A study of compaction and cementation of sandstones in the Athabasca Basin, Northern Saskatchewan, Canada, and implications for unconformity-related uranium mineralization

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Yumeng Wang, candidate for the degree of Master of Science in Geology, has presented a thesis titled, *A Study of Compaction and Cementation of Sandstones in the Athabasca Basin, Northern Saskatchewan, Canada, and Implications for Unconformity-related Uranium Mineralization*, in an oral examination held on November 13, 2019. The following committee members have found the thesis acceptable in form and content, and that the candidate demonstrated satisfactory knowledge of the subject material.

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Abstract

The Athabasca Basin hosts numerous high-grade unconformity-related uranium (URU) deposits, which typically occur near the unconformity between the metamorphosed basement and the quartzose sedimentary rocks. Previous studies have generally agreed upon a diagenetic-hydrothermal model in which the URU mineralization occurred at elevated temperature (>200°C) and deep-burial (5-7 km) conditions under normal geothermal conditions. However, a recent study invoked that the URU mineralization may have occurred at a relatively shallow burial depth (<3 km) and alternatively interpreted elevated temperatures (>200°C) throughout the basin as the result of higher-than-normal thermal gradient, which may be associated with deep-seated geodynamic processes. In consideration of this debate, this study investigated the compaction and cementation characteristics of the sandstones from four drill cores (Rumpel Lake, WC79-01, BL-08-01, DV10-001) located in the central part of the basin. Samples were collected from different stratigraphic levels, including the Read, Manitou Falls, Lazenby Lake, Wolverine Point, and Locker Lake formations. Three petrographic parameters, including contact index (CI), tight packing index (TPI) and intergranular volume (IGV), were used as indicators of the degree of compaction. Only moderately well- to well-sorted and matrix-poor quartz arenite samples were selected for point
counting. It was found that below the mud-rich Wolverine Point Formation, the degree of compaction increases downward sharply from the Lazenby Lake Formation to the Read Formation, whereas the degree of quartz cementation decreases downward. However, high-degree compaction was also observed in the Locker Lake Formation above the Wolverine Point Formation. The dominance of point contacts and high abundance of quartz cement (up to 23%) in the Lazenby Lake Formation indicate a shallow burial condition at the time of quartz cementation. Reactive mass transport modeling results by using the TOUGHREACT program suggest that the petrographically discerned compaction and cementation pattern was controlled by fluid convection within sandstone successions that were confined by low-permeability aquitards. Combined with previous fluid inclusions studies indicating that the diagenetic fluid is relatively high-temperature (generally>120°C), high-salinity brine with high uranium concentration, this study further provides petrographic evidence supporting the notion that the shallow (<3 km) burial model of URU mineralization may be reasonable and large-scale fluid convection may have occurred at a time when the sedimentary rocks were not subjected to significant compaction.
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Chapter 1: Introduction

1.1 Rationale of this study

The Athabasca Basin, located in Northern Saskatchewan and Alberta, is a uraniferous basin hosting numerous world-class, high-grade (generally >1%, up to ~20%), and large-tonnage (up to ~200,000 t U) uranium deposits (Jefferson et al., 2007). The uranium deposits typically occur near the unconformity between the overlying quartz-dominated Athabasca Group and the underlying metamorphosed crystalline basement (Jefferson et al., 2007). The deposits are generally considered to have formed from the redox reaction between basinal, oxidized, uraniferous brines and reduced basement fluids and/or lithologies under diagenetic-hydrothermal conditions (Pagel, 1975; Hoeve and Sibbald, 1978; Pagel et al., 1980; Hoeve and Quirt, 1984; Kotzer and Kyser, 1995; Derome et al., 2005; Richard et al., 2011; Mercadier et al., 2012).

Pagel (1975) performed fluid inclusion P-T-X analyses on fluid inclusions in quartz overgrowths in the sandstones that were collected from the Rumple Lake drill core and the Cluff Lake deposit and his results suggested that the temperature at the basal part of the basin could reach up to 220°C. Besides, he further estimated that the fluid pressure
ranges from 900 bars to 1500 bars, which corresponds to a burial depth of 5–7 km if assuming a fluid pressure regime under lithostatic conditions and a rock density of 2.65 g/cm³, and finally a geothermal gradient of about 35°C/km was estimated. Subsequent fluid inclusion microthermometric and clay thermometric studies suggested that the Athabasca Basin may have been subjected to temperatures ranging from 180°C to 250°C extensively, both in mineralized places and barren areas (Hoeve and Quirt, 1984; Kotzer and Kyser, 1995; Derome et al., 2005; Cloutier et al., 2009; Richard et al., 2016; Chu and Chi, 2016). These studies were generally taken to support the viewpoint that the unconformity-related uranium deposits were formed at a deep burial depth (5–7 km) at a normal geothermal gradient (Hoeve and Quirt, 1984; Kotzer and Kyser, 1995; Derome et al., 2005; Cloutier et al., 2009; Richard et al., 2016). However, the results of some other studies seem to be contradictory with Pagel’s model. Wilson et al. (2007) reported that the Douglas Formation at the top of the Athabasca Group, which is about 2.5 km above the unconformity, may also have undergone temperatures ranging from 160 to 200°C, based on analyses of vitrinite reflectance of organic matter. The results contradict with the inference of the geothermal gradient of 35°C/km by Pagel (1975). Besides, Chu and Chi (2016) reported a new series of temperature data based on fluid inclusions and clay microthermometry and the results suggested that the elevated fluid temperatures ranging from 160 to 200°C were extensively developed throughout the stratigraphy, which is
contradictory with the systematic thermal gradient proposed by Pagel (1975).

Chi et al (2015) further argued that there may have been some uncertainties in the estimation of the burial depth based on the fluid pressure interpreted from the fluid inclusion studies by Pagel (1975). Firstly, fluid overpressure was unlikely to develop in the Athabasca Basin which is a high-permeability, quartz-dominated basin and is favorable for fluid flow and thus the releasing of pore water pressure, and therefore the fluid pressure regime was more likely to be near-hydrostatic (Chi et al., 2013); this is contradictory with the assumption of Pagel (1975) that fluid regime of the Athabasca Basin is lithostatic. Chi et al (2014) further suggested that the fluid pressure may have been at a near-hydrostatic condition no matter whether hydrocarbon generation in the organic-rich Douglas Formation was considered or not. Therefore, if assuming the fluid pressure estimated by Pagel (1975) was correct and fluid regime was actually at a near-hydrostatic condition, a maximum burial depth would be calculated to be about 15 km, which would result in an unreasonably low geothermal gradient of about 13.3°C/km (Chi et al., 2015).

In consideration of the uncertainties about the burial depth of the basin at the time of uranium mineralization, Chi et al (2018) advocated a revised genetic model in contrast to
the widely-accepted, conventional diagenetic-hydrothermal mineralization model, invoking that the primary uranium mineralization in the Athabasca Basin likely occurred at a shallow burial depth (<3 km) based on a regional geochrono-stratigraphic research. In opposition to the conventional deep-burial model in which uranium mineralization occurred at a deep burial depth of approximately 5 to 7 km, the relatively-shallow burial mineralization model implies that, at the time of URU mineralization, the Athabasca basin may have undergone a relatively low degree of compaction, which results in sedimentary rocks having high porosity and permeability at that time. Besides, the new relatively-shallow burial model also implies that the URU mineralization may be associated with a large-scale thermal disturbance and its resultant anomalously high geothermal gradients. This study aims to constrain the burial depth of the Athabasca basin by studying the compaction and cementation characteristics of sandstones, as the compaction degree of sandstones is typically deemed as an indicator of the maximum burial depths that the rocks have been subjected to. Furthermore, the maximum burial depth of the basin must be greater than or equal to the burial depth of the basin at the time of uranium mineralization. Therefore, this study aims to constrain the mineralization depth by studying the compaction characteristics of the sandstones and estimating the maximum burial depth of the basin.
1.2 Literature review of compaction and cementation of sandstones

Compaction and cementation are two of the most important diagenetic processes of sedimentary rocks. Compaction and cementation turn loose sediments into consolidated rocks during the burial history. Sandstone is a type of sedimentary rock composed of sand-size grains of minerals called detrital grains, with variable amounts of cements and matrix of silt- or clay-size particles between sand grains. Sandstones typically present different compaction characteristics at different burial depths as the result of the loading pressure exerted by overlying sediments, and in turn compaction characteristics can be used to evaluate the maximum burial depths that rocks have undergone.

