upper Palaeozoic. Formations thermal conductivity were estimated by Busby (2019) in the UKGEOS Open Report OR19015 by combining the thermal conductivities of individual lithological units as a harmonic mean, based on the data recently acquired from cores at the Glasgow Geothermal Energy Research Field Site. Due to the absence of thermal conductivity values for the Scottish Coal Measures, the Passage Formation and the Kirkwood Formation, the thermal conductivity for representative sections in those formations were estimated from values measured on similar lithologies in 5 boreholes located in the Pennine Coal Measures of northern England.

\begin{table}[h!]

\centering

\begin{tabular}{||l l l||}

\hline\hline

a) Lithology & K (W/m/K) & Reference\\

\hline\hline

Coal & 0.31 & Lee et al., 1984\\

Coal Measures Sandstone & 3.31, 3.58\* & Rollin, 1987; \*Busby (2019) \\

Coal Measures Siltstone & 2.22, 2.23 & Rollin, 1987; \*Busby (2019)\\

Coal Measures Mudstone & 1.49, 1.85\* & Rollin, 1987; \*Busby (2019)\\

Coal Measures Ironstone & 2.35\* & \*Busby (2019)\\

Namurian Millstone Grit Limestone & 3.75 & Rollin, 1987\\

Lower Carboniferous limestone & 3.14 & Rollin, 1987\\

Upper Old Red Sandstone & 3.26 & Rollin, 1987\\

Silurian slates & 3.33 & Rollin, 1987\\

Hercynian granites & 3.3 & Rollin, 1987\\

Basalt & 1.8 & Rollin, 1987\\

Shale & 1.3 & Busby (2019)\*\* \\

Fireclay & 0.59 & Busby (2019)\*\* \\

\hline\hline

b) Formation & K (W/m/K) & Reference) \\

\hline\hline

Scottish Coal Measures Formation & 1.91 ± 0.25 & Browne et al., 1985 \\

Scottish Middle Coal Measures & 2.02 & Busby (2019)\\

Scottish Lower Coal Measures Formation & 1.91 ± 0.25 & Busby (2019)\\

Passage Formation & 2.91 ± 0.15 & Browne et al., 1985\\

Upper Limestone Formation & 2.25 & Busby (2019)\\

Limestone Coal Formation & 2.24 & Busby (2019)\\

Lower Carboniferous & 2.12±0.25 & Browne et al., 1985 \\

Lower limestone Formation & 1.88 & Busby (2019)\\

Lawmuir formation & 4.36 & Busby (2019)\\

Kirkwood Fm & 2.1 & Busby (2019)\\

Clyde Plateau Volcanic Formation & 2.2 & Busby (2019)\\

Clyde Sandstone Formation & 4.19 & Busby (2019)\\

Ballagan Formation & 3.14 & Busby (2019)\\

Kinnesswood formation & 3.66 & Busby (2019)\\

\hline\hline

\end{tabular}

\caption{a) Thermal conductivity for specific lithologies found in the MVS succession, estimated based on laboratory measurements made on water saturated samples extracted from boreholes. \*representative values for the Scottish Middle Coal Measured estimated from 5 boreholes in England. \*\* averages from the Boreland (Anderson, 1940) and Glenrothes (Gebski et al., 1987) boreholes. b) Average thermal conductivity for the formations of the Carboniferous succession of the MVS (from Busby, 2019), estimated from the Maryhill, Hurlet House, Clachie Bridge, Barnhill, Kipperoch (Oxburgh, 1982), Boreland (Anderson, 1940) and Glenrothes (Gebski et al., 1987) borehole logs}

\label{table:2}

\end{table}

\subsection{Hydrology}

Carboniferous sedimentary sequences in the Midlothian Coalfield are composed of low permeability layers (mudstone) interbedded with higher permeability strata (sandstone, limestone) that tend to form complex multi-layered aquifers, both in confined or unconfined conditions. Those aquifers are generally described as minor or moderately productive aquifers due to a relatively low rock permeability, providing borehole yields in the range 5-15 l/s (MacDonald et al., 2004; Robins, 1990; Ball, 1999). The natural layering tends to create complex flow path, dominated by horizontal inter-granular flow and fracture flow. In those formations, groundwater residence time has been estimated to be in excess of 60 years (Ó Dochartaigh et al., 2011). In contrasts, the Passage Formation forms an extensive productive aquifer where groundwater moves dominantly by intergranular flow (Ball, 1999). Lateral compartmentalization also results from the intrusion of igneous rock (dykes, sills, plugs) or faults that can act either as permeable pathways or barriers to groundwater flow.