Compaction can be roughly divided into mechanical compaction and chemical compaction. Mechanical compaction is controlled by the effective stress that framework grains take on and dominates at a relatively shallow burial depth down to 2 to 4 km depth, which roughly equals to 80–100°C depending on geothermal gradients (Fig. 1.1; Szabo and Paxton, 1991; Bjørlykke, 1997; Bjørkum et al., 1998, Bjørlykke and Jahren, 2010). During this stage, grain reorientation, pore water expelling, fragile grains breakage and plastic grains deformation occur, resulting in porosity diminishing, bulk volume decreasing and density increasing. However, in addition to burial depth, other factors
including mineralogical composition, clay content, fragile grains content and sorting, as well as the development of overpressure, may also influence the response of bulk rocks to the loading pressure during the mechanical compaction stage. Specifically, relative to rigid detrital grains (quartz and feldspar) that can resist strong mechanical compaction, clayey and fragile grains (mica and altered lithic fragments) that are ductile and soft materials, are more easily squeezed, deformed and crushed. If compaction is evaluated based on porosity, poorly-sorted sandstones tend to be more compacted than well-sorted sandstones at the same burial depth because relatively small grains tend to be squeezed into the interstitial pore space, which significantly reduces primary pore space. Other factors that may affect the response of sediments to compaction is overpressure. Overpressure refers to a state where porewater pressure in porous sediments is much higher than hydrostatic pressure (Osborne and Swarbrick, 1997). In overpressured sediments, porewater shares the loading pressure with framework grains, aiding sedimentary rocks to retard both the mechanical and chemical compaction, and therefore it is beneficial to the preservation of primary intergranular porosity and the development of high-quality reservoirs (Stricker and Joines, 2018). Chemical compaction refers to dissolution at grain contacts and precipitation at nearby open pore, which is regarded to be controlled by both the effective pressure and the relatively-high temperature (>90−100°C) at a relatively deep burial depth (Bjørlykke and Kaare, 1997; Bjørlykke and
Jahren, 2010). Chemical compaction can be observed at grain contacts between different types of detrital components including detrital quartz, feldspar and lithic fragments. Detrital quartz that is more thermodynamically stable than feldspar and lithic fragments can stay chemically stable within the first 2 km burial depth, which approximately corresponds to temperatures ranging from 80 to 90 °C (Bjørlykke and Jahren, 2010). In some literature, chemical compaction has been interpreted to be controlled by effective pressure at grain-grain contacts and was therefore named as ‘pressure solution’ in the past (Houseknecht, 1984, 1987, 1988). However, more recent studies advertised that the most important factor controlling the kinetics of pressure solution is temperature and geological time rather than effective pressure, and alternatively use the term of ‘chemical compaction’ in place of ‘pressure solution’ (Fig. 1.1; Bjørlykke and Kaare, 1997, 2014; Bjørkum et al., 1998; Sheldon et al., 2003). Besides, some studies reported that pressure solution can be induced by clay minerals at a relatively high temperature even when effective pressure is relatively low, which challenges the importance of effective pressure on controlling the kinetics of chemical compaction (Heald, 1974; Weyl, 1959; Bjørkum, 1996). Whereas, Stricker et al (2016) reported that the development of overpressure is an indispensable factor in controlling the chemical compaction as supported by the observation that highly-overpressured sandstones are not significantly compacted compared to the counterparts within hydrostatic pressure regime. Chemical compaction
happening at grain contacts can be called intergranular pressure solution (IPS), which can also develop more extensively and form a planar-shaped, laterally-extensive and interdigitated zone called stylolites. IPS and stylolites are regarded to have the same formation mechanism, and the only difference is the scale (Tada and Siever, 1989). Stylolites in sandstones are generally filled by insoluble materials, e.g. clay, iron-oxide and heavy minerals. The occurrence of stylolites can be used to indicate the direction of maximum principal stress (Tada and Siever, 1989). Stylolites that are parallel to sedimentary beddings are interpreted to form as the result of progressive burial, while those obliquely cutting through the beddings are interpreted to develop due to lateral tectonic compression (Railbeck and Andrew, 1995). Most researchers agreed upon that stylolites typically develop at a relatively deep (at least >1.5 km) burial depth in quartzose successions, but the temperature and pressure threshold for the development of stylolites are still uncertain (Tada and Siever, 1989). The compaction characteristics of sandstones are affected by cements, especially if cements are composed of rigid minerals, e.g. syntaxial quartz overgrowths and calcite cements. Detrital grains that were cemented earlier would be protected from subsequent compaction due to the consolidation effect (Fig. 1.2; Bjørlykke and Jahren, 2010).
Fig. 1.1 Sandstone diagenesis process as a function of temperature and time. Note that chemical compaction and quartz cementation start at a temperature of 80–90°C, and continue until basin uplifting which decreases burial temperature to below the required temperature of chemical compaction (From Bjørlykke and Jahren, 2010)
Fig. 1.2. Porosity evolution of sandstones due to compaction. At a burial depth <2 km, sandstones compact mechanically as a function of effective stress (depth), during which time grain reorientation, pore water expelling, plastic grains deformation and grains breakage occur. At a deeper burial depth >2 km with temperature increasing up to about 80–100°C, chemical compaction and quartz cementation start to occur. It can be assumed that the chemical compaction degree of the sandstones is a function of depth, and thus also a function of hydrostatic pore pressure (hydrostatic pore pressure equals to the product of depth and hydrostatic pressure gradient). However if sandstones are cemented by quartz overgrowths, they become overconsolidated relative to strain level, and thus further compaction requires an extra strain $X_2$ to overcome cementation consolidation. If overpressure develops, the effective pressure that grain frameworks take on decreases due to the supporting of pore water to detrital grains, and in this case, further compaction requires an extra strain $X_1$ to overcome overpressure consolidation (From Bjørlykke and Jahren, 2010).
Though some case studies report quartz cementation can take place at a very shallow burial depth, and can be related to specific biological progress and surficial hydrolysis of unstable minerals (e.g. volcanic glass) during chemical weathering (Milliken, 1979; McBride, 1989; Aplin and Warren, 1994; Aase et al., 1996), most researchers proposed that quartz cement can only become a major pore-occluding phase at a relatively deep (>2 km) burial condition (McBride, 1989; Bjørlykke and Jahren, 2010). In deeply buried conditions of relatively high temperature, some minerals are dissolved, transformed and replaced, which is followed by the precipitation of authigenic minerals within intergranular space. Authigenic quartz is the most common pore-filling cement of quartz-rich sandstones that have been subjected to deep (>2500 m) burial depth because quartz solubility is extremely low at low temperature (<60 °C) but has a positive relationship with temperature, and thus can only become a major authigenic pore-occluding phase in a relatively-deeply-buried sandstones that are heated to above 80°C (Fig. 1.2; Bjørlykke and Jahren, 2010). The temperature of 80°C is regarded as the threshold temperature for quartz cementation (Fig. 1.2; Bjørlykke and Jahren, 2010). However, some studies advocated that authigenic quartz cementation can also happen in a kinetically controlled process at a low temperature (<80°C) environment with small amounts of quartz cement precipitating at a very slow rate (Worden and Morad, 2000). On the other hand, clay coatings can impede quartz cementation by covering clean quartz
surface which is necessary for the precipitation of quartz overgrowth (Worden and Morad, 2000). It is generally agreed that the development of quartz cementation involves three steps, including extraction of silica from the source, the transportation of silica from the site of source to the site of cementation, and precipitation (Worden and Morad, 2000). The silica for quartz cementation may be derived from intergranular pressure solution, stylolitization, feldspar leaching and clay minerals transformation (McBride, 1989; Worden and Morad, 2000). Many researchers believe that silica transportation in porous sandstones is mainly achieved by molecule diffusion rather than advection (Bjørlykke et al., 1988; Bjørlykke and Gran, 1994), whereas some researchers proposed that large-scale fluid convection may have fundamentally controlled silica transportation in sedimentary basins (Wood and Hewett, 1982). The precipitation of quartz is generally assumed to be a continuous process rather than episodic (Walderhaug 1996; Oelkers et al. 2000), which has been recently demonstrated by high-precision in-situ isotopic techniques (Harwood et al. 2013). Based on the continuous quartz precipitation hypothesis, the kinetics of quartz precipitation is approximately calculated based on a series of studies involving petrography (Blanche and Whitaker 1978; Oelkers et al., 1996), fluid inclusion studies (Haszeldine et al. 1984; Walderhaug 1994a, 1994b), isotopic geochemistry (Harwood et al. 2013) and basin burial history analysis. Thus, a kinetically-controlled quartz cementation model has been gradually accepted and shown to be effective in predicting
quartz cementation with the consideration of the depth, grain surface area, temperature and time (Walderhaug 1996, 2000; Lander and Walderhaug, 1999; Lander et al. 2008; Harwood et al. 2013; Schmoker and Gautier, 1988, 1989). This kinetically-controlled geological model forms the basis for the simulation of quartz cementation in sandstones and is commonly used to predict porosity evolution of petroleum reservoirs.
1.3 Objectives of study

The main purpose of this study is to evaluate the burial depth of the unconformity between the basin and basement during the URU mineralization through the study of the compaction and cementation characteristics of the sandstone. The specific objectives can be divided into the following three parts: 1) to characterize the compaction and cementation pattern of sandstones with respect to burial depth; 2) to determine the controlling factors of the compaction and cementation patterns of sandstones; and 3) to understand the linkage between this study and the burial depth of the Athabasca basin during the URU mineralization.
Chapter 2: Geological Setting

2.1 Regional geology

Located in the northwestern part of the Canadian Shield, the basement of the Athabasca Basin belongs to the Taltson magmatic zone, Rae Subprovince and Hearn Subprovince from west to east (Fig. 1a; Jefferson et al., 2007). The Rae and Hearn Subprovinces are separated by the northeast-trending Snowbird tectonic zone (Hoffman, 1990). The 1.99–1.92 Ga granitoid intrusions, occurring in the Rae Subprovince, are interpreted to be associated with Thelon-Talston orogeny (Card et al., 2007), while the 1.9–1.8 Ga intrusions that were emplaced in the Hearn Province are interpreted to be related with the Trans-Hudson orogeny (Card et al., 2007).

The Athabasca Basin is generally considered to be an intracratonic basin based on the following characteristics (Jefferson et al., 2007): 1) a regular, saucer-shaped geometry, 2) a flat-lying stratigraphic configuration without boundary faulting systems, and 3) infills are predominantly continental sediments that are alluvial-fluvial facies, with minor volumes of marine sands, shales, and carbonates. However, Ramaekers et al (2017) proposed a more complex basin model in which the deposition of the Jackfish and Cree
subbasins was contemporary with the Trans-Hudson orogeny, and therefore the Athabasca Basin was a foreland basin at the time. Later on, with the breakup of Nuna supercontinent, the basin became a part of a rifting system, during which time the mirror subbasin developed. According to the dogmatic definition, a foreland basin subsides and accepts sediments during the orogeny. However, based on the geochronostratigraphic dating, the deposition time (<1.75 Ga) of the Athabasca Basin is much later than the active stage of the Trans-Hudson orogeny (1.9 Ga–1.8 Ga) (Alenxandre et al., 2009; Card et al., 2007). Therefore, it is problematic to consider the Athabasca Basin as a foreland basin. Nevertheless, the Athabasca Basin does have some features of foreland basins as suggested from the following sedimentary characteristics: 1) the orientation of sedimentary systems is from east to west as indicated by the directions of paleocurrents systems, which suggests that the sediments were mainly derived from the Trans-Hudson orogeny to the east of the Athabasca Basin; 2) sedimentary troughs controlled by reactivated Trans-Hudson faulting systems suggest that the Trans-Hudson orogeny might be reactivated during the deposition of the Athabasca Group (Jefferson et al., 2007).

As for the hypothesis that the Mirror subbasin is a rift basin (Ramaekers et al., 2017), there has been no unequivocal evidence (e.g. syn-sedimentary normal faulting systems) discovered yet. The currently known extensional structures related to the Mackenzie
diabase dikes that were emplaced in the Athabasca Basin at ca.1267 Ma (LeCheminant and Heaman, 1989) is significantly younger than the preserved strata of the Athabasca Basin. Therefore, this study deems the Athabasca Basin as a post-orogenic intracratonic basin, which is intimately related to the Trans-Hudson orogeny in terms of sediments source and tectonic reactivation.
Fig. 2.1. (a) Location of the Athabasca Basin in Canada; (b) Geological map of the Athabasca Basin showing each formation and the location of the studied four drill holes: WC-79-1, Rumple Lake, BL-08-01, DV10-001 (modified from Ramaekers et al., 2007; Jefferson et al., 2007; Bosman et al., 2011). B – basement; FP – Fair Point; S – Smart/Manitou Falls; RD – Read; MF – Manitou Falls; LZ – Lazenby Lake; W – Wolverine Point; LL – Locker Lake; O – Otherside; D – Douglas; C – Carswell.
2.2 Stratigraphy

There are currently two classification schemes for dividing the sedimentary infills of the Athabasca Basin. The nomenclature used in this study is a classical one which incorporates the sediments in the Athabasca Basin as the Athabasca Group. The Athabasca Group is mainly composed of fluvial to shallow marine facies quartzose sandstones and is divided into four depositional sequences according to the basin-wide unconformities (Fig. 2.2). Sequence 1 (Fair Point Formation) is a fluvial quartzose succession composed of conglomerate, conglomeratic quartz arenite and coarse-, medium-grained quartz arenite with a few pebbly mudstone clasts. Sequence 2 (Read/Smart and Manitou Falls formations) is composed of several fluvial quartzose successions including conglomerate, conglomeratic quartz arenite, pebbly quartz arenite, coarse-/medium-/fine-grained quartz arenite, quartz wacke and some clay intraclasts and siltstone interlayers. Sequence 3 (Lazenby Lake and Wolverine Point formations) is a shallow-marine and relatively mud-rich sequence. The Lazenby Lake Formation consists of quartz arenite and mud-rich quartz wacke, and the Wolverine Point Formation is composed of mud-rich quartz wacke, mudstones and quartz arenite. Sequence 4 (Locker Lake, Otherside, Douglas, and Carswell formations) is composed of fluvial successions and marine facies sediments. The Locker Lake and Otherside formations are composed of
fluvial quartz arenite, whereas the Douglas Formation is composed of quartz wacke, quartz arenite, and organic-rich mudstone. The Carswell Formation is composed of dolomite and stromatolite. It should be noted that the Douglas and Carswell formations are only preserved around the Carswell Structure, which is a meteorite impact structure formed in the Ordovician (Fig. 2.1b; Ramaekers et al., 2007).

Sedimentary rocks of the Athabasca Group can be globally divided into four packages of hydrological units. The Douglas and Carswell formations that are mainly composed of organic-rich mudstones and carbonates can be regarded as a low-permeability aquitard at the top. The middle unit of the Locker Lake and Otherside formations that are mainly composed of fluvial sandstone can be regarded as a high-permeability aquifer. The Wolverine Point Formation that is mainly composed of marine mudstone and mud-rich sandstone can be regarded as a low-permeability aquitard in the middle. The bottom unit of the Manitou Falls and Read formations that are dominantly composed of alluvial/fluvial sandstones can be regarded as a high-permeability aquitard above the low-permeability crystalline basement. This sandwiches-like hydrological geometry was deemed as the controlling factor of the formation water flow in this study.

Based on the isopaches of the four sequences, the Athabasca Basin is divided into three
subbasins: Jackfish subbasin in the west, Mirror subbasin in the mid-west and Cree subbasin in the east. The Jackfish subbasin is the place where the Sequence 1 (Fair Point Formation) was deposited, and the Cree subbasin is the place where sequence 2 has the largest thickness. The Mirror subbasin is characterized by the largest thickness of sequence 3 and 4. The subbasins arrangement indicates the migration of the depocenter of the Athabasca Basin. The depocenter moved from the Jackfish Subbasin to the Cree Subbasin from the deposition of sequence 1 to the sequence 2, and then moved to the Mirror subbasin during the deposition of sequence 3. Geochronological data indicate that the deposition of the Athabasca Group may have started after the Trans-Hudson Orogeny at about ca. 1750 Ma, which may represent the maximum age of the Fair Point Formation (Annesley et al., 1997; Alexandre et al., 2009). The age of the Wolverine Point Formation was reported to be 1644 ± 13Ma (Rainbird et al., 2007), and that of the Douglas Formation was estimated to be 1541 ± 13 Ma (Creaser and Stasiuk, 2007).

Recently, a revised stratigraphic classification scheme has been presented raising the Athabasca Group to the Athabasca Supergroup (Bosman and Ramaekers, 2015). The Athabasca Supergroup is composed of four stacked sedimentary basins that are the Martin, Jackfish, Cree and Mirror basins. The Fair Point, Manitou Falls, Lazenby Lake and Wolverine Point formations were raised to group status. While the Carswell, Douglas,
Locker Lake formations were grouped together as the McFarlane Group.

However, the classical nomenclature is still in use in this study for ease to compare the results with previous studies. This is not a problem considering the nature of this study, i.e., a diagenetic study (with a focus on compaction and cementation) rather than a sedimentological and stratigraphic study.
### Fig. 2.2. Lithology and stratigraphy of the Athabasca Basin (based on Ramaekers et al., 2007)

<table>
<thead>
<tr>
<th>Age</th>
<th>Formation</th>
<th>Jackfish Subbasin</th>
<th>Mirror Subbasin</th>
<th>Cree Subbasin</th>
<th>Sequence</th>
<th>Petrographic Description</th>
<th>Sedimentary Environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>1541 Ma</td>
<td>Carswell Fm</td>
<td></td>
<td></td>
<td></td>
<td>4</td>
<td>Dolomite and stromatolite</td>
<td>Marine, lagoonal and distal braided fluvial</td>
</tr>
<tr>
<td></td>
<td>Douglas Fm</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Black mudstone and fine quartz arenite</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Otherside Fm</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Quartz arenite and pebbly quartz arenite</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Locker Lake</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Quartz arenite, pebbly quartz arenite and conglomeratic quartz arenite</td>
<td></td>
</tr>
<tr>
<td>1644 Ma</td>
<td>Wolverine Point Fm</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Quartz wacke, quartz arenite and mudstone</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Lazenby Lake</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Quartz arenite pebbly quartz arenite, quartz wacke, siltstone and mudstone</td>
<td>Marine and braided fluvial</td>
</tr>
<tr>
<td></td>
<td>Manitou Falls Fm</td>
<td></td>
<td></td>
<td></td>
<td>3</td>
<td>Quartz arenite, pebbly quartz arenite and conglomeratic quartz arenite with minor clay interbeds</td>
<td>Braided fluvial and alluvial fan</td>
</tr>
<tr>
<td></td>
<td>Read/Smart</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Pebbly quartz arenite and conglomerate</td>
<td></td>
</tr>
<tr>
<td>1750 Ma</td>
<td>Fair Point Fm</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Pebbly quartz arenite, quartz arenite, and conglomerate with basal pebbly mudstone</td>
<td>Alluvial fan and braided fluvial</td>
</tr>
<tr>
<td></td>
<td>Basement</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Metamorphosed Archean and Paleoproterozoic granitoid gneiss and Paleoproterozoic metasedimentary rocks</td>
<td></td>
</tr>
</tbody>
</table>
**2.3 Unconformity-related uranium deposits**

Unconformity-related uranium (URU) deposits in the Athabasca Basin typically occur near the unconformity between the basement and the basin and are spatially associated with reactivated basement faults and graphitic lithologies (Hoeve and Sibbald, 1978; Jefferson et al., 2007; Kyser and Cuney, 2008). The uranium deposits are classified into two types, i.e., unconformity-hosted deposits and basement-hosted deposits. The unconformity-hosted uranium deposits develop immediately above or below the sub-Athabasca unconformity (e.g., Cigar Lake, Midwest), whereas the basement-hosted uranium deposits occur up to hundreds of meters below the unconformity (e.g., Eagle Point deposit, Rabbit Point, Millennium). Some deposits comprise both unconformity-hosted and basement-hosted orebodies (e.g., McArthur River deposits).

Local uranium mineralization occurring in the sandstones of the basin hundreds of meters above the unconformity is termed as perched uranium mineralization (Fayek et al., 1997).

Uranium orebodies are generally enveloped by plume-shaped alteration haloes that formed from long-term chemical reaction between ore-forming fluids and host rocks. Alteration halos are believed to be genetically linked with the mineralization, and thus can serve as exploration indicators. Based on the assemblages of clay minerals, alteration
haloes can be classified into illite-dominated haloes, kaolinite-dominated haloes and chlorite-rich haloes (Jefferson et al., 2007). It is generally considered that URU deposits formed from the redox reaction between uranium-rich, oxidized, high-salinity basinal brines and reduced basement fluids or lithologies at reactivated basement faults (Jefferson et al., 2007). There are still many uncertainties about the genetic model of URU mineralization, and the one of the most important issue that is unsolved and hot-debated is the source of the U. One group of researchers advocated that the primary U source is the basin sediments (Hoeve and Sibbald, 1978; Hoeve and Quirt, 1984; Kotzer and Kyser, 1995; Komninou and Sverjensky, 1996; Fayek and Kyser, 1997; Chi et al., 2019), while scholars from the other school believed that uranium is mainly sourced from the basement (Annesley et al., 1999; Cuney et al., 2003; Richard et al., 2010; Mercadier et al., 2013). Both schools have their own reasons to support their viewpoints. The former school maintains that basin sediments that are composed of high-permeability clastic redbeds were favorable for the movement of oxidized fluids, and thus fluid circulation developed in sandstone successions may have leached U from U-rich minerals in the basin. On the other hand, the basement that is composed of crystalline rocks has an extremely low-permeability and reduced environments, which is unfavorable for both U mobilization and fluid movement, and therefore is unlikely to be the primary U source. However, the latter school argued that based on the chemical analysis of elemental
composition on rocks from the basement and basin, the basement is much more enriched in U than the quartz-dominated basinal sediments and therefore it is plausible that U was uptaken from the basement rocks through the penetration of oxidizing basinal brines into the basement along permeable structures. Irrespective of the disagreement on uranium source, both schools have paid significant attention on the fluid circulation and its importance to U mineralization in consideration of the geological and geochemical records of intense water-rock reaction discovered near the uranium deposits.

Most recently, a fluid inclusion study of trace elements within quartz overgrowths of sandstones from the central basin far away from any known U deposits suggested that U-rich basinal brines were once pervasively developed throughout the basin (Chi et al., 2019). The timing of the development of theuraniferous brines within the basin, as revealed by the fluid inclusions entrapped near the quartz overgrowth – detrital quartz boundaries, have important implications for the relationships between sediments compaction, cementation and uranium mineralization.
Chapter 3: Study Methods

3.1 Petrographic analysis

3.1.1 Study materials

The basic part of this study is petrographic analysis involving petrographic observation and point counting. A total of 250 thin sections that were collected from four drill holes (Rumpel Lake, WC-79-1, BL-08-01, DV10-001) in the central Athabasca Basin have been prepared and petrographically examined in two previous studies (Chu, 2016; Scott, 2016) (Fig. 2.1). The four drill cores were selected for diagenesis studies because they are far (several tens to hundred kilometers) away from any known uranium deposits and are thus unaffected by mineralization events. Furthermore, these four drill cores have more strata information because strata are relatively well preserved in the central part of the basin. Among all the collected thin sections, 111 samples were collected from the Rumpel Lake drill core, 40 samples were collected from the WC-79-1, 44 samples were collected from the BL-08-01 and 55 samples were collected from the DV10-001 (Table 3.1). Most of the thin sections that have been studied are quartz-dominated sandstones, while some are conglomerates and a few of them are mudstones.
The four studied drill cores were lithologically logged by former researchers (Fig. 3.1; Bosman et al., 2012; Scott, 2016; Chu et al., 2016). The Rumpel Lake drill core, which is the deepest drill well at the geographic center of the Athabasca Basin, has penetrated the Read, Manitou Falls, Lazenby Lake, Wolverine Point, and Locker Lake formations and intersected with the metamorphic basement at the depth of 1457 m. The DV10-001 drill core, which is located at the depositional center of the basin, has penetrated through the Wolverine Point, Lazenby Lake and Manitou Falls formations, reaching the basement at the depth of 1045 m. The WC-79-1 drill core in the mid-east of the basin has drilled through the Lazenby Lake, Manitou Falls and Read formations, reaching the basement at the depth of 940 m. The BL-08-01 drill core, located to the southeast of the Carswell Structure, has penetrated the Wolverine Point, Lazenby Lake, Manitou Falls, and Read formations and intersected the sub-Athabasca unconformity at the depth of 956 m.
Table 3.1. The number of core samples studied at each formation

<table>
<thead>
<tr>
<th>Formation</th>
<th>Rumpel Lake</th>
<th>WC79-01</th>
<th>BL-08-01</th>
<th>DV10-001</th>
</tr>
</thead>
<tbody>
<tr>
<td>Locker Lake</td>
<td>6</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Wolverine Point</td>
<td>20</td>
<td>-</td>
<td>6</td>
<td>12</td>
</tr>
<tr>
<td>Lazenby Lake</td>
<td>12</td>
<td>1</td>
<td>7</td>
<td>24</td>
</tr>
<tr>
<td>Manitou Falls</td>
<td>59</td>
<td>38</td>
<td>27</td>
<td>19</td>
</tr>
<tr>
<td>Read</td>
<td>14</td>
<td>1</td>
<td>4</td>
<td>-</td>
</tr>
<tr>
<td>Total number</td>
<td>111</td>
<td>40</td>
<td>44</td>
<td>55</td>
</tr>
</tbody>
</table>
Fig. 3.1. Core logging of the Rumpel Lake, WC-79-01, BL-08-01, and DV10-001 drill holes based on the lithological description of previous studies by Chu (2016) and Scott (2016).
3.1.2 Petrographic study of sedimentary components

Petrographic research on the composition and fabric of sandstones was conducted using an Olympus BX51 microscope. K-feldspar staining has verified that no K-feldspar is present in the studied thin sections (Scott, 2016). The staining of thin sections for feldspar identification was conducted by Scott (2016) utilizing the methods of Bailey and Stevens (1960). The components of sandstones were classified into framework grains, depositional matrix, and authigenic cements, which form the basis of point counting work. The hematite coatings that cover detrital quartz grains greatly facilitated the differentiation of detrital quartz from syntaxial quartz overgrowth. Authigenic kaolinite/dickite were identified based on the crystalline aggregation morphology which can be vividly described as ‘book texture’. Illite was distinguished from kaolinite/dickite by its colorful interference colors and ‘needle-shaped’ morphology. Argillaceous materials in sandstones can be divided into sedimentary matrix and pseudomatrix. Pseudomatrix is an altered lithic fragment with a ‘dirty’ appearance and can be easily distinguished from depositional matrix by its granular shape. Without the impregnation of color-dyed resin, pore space in this study was distinguished by ‘transparent’ color under plane-polarized light and complete extinction under cross-plane polarized light. Fake pore space that was created from grain loss during thin section preparation was occasionally observed in this study.
3.1.3 Component quantification

Relative abundances of each component can be measured with conventional point counting study. Three hundred (300) conventional point counts per thin section were performed to obtain the percentage abundance of detrital grains and interstitial components including authigenic quartz cement, authigenic illite, kaolinite/dickite, hematite, matrix and porosity. The specific method for each count is to move the selected thin section randomly under the microscope to determine a random observation scope, at which scope the solid phase strictly at the center of the cross hair is distinguished and recorded as Phase (n) (point counting number\(\text{phase(n)}\) = point counting number\(\text{phase(n) + 1}\)). After 300 times point counting, the percentage abundance of each component can be calculated with the following formula:

\[
\text{percentage}_{\text{phase (n)}} = \frac{\text{Point counting number}_{\text{phase (n)}}}{300} \quad n = 1, 2, 3 \ldots
\]

As for the error of the conventional point counting, previous researchers reported that errors are closely related to counting times and the true composition percentage of components (Folk, 1974; Dutton, 1997). The point counting result of a component truly occupying 50% volume has a maximum error of ± 3.6% when 300 times point counting is carried out. While the error margin for 300 times point counting on a component truly occupying 10% or 90% volume is only about ± 2.1% (Folk, 1974; Dutton, 1997). It has
been found that significant error may be introduced by distinguishing components mistakenly. For example, being unable to differentiate syntaxial quartz overgrowth from detrital quartz grains can increase the results of the percentage abundance of detrital quartz grains. This error were avoided in this study thanks to the well-developed hematite coatings, which serve as the marks of the boundaries of the detrital grains. Besides, mistakenly counting the ‘artifact porosity’ as true porosity (Houseknecht, 1987; Ehrenberg, 1989) can increase the results of porosity. The ‘artifact porosity’ that was created during thin section preparation was identified and excluded from point counting results in this study.
3.1.4 Petrographic study of compaction texture characteristics

To evaluate the compaction degree of sandstones, this study quantified grain contact textures of sandstones from different depth intervals of the drill cores. Grain contacts are divided into four types, which are point contact, long contact, concave-convex contact and sutured contact (Taylor, 1950). Point contact is a point-shaped grain boundary that forms from two detrital grains touching at a point without discernable dissolution. Long contact is a linear-shaped boundary between two grains and may form from grains reorientation or moderate pressure solution. Concave-convex contact is an irregular-shaped grain boundary that forms from grains interpenetration and suggest that significant pressure solution has occurred at grain contacts. Sutured contact is an irregular, jagged-shaped grain boundary that forms as the result of significant pressure solution and grain interpenetration (Fig. 3.2). Grain contact textures are irreversibly modified during burial-diagenesis, and thus can be used as an indicator of compaction degree (Taylor, 1950). Percentage of each type of grain contacts has a close relationship with burial depth. That is, right after the deposition of clastic sediments, grain contacts are dominantly point contacts or null-contacts (floating grains). With overlying sediments progressively accumulating above, detrital grains of sandstones rotate and approach to each other with larger numbers of tight contacts developing as a result of pressure solution (Taylor, 1950).
Fig. 3.2 Schematic diagram showing different types of grain contacts in sandstones as defined by Taylor (1950).
3.1.5 *Compaction texture quantification*

Three parameters are chosen as the indicators of compaction quantification, including contact index (CI = the average number of all kinds of intergranular contacts), tight packing index (TPI = the average number of long, concave-convex and sutured contacts) and intergranular volume (Fig.3; IGV = the sum of intergranular porosity, intergranular authigenic cement and sedimentary matrix) (Taylor, 1950; Wilson and McBride, 1988; McBride et al, 1991). A total of 100 target grains per thin section were randomly selected for counting the number of any types of grain-grain contacts (point, long, concave-convex and sutured contacts) in each selected samples as proceeded by former researchers (Wilson and McBride, 1988; McBride et al, 1991). The selection of a target grain is achieved by moving a thin section under the microscope randomly to determine a random scope at which the grain strictly at the center of the cross hair is the target grain, and the number of detrital grains directly contacting the target grain, as well as their corresponding grain contact types, are recorded (as illustrated in Fig. 3.4). Besides, we also measured the maximum diameters of all target grains in order to calculate the grain size and sorting index of the selected thin section. Grain size is represented by the mean value of the maximum diameters of the 100 target grains, and sorting index is represented by the standard deviation of the maximum diameters of the 100 target grains.
Fig. 3.3 Schematic diagram illustrating intergranular volume (IGV) defined as the sum of intergranular porosity (%), cement (%), and depositional matrix (%). Note that view (a) and (b) have the same value of IGV, but intergranular porosity of view (b) is much lower than that of view (a) due to the presence of matrix and cement. Framework grains have the same distance in both views, indicating the two samples were compacted to the same degree (modified from Paxton et al., 2002).
Fig. 3.4 Schematic diagram illustrating the specific method at one point counting step. This scope is randomly obtained by moving a thin section randomly under microscope. The target grain is the detrital grain exactly at the center of the cross wire. The grain contacts between the target grains and adjacent grains are recognized and recorded. In this figure, there are one concave-convex contact, one long contact and two point contacts between the target grain and the adjacent grains.
Intergranular volume, a parameter measuring the proximity of framework grains, is a good indicator of compaction and can only be used to evaluate relatively well-sorted, matrix-poor sandstones (not applicable to poorly-sorted or matrix-rich sandstones) (Paxton et al., 2002). Intergranular volume (IGV) can be calculated by summing intergranular porosity, authigenic cement and depositional matrix (Houseknecht, 1987, 1988, 1989; Paxton et al., 1990; Paxton et al., 2002). Mechanical compaction reduces intergranular volume mainly by pushing detrital grains towards each other, resulting in porosity reduction and grain frameworks closest-packed. Another mechanism for IGV decreasing is chemical compaction (pressure solution) that dominates at a deep burial depth (>2 km). During chemical compaction, point contacts transform into long, concave-convex and sutured contacts, which makes the center of framework grains move closer towards each other (Paxton et al., 2002). Mechanical compaction can decrease intergranular volume of well-sorted quartz-rich sandstones (no ductile fragments) from 42% (initial value of IGV) to 26% (the closest packing mode), while the additional decrease of IGV to less than 26% mainly results from chemical compaction (Paxton et al., 2002).
All thin sections were petrographically examined under the optical microscope, but not all of them were used for point counting. The reason is that the main objective of this study is to understand the relationship between compaction and depth, but there are many geological factors other than burial depth that may influence the compaction characteristics of sandstones (e.g., sorting, matrix, and clay coating). Therefore, it is important to select representative samples of similar sedimentary characteristics (grain size, sorting, matrix content) to ensure the selected samples are good indicators of burial depth. The samples that were selected for point counting have the following features: 1) grain-supported (matrix<15%) fabric, 2) fine-/medium-/coarse- grained sandstones (conglomerates were excluded), 3) moderately-well to well-sorted, and 4) no development of clay coatings.

Sorting has been recognized as one of the most important influencing factors if using contact index, tight packing index and intergranular volume to evaluate the degree of compaction (Beard and Weyl, 1973; Rossi and Alaminos, 2014). In poorly-sorted sandstones, relatively small grains tend to cram into the interstitial space between larger grains, which makes the poorly-sorted sandstones appear to be more compacted because finer grains filling the interstitial space between larger grains makes sandstones less porous, and besides, the finer grains occurring in the interstitial space can contact with
the framework grains, which results in a larger number of grain contacts (Fig. 3.5). In contrast, relatively-well-sorted sandstones, in which framework grains support the loading pressure, can be adopted as good indicators of compaction. The sorting of sandstones was visually estimated using the method of Beard and Weyl (1973), and was quantitatively calculated using the standard deviation of the maximum diameters of 100 randomly selected target grains.

Matrix abundance is another factor that significantly influences the compaction characteristics and the results of point counting. If matrix abundance is more than 15%, sandstones typically present a matrix-supported fabric. In this case, detrital grains appear to be ‘cushioned’ and ‘protected’ from compaction by soft clayey matrix (Fig. 3.6). However, even if matrix is less than 15% and detrital grains frameworks present a grain-supported mode, the sample may still not be adopted for study due to the presence of clay coatings, which may chemically promote pressure solution at grain contacts (Fig. 3.6; Wely, 1959; Rneard et al., 1997). An alternative explanation is that clay coatings may inhibit the precipitation of quartz overgrowth that consolidates detrital frameworks and counteracts subsequent compaction after the cementation, which makes the sandstones more compacted than counterparts without clay coatings (Sibley and Blatt, 1976). Therefore, based on the above mentioned sample selection criteria, a total of 21, 16, 13
and 9 thin sections were selected from the Rumpel Lake, WC-79-1, BL-08-01 and DV10-001 drill cores, respectively. These selected samples were point counted in terms of composition, sorting index and compaction degree.
Fig. 3.5 Schematic diagram illustrating that compaction quantification is significantly influenced by the sorting of samples. Note that the larger grains are the same distance apart in both views of a and b, indicating that both samples have been compacted to the same degree. However, it should be noticed that the numbers of grain contacts (marked red) and intergranular volume (white area) are largely different between view (a) and (b). This figure explains the influence of sorting on the values of CI, TPI and IGV.
Fig. 3.6 Schematic diagram illustrating that compaction quantification is closely related to matrix abundance and grain supporting mode. Note that sedimentary matrix can protect grains from compaction. (a) Illustrating that grain contacts of sandstones initially of matrix-supported texture are protected from compaction. (b) Illustrating that grain contacts of sandstones initially of grain-supported texture can be significantly compacted despite the high abundance of matrix.
Considering that distinguishing grain contact types between detrital grains optically are highly subjective (McBride et al., 1991), the error must be carefully evaluated to make sure the accuracy and reproducibility of the point counting results. In this study, error tests were conducted on two samples. For each test, a total of 150 times point counting was performed, and the values of CI and TPI were calculated using the results of every 10n (n=1,2,...,15) times count. The results were plotted together for error examination as illustrated in Fig. 3.7a and Fig. 3.7b. The blue parts in Fig. 3.7a and Fig. 3.7b indicate the areas in which the same results of CI or TPI can be obtained by reserving one decimal. It was observed that the results of the CI and TPI fluctuated greatly when counting times were less than 50, and gradually approximated a stable state with counting times increasing; after 100 times, results were in a stable state. Therefore, the error tests have demonstrated that a total of 100 times point counting work is sufficient for researchers to obtain accurate, reproducible and indicative values of CI and TPI.
Fig. 3.7a Diagram illustrating the results of CI and TPI fluctuate with the increasing of point counting times. Note that the results gradually approach a specific value and become stable. The blue areas indicate that the same results can be obtained by keeping one decimal place. It is observed that reproducible CI and TPI results can be obtained with point counting times more than 100.
Fig. 3.7b Diagram illustrating the results of CI and TPI fluctuate with the increasing of point counting times. Note that the results gradually approach a specific value and become stable. The blue areas indicate that the same result can be obtained by making a round-off to keep one decimal place. It is observed that reproducible CI and TPI results can be obtained with point counting times more than 100.
3.2 Reactive mass transport modeling

3.2.1 Reactive mass transport and TOUGHREACT program

Reactive mass transport modeling couples the calculation of mass transport, fluid flow and chemical reaction and can be carried out by using the TOUGHREACT program; it is a non-isothermal reactive mass transport modeling tool that utilizes Integral Finite Difference (IFD) methods (Xu et al., 2004). IFD facilitates flexible spatial discretization using irregular grids for developing complicated geological media (Xu and Pruess, 1998, 2001). Chemical species for simulating chemical reactions can be accommodated in liquid, gas and solid phases, and can be coupled into fluid flow and solute transport models through iterative computing (Xu et al., 2004). Fluid flow and solute transport can be simulated both in porous and fractured media, with thermophysical and geochemical properties (such as fluid density and viscosity, and thermodynamic and kinetic data) able to be considered in the simulation (Xu et al., 2004). The transport of aqueous and gaseous species by advection and molecular diffusion coupled with fluid flow can be simulated. Aqueous complexation, acid-base, redox, gas dissolution/exsolution, and cation exchange, can be considered under local equilibrium assumption. Mineral dissolution and precipitation can be considered under either local equilibrium or kinetic conditions (Xu et al., 2004). Fluid flow driven by pressure, viscous and gravity forces coupled with heat
flow, conduction and convection is able to be modeled (Xu et al., 2004).

Reactive mass transport modeling is conducted in the following steps (Fig. 3.8; Xu et al., 2004): Firstly, to solve flow equations in each time step, fluid velocities and phase saturation are calculated, and the results are substituted into chemical transport equations that are solved on a component-by-component basis. The concentration of each component obtained by solving transport equations are then substituted into chemical reaction equations that are solved on a grid-block-by-grid-block basis by the Newton-Raphson iteration. Chemical transport and fluid chemical reactions are solved iteratively until convergence. Porosity and permeability change due to the precipitation and dissolution of minerals is considered. Porosity change is calculated to be equal to the change of volume fraction of minerals using the calculation formula defined as (Xu et al., 2004):

\[ \phi = 1 - \sum_{m=1}^{nm} f_{\text{fm}} - f_{\text{fu}} \]

Where \( nm \) is the total numbers of minerals, \( f_{\text{fm}} \) is the volume fraction of a mineral after the reaction in the whole rock (\( V_{\text{mineral}}/V_{\text{rock}} \)), and \( f_{\text{fu}} \) is the volume fraction of a mineral before the reaction in the whole rock. Porosity change can be calculated in each time step, and permeability change that is assumed to a function of porosity change can be calculated using the Carman-Kozeny equation, which is defined as (Bear, 1972; Xu et al.,
2004):

\[ k = k_i \left( \frac{1 - \phi_i}{1 - \phi} \right)^2 \left( \frac{\phi}{\phi_i} \right)^3 \]

Subsequently, the new porosity and permeability of the grid are updated, which would influence fluid flow in the next time step. All the above mentioned calculations and iterations will end when the assigned maximum time steps are reached (Xu et al., 2004).
Fig. 3.8 Flowchart of the TOUGHREACT program (from Xu et al., 2004)
3.2.3 Reactive mass transport modeling of chemical reaction within thermal convection systems

Reactive mass transport modeling attempted to model a quartz dissolution versus precipitation pattern that is similar with the petrographic results. Previous numerical modeling studies of fluid reported that large-scale thermal convection may have occurred in the Athabasca Basin. Thermal convection, an important heat and mass transport mechanism within porous media, typically occurs in a situation that heat conduction is unable to homogenize temperature differences within a non-isothermal fluid domain (Turcotte and Schubert, 2004). Fluid flow of thermal convection is actually driven by density differences that are related to the heat-expansion and cold-contraction phenomenon. Due to thermal gradients in sedimentary basins, warmer fluid at a deep part of the basin has a relatively low fluid density and thus experiences a buoyancy force with a tendency to rise upward, while colder fluid at a shallow part of the basin has a relatively high fluid density and tends to sink. In this case, adjacent upwelling and downwelling plumes make up a fluid convection cell in which fluid circulates up and down, which is of great significance to mass transport in sedimentary basins (Worden and Morad, 1982). Within fluid-saturated, homogeneous, isotropic and confined porous media, the tendency of fluid migration towards fluid convection can be quantitatively evaluated by a dimensionless Rayleigh number (Ra), whose formula is mathematically defined as
(Turcotte and Schubert, 2004):

\[ Ra = \frac{\rho_f g \beta \rho_f K H \Delta H}{\eta \lambda_m} \]

Ra, the Rayleigh number in the right part of the formula, represents the ratio of buoyancy forces to viscous forces (Turcotte and Schubert, 2004). Where \( \rho_f \) is the average fluid density, \( \beta \) is the thermal expansion coefficient of the fluid (K\(^{-1}\)), \( g \) is the acceleration of gravity, \( c_{pf} \) is the isochoric heat capacity (J (kg \cdot K\(^{-1}\)), \( K \) is the porous medium permeability (m\(^2\)), \( \Delta T \) is the temperature difference across height \( H \), \( \eta \) is the fluid viscosity and \( \lambda_m \) is the thermal conductivity of the porous medium (J (s \cdot m \cdot K)). The threshold value of Ra for fluid convection to occur is named critical Rayleigh number. When \( R_a < \) critical Rayleigh number, heat transfer are primarily in the form of heat conduction, in which case fluid convection is not important to mass transport. When \( R_a > \) critical Rayleigh number, fluid convection becomes the major driving mechanism for mass transport. The TOUGHREACT program has the capability of simulating mineral dissolution versus precipitation process by coupling fluid flow, heat transfer, solute transport and chemical reaction in geological systems in which thermal convection prevails.

The TOUGHREACT program enables users to model reactive transport reaction using kinetic or equilibrium approaches. Kinetic approach enables users to model chemical
reaction system of high fluid velocity, in which case local chemical equilibrium cannot be reached within a short period of reaction times (Simunek and Valocchi, 2002).

Equilibrium approach is applicable to chemical reactions of which the chemical reaction rates are much faster than the rates of mass advection and dispersion (Knapp, 1989, Bahr, 1990, and Lichtner, 1996). A Damkohler number (Da) that represents the ratio of chemical reaction rate to diffusion and/or dispersion rate is proposed to differentiate whether equilibrium approach is valid or not (Kuhn, 2004; Fogler, 2006). If Da >> 1, the chemical reaction rate is much faster than the diffusion and/or dispersion rate so that the equilibrium approach is valid; while if Da << 1, the rate of mass diffusion and/or dispersion is much faster than the chemical reaction rate, and the kinetic approach is the only choice (Steefel and Lasaga, 1990; Kuhn, 2004). Considering flow velocities of fluid convection associated with thermal gradient are at a very slow rate of ~1 m/year, quartz dissolution versus precipitation process associated with fluid convection may be at a local equilibrium condition, and therefore, the local equilibrium approach is utilized.
3.2.4 Model setup and parameters used in computation

Four physical (hydrological) models were established in this study, and each of them was composed of a series of physical parameters, mainly including porosity, permeability, specific heat capacity, thermal conductivity and thermal gradient. Porosity and permeability are assigned based on the consideration that deeply buried sandstones are relatively impermeable compared with the counterparts that have been buried at a relatively shallow burial depth. The Physical models established in this study are all 2D models in which the length at the out-of-plane direction is arbitrarily assigned to be 50 m.
3.2.5 Physical models

Based on the lithological framework of the Athabasca Basin, the mud-rich, fine-grained Wolverine Point and Douglas formations have been widely considered as sedimentary aquitards (Hiatt and Kyser 2007); while the sandstone packages composed of the Read, Manitou Falls, Lazenby Lake formations have been deemed as a large-thickness, high-permeability, quartz-rich aquifer that is confined by the basement and the Wolverine Point Formation; At the stratigraphically higher units above the Wolverine Point Formation, the sandstone succession composed of Locker Lake and Otherside formations can be regarded as another high-permeability, large-thickness aquifer that is confined by the Douglas and Wolverine Point formations. Previous numerical modeling studies proposed that fluid convection associated with geothermal gradients may have developed in the sandstone successions of the Athabasca Basin (Raffensperger and Garven., 1995; Cui et al., 2012; Li et al., 2016).

Physical model 1 and 2 are specifically designed to model quartz dissolution versus precipitation pattern in one confined aquifer that corresponds to the sandstone succession composed of the Read, Manitou Falls and Lazenby Lake formations. The two physical models were assigned with the same formation thicknesses and geometric shape. However, other physical parameters including depth, permeability and porosity were
assigned differently. Both the physical model 1 and 2 were assigned to have a
metamorphosed basement of 1000 m thickness at the base, a sandstone-dominated aquifer
of 1500 m thickness in the middle and a mud-rich aquitard of 300 m thickness at the top.
The interface between the basement and the sandstone-dominated aquifer represents the
sub-Athabasca unconformity where unconformity-related U mineralization might occur if
favourable conditions were met. In physical model 1, this interface was at 3 km depth,
indicating a relatively shallow burial condition; in the physical model 2, it was at 6 km
depth (Fig. 3.9).

Physical model 3 and 4 were used to model quartz dissolution versus precipitation pattern
within two confined sandstone units. One sandstone aquifer was confined by the
basement and the Wolverine Point Formation, and the other one is confined by the
Wolverine Point and the Douglas formations. Similarly, the physical model 3 and 4 were
assigned to have the same formation thickness and geometry with the assignment of
different depth, permeability and porosity. The physical model 3 and 4 had a 1000 m
thickness basement, a 1500 m thickness sandstone aquifer above the basement, a 300 m
thickness aquitard in the middle, a 600 m thickness sandstone aquifer above the aquitard,
and a 600 m thickness cover (aquitard) at the top (Fig. 3.10). The sub-Athabasca
unconformity was located at 3 and 6 km depth in the physical model 3 and 4 (Fig. 3.10).
Fig. 3.9 Cross-sectional view of the physical model 1 and 2. Note the main differences between the physical model 1 and 2 are the burial depth and sandstone permeability. (a) Physical model 1 represents the shallow burial condition with sandstone permeability equal to 100 md. (b) Physical model 2 represents the deep burial condition with sandstone permeability equal to 200 md.
Fig. 3.10 Cross-sectional view of the physical model 3 and 4. Note the differences between the physical model 3 and 4 are the burial depth and the permeability of sandstones. (a) Physical model 3 was used to model the shallow burial condition (Sub-Athabasca unconformity at 3 km depth). (b) Physical model 4 was used to model the deep burial condition (Sub-Athabasca unconformity at 6 km depth).
3.2.6 Physical parameters

Individual hydrological unit was arbitrarily assigned to be homogeneous and isotropic as proceeded by several previous numerical simulation studies in order to save running time of the program (Raffernsperger and Garven, 1995; Cui et al., 2012; Li et al., 2016). Physical parameters of rock properties assigned mainly include density, thermal conductivity, specific heat capacity, porosity and permeability. The density, thermal conductivity and specific heat capacity of each hydrological unit were assigned according to the published studies and compilations (Table 3.2; Yang et al., 2004; Cui et al., 2012; Chi et al., 2013, 2014; Li et al., 2016).

The assignment of porosity and permeability was based on the consideration of burial depth. The sediments that were buried at a deep burial depth were subjected to significant compaction, and therefore have low porosity and permeability compared with the counterparts that only have been buried to a shallow depth. The physical model 1 that was designed to model a hot and shallow diagenetic environment was assigned to have relatively high porosity and permeability, in which case the basement, sandstones and confining aquitard were assigned to have permeability of 0.01 md, 100 md, and 0.01 md respectively. The physical model 2 that was designed to model a deeply buried environment, in which case the basement, sandstones, and confining aquitard were
assigned to have permeability of 0.01 md, 50 md and 0.01 md respectively (Table 3.3). In
the physical model 3 which was designed to model a shallow burial condition, for the
hydrological units from the bottom to the top, the basement, lower sandstones,
intermediate aquitard, upper sandstones and top aquitard were assigned to have
permeability of 0.01 md, 100 md, 0.01 md, 200 md and 0.01 md separately. In the
physical model 4 which was designed to model a deeply buried condition, for the
hydrological units from the bottom to the top, the basement, lower sandstones,
intermediate aquitard, upper sandstones and top aquitard were assigned to have
permeability of 0.01 md, 50 md, 0.01 md, 100 md and 0.01 md respectively (Table 3.3).
Table 3.2 Input parameters of hydrological units used in the setups of physical model 1 and 2 (after Cui et al., 2012; Li et al., 2016)