Despite the presence of coal seams is adding more low-permeability layers between the sandstone aquifer units, the Carboniferous strata extensively mined for coal generally have slightly higher productivity than the aquifers not extensively mined for coal (Dochartaigh et al., 2015). Mining activities indeed resulted in an increase in permeability of the Carboniferous strata both by increasing the void space in zones where seams were mined out and through fracturing of competent horizons above the mined seams. Mining has tend to form “anthropogenically enhanced aquifers” (Banks, 1997) with greater transmissivity than the undisturbed aquifers and locally increased storage capacity. Collapse of roofs above large voids and deformation of the surrounding rock mass might however lead to the opposite effect by reducing both transmissivity and storage (Younger and Robins, 2002). However, only a few aquifer properties data from pumping test are available for boreholes intercepting former mines. Borehole yields used to dewater mines in Scotland during mining activities are shown in table XXX.

A total of 61 topography corrected temperatures acquires from onshore boreholes are reported in Burley et al. (1984), and 17 additional borehole temperature-depth measurements located in the west Midland Valley are reported in the 3rd version of the Catalogue (). Despite most of the measured temperatures are assumed to represent equilibrium temperature, i.e. little affected by cooling due to circulating drilling mud or by mine ventilation systems, Farr et al. (2016) warned that some uncertainties still limit the reliability of geothermal gradient estimates. This includes the lack of knowledge of the depth and timing of temperature measurement after the drilling of the boreholes, the effect of circulation of water within the borehole and the influence of intercepted mine working. Browne et al. (1985) suggested that equilibrium temperatures obtained during heat flow measurements are the most reliable and accurate values, however, they are relatively rare in the MVS.

Knowing the long term temperature variation does matter to know the actual deep heat flux and quasi steady state geothermal gradient. for shallow extraction (< 20 m), daily / annual fluctuations might be of greater interest as it is mostly impacted by daily fluctuations.

Due to the narrow spread of the data, this suggests that the estimated average geothermal gradient might be consistent across Scotland and characterise a regional background heat flow, unperturbed across the basement-cover interface. It is however suggested that the proximity of offshore area or the stretched crust in the MVS, where the crust may be thinner and thus the heat flow higher, may account for a higher geothermal gradient in the deepest part of the curve. This might also be explained by more radioactive granitic rocks within the crust, or continued heat flow from Cenozoic igneous activity (Browne et al., 1987). Other theories have suggested that the heat flow anomaly in the MVS can be explained by a regional upflow of groundwater (Browne et al., 1987; Lee et al., 1987; Robins, 1988), with areas of recharge (i.e. downward flow) mainly located along the northern and southern boundaries of the MVS, where the elevation is higher (Robins, 1988). However, this model has been refuted by Robins et al. (1988) and Browne et al. (1987) based on the lack of evidence from geochemical data and on the fact that deep groundwater circulation is likely to be moderate in volume and confined within isolated discrete pathways.