<table>
<thead>
<tr>
<th>Units/properties</th>
<th>Density (kg/m³)</th>
<th>Porosity</th>
<th>Permeability (m²)</th>
<th>Thermal conductivity (W/(m · °C))</th>
<th>Specific heat capacity (J/(kg · °C))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Physical model 1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cover</td>
<td>2400</td>
<td>0.1</td>
<td>1.0×10⁻¹⁶</td>
<td>2.5</td>
<td>803</td>
</tr>
<tr>
<td>Sandstone</td>
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<td>803</td>
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<tr>
<td>Basement</td>
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<td>0.1</td>
<td>1.0×10⁻¹⁶</td>
<td>2.5</td>
<td>803</td>
</tr>
<tr>
<td>Physical model 2</td>
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<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cover</td>
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<td>0.1</td>
<td>1.0×10⁻¹⁶</td>
<td>2.5</td>
<td>803</td>
</tr>
<tr>
<td>Sandstone</td>
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<td>5.0×10⁻¹⁴</td>
<td>3.5</td>
<td>803</td>
</tr>
<tr>
<td>Basement</td>
<td>2650</td>
<td>0.1</td>
<td>1.0×10⁻¹⁶</td>
<td>2.5</td>
<td>803</td>
</tr>
<tr>
<td>Units/properties</td>
<td>Density (kg/m³)</td>
<td>Porosity</td>
<td>Permeability (m²)</td>
<td>Thermal conductivity (W/(m · °C))</td>
<td>Specific heat capacity (J/(kg · °C))</td>
</tr>
<tr>
<td>-------------------</td>
<td>-----------------</td>
<td>----------</td>
<td>-------------------</td>
<td>----------------------------------</td>
<td>-------------------------------------</td>
</tr>
<tr>
<td>Physical model 3</td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
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<td>1.0×10^{-16}</td>
<td>2.5</td>
<td>803</td>
</tr>
<tr>
<td>Aquifer 2</td>
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<td>2.0×10^{-13}</td>
<td>3.5</td>
<td>803</td>
</tr>
<tr>
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<td>1.0×10^{-16}</td>
<td>2.5</td>
<td>803</td>
</tr>
<tr>
<td>Aquifer 1</td>
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<td>1.0×10^{-13}</td>
<td>3.5</td>
<td>803</td>
</tr>
<tr>
<td>Basement</td>
<td>2650</td>
<td>0.1</td>
<td>1.0×10^{-16}</td>
<td>2.5</td>
<td>803</td>
</tr>
<tr>
<td>Physical model 4</td>
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<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cover 2</td>
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<td>0.1</td>
<td>1.0×10^{-16}</td>
<td>2.5</td>
<td>803</td>
</tr>
<tr>
<td>Aquifer 2</td>
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<td>1.0×10^{-12}</td>
<td>3.5</td>
<td>803</td>
</tr>
<tr>
<td>Cover 1</td>
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<td>1.0×10^{-16}</td>
<td>2.5</td>
<td>803</td>
</tr>
<tr>
<td>Aquifer 1</td>
<td>2500</td>
<td>0.25</td>
<td>5.0×10^{-13}</td>
<td>3.5</td>
<td>803</td>
</tr>
<tr>
<td>Basement</td>
<td>2650</td>
<td>0.1</td>
<td>1.0×10^{-16}</td>
<td>2.5</td>
<td>803</td>
</tr>
</tbody>
</table>
3.2.7 Initial and boundary conditions

Previous numerical modeling studies suggest the fluid pressure regime within the Athabasca basin was at a near-hydrostatic condition rather than lithostatic condition because the high permeability of quartzose sandstones is favorable for fluid migration and pressure release, in which case overpressure is unlikely to develop (Chi et al., 2013, 2014). Therefore, a hydrostatic pressure gradient \(1.013 \times 10^4 \text{ pa/m}\) was assigned and the initial fluid velocity was arbitrarily assigned to be zero automatically by the TOUGHREACT program. The upper, bottom and side boundaries of the models were assumed to be impermeable to fluid flow. As for initial geothermal gradients, an elevated geothermal gradient of \(57^\circ\text{C}/\text{km}\) was assigned to the physical model 1 and model 3 according to the shallow burial model of unconformity-related U mineralization (Chi et al., 2018); a normal geothermal gradient of \(30^\circ\text{C}/\text{km}\) was assigned to the physical model 2 and model 4 that were used to model the condition corresponding to the conventional deep burial diagenetic-hydrothermal model (e.g. Pagel, 1975).
3.2.8 Chemical reactions and parameters

The solubility of quartz varies at different geological environments. The solubility of quartz is widely known to be related with temperature and pH conditions. Generally, at a relatively high temperature and high pH condition, quartz tends to be dissolved. While at a low temperature and low pH condition, the dissolved quartz tends to precipitate. Quartz dissolution versus precipitation process in aqueous solution can be expressed by the following chemical equations:

\[
\text{SiO}_2(s) + 2\text{H}_2\text{O}(aq) \leftrightarrow \text{H}_4\text{SiO}_4(aq)
\]

\[
\text{H}_4\text{SiO}_4(aq) + \text{OH}^- \leftrightarrow \text{H}_2\text{SiO}_4^-(aq) + \text{H}_2\text{O}(aq)
\]

If an aqueous solution is acidic or neutral, the influence of pH on quartz solubility is negligible because the dissociation of \( \text{H}_4\text{SiO}_4 \) requires a high concentration of hydroxide, which is at a very low concentration in acidic or neutral environments (Weast et al., 1986, as illustrated in Fig. 3.11). Therefore, in this case, the chemical formula of the quartz dissolution versus precipitation can be briefly expressed as:

\[
\text{SiO}_2(s) \leftrightarrow \text{SiO}_2(aq)
\]

The equilibrium constant can be expressed by the following equation:

\[
K = \frac{\left[ \text{Activity}_{\text{SiO}_2(aq)} \right]}{\left[ \text{Activity}_{\text{SiO}_2(s)} \right]} = \left[ \text{Activity}_{\text{SiO}_2(aq)} \right] = \left[ \text{Solubility}_{\text{SiO}_2(aq)} \right]
\]

The chemical reaction equilibrium constant (K) at a variety of temperature has been
incorporated into the thermodynamic database of the TOUGHREACT program (Table 3.4). In the thermodynamic database, internal values between adjacent temperature points are linearly extrapolated as illustrated in Fig. 3.12. It should be noticed that the default data within the thermodynamic database do not consider the influence of pH and therefore are only applicable to the simulation of quartz dissolution versus precipitation process in acidic or neutral (pH<8) environments, which is just the case of the Athabasca Basin (Jefferson et al., 2007).
Fig. 3.11 (a) Activities of dissolved silica species in an equilibrium condition at 25°C. Note that silica solubility is pH-independent at pH<9, which increases sharply with the pH increasing at pH>9. (b) Solubility of quartz as a function of temperature at pH<7 (Weast et al., 1986).
Table 3.4 Solubility equilibrium constant $K$ and solubility of quartz at different temperature points in the TOUGHREACT thermodynamic database (Xu et al., 2004).

<table>
<thead>
<tr>
<th>Temperature ($^\circ$C)</th>
<th>0</th>
<th>25</th>
<th>60</th>
<th>100</th>
<th>150</th>
<th>200</th>
<th>250</th>
<th>300</th>
</tr>
</thead>
<tbody>
<tr>
<td>mol/L</td>
<td>0.00008</td>
<td>0.00018</td>
<td>0.00045</td>
<td>0.00102</td>
<td>0.00228</td>
<td>0.00432</td>
<td>0.00622</td>
<td>0.00962</td>
</tr>
<tr>
<td>g/L</td>
<td>0.005</td>
<td>0.012</td>
<td>0.029</td>
<td>0.065</td>
<td>0.146</td>
<td>0.276</td>
<td>0.398</td>
<td>0.615</td>
</tr>
<tr>
<td>ppm</td>
<td>5.34</td>
<td>11.67</td>
<td>28.65</td>
<td>65.19</td>
<td>145.94</td>
<td>276.17</td>
<td>398.27</td>
<td>615.43</td>
</tr>
</tbody>
</table>

Fig. 3.12 Quartz solubility as a function of temperature in the TOUGHREACT thermodynamic database (Xu et al., 2004).
In terms of chemical input, sandstones were assigned to be fully composed of quartz, and pore water was arbitrarily assigned to solely accommodate SiO$_2$(aq) without any other aqueous species considered; this assignment was greatly simplified, but was likely to be valid considering that aqueous SiO$_2$ cannot react with any other components in diagenetic environments (T<250°C). The sedimentary aquitard and basement were arbitrarily assigned to be free from chemical reaction, which was based on the consideration that the diagenesis system in the low-permeability aquitard is geochemically closed (Bjørlykke and Jahren, 2012). In other words, the reactive mass transport modeling is simplistically considered within the confined sandstone packages. Given that the fluid flow velocity of free convection associated with thermal gradient is at a very slow rate (about 1 m/year), quartz dissolution versus precipitation process was assigned to stay at a local equilibrium condition. This assignment was based on the assumption that the rates of quartz dissolution and/or precipitation are faster than the rates of advection and dispersion of silica associated with fluid convection in subsurface environments (Knapp, 1989, Bahr 1990, and Lichtner, 1996).
### 3.2.9 Computation grid and time step

Appropriate space discretization and time step choice are significant for obtaining accurate numerical