The thermal character of a reservoir rock is mainly dependent on the effective heat capacity Cp (J/K), thermal conductivity $\lambda$ (W/m.K) and thermal diffusivity $\alpha$ (m2/s) of the materials that forms it, as well as the amount of saturation. Where conductive processes are dominant, the effective fluid-rock heat capacity and heat conductivity both influence heat transfers. Thermal dispersivity mostly account for heat dispersion in areas where convective heat transfers are dominant (Deethlefsen et al., 2016). However, while the specific heat capacity, conductivity and diffusivity of groundwater are well known, the thermal properties of rocks highly depend on their material composition, texture and structure (Allen et al., XXX; Deethlefsen et al., 2016). Studies driven by Allen et al. (XXX) showed that the increase in rock specific heat capacity with increasing temperature was notable for temperature closed to 0°C. Thermal conductivity of most rock types (i.e. quartzitic sandstone, slate, limestone, marble, calcite, gneiss and granite) tends to reduce with increasing temperature, but the formation of cracks rocks does not systematically lead to a decrease of thermal conductivity. Thermal diffusivity appears to be highly dependent on quartz and feldspar content. Quartz-rich rocks (i.e. sandstone or quartzite) tend to have a high diffusivity, while those with a high feldspar content (i.e. limestone, marble) have a lower diffusivity, that was explained by the poor conductive properties of feldspar. For a similar composition, crystalline rocks moreover tend to have higher thermal diffusivities compared to sedimentary rocks Allen et al. (XXX).

**Low-conductivity horizontal rock layers such as mudstone or coal are finally likely to form cap rocks for the inflow of geothermal heat. If heat is generated from an underlying heat source, this might contribute to the formation of thermal anomalies beneath thick layers of coal-bearing formations.**

\section{Mine-water temperature}

Minewater temperatures reported in the BGS Catalogue of Geothermal Data for the UK (Burley et al., 1984) have been analyzed by (Gillespie et al., 2013). Temperatures measured from nine boreholes in the Midland Valley range from 12 to 21°C, with a mean and median of 17 °C. However, despite the temperature of mine waters generally increases with depth according to the geothermal gradient, no correlation has been found between the depth of the measurements and the observed water temperature in the Midland Valley. Robin (1990) reported a range of temperature comprised between 8 and 15 °C for water circulating above 200 m. In autumn 2008, new groundwater samples were collected from boreholes abstracting water from Carboniferous sedimentary aquifers across the Midland Valley, as part of the Scotland Baseline Project. Four of them originate from abandoned flooded mine workings (i.e. either pumped or from gravity-driven discharge zones). Temperatures of two of the measured pumped mine-water were 11.7 and 14.5 °C, in accordance with the typical temperature of natural groundwater from Carboniferous aquifers in the Midland Valley. Temperature of the sampled mine-water pumped from the Polkemmet mine shaft was 19.2 °C, while the measured temperature of the gravity-driven discharged mine-water was 9.8°C. Average mine water temperatures of 11.5-13.3 °C and 14-15°C were moreover reported for the Crynant, Dulais Valley, South Wales Coalfield (Farr et al., 2016) and Caphouse Colliery (Burnside et al., 2019), respectively.

In their study, Farr et al. (2016) assessed the low enthalpy heat recovery potential from coal mine discharges in the South Wales Coalfield based on water chemistry analysis. Despite relatively constant temperature measured at each of the 16 sites during the monitoring period, temperature differences between each of them were attributed to the position of the logger relative to the surface features. By comparing the mine water temperature to the average monthly air temperature, Farr et al. (2016) showed that measured temperature can be influenced by 1) cold recharge entering the drainage system (i.e. shallow logger in gravity driven adit, correlated with air temperature), 2) localized recharge events such as period of intense rainfall (i.e. adits with transfer pumps with more stable yearly temperature) or 3) the absence of connection with surface recharge (i.e. deep warm water with constant temperature throughout the year). Groundwater of different temperature may indeed enter the borehole at different depths, exchanging heat with the surrounding rocks to a degree that will depend on the flow rate, making it difficult to predict the temperature of the pumped water.

Additional temperature data have been collected by the Coal Authority since 1994 (TO DEVELOP). Those includes temperature time series, acquired from data logger placed closed to the surface in shafts. Shafts are generally capped, and can sometimes be filled by collapsed materials from the mines. Temperature profiles in boreholes are moreover punctually acquired during surveys. Table XXX summarizes the data available for the Bilston Glen - Easthouses area, located in the Midlothian Coalfield. \\

\begin{figure}[htp]

\centering

\includegraphics[width=8cm]{Mine\_T.png}

\caption{Temperature of mine water measured in shafts}

\label{fig:Mine\_T}

\end{figure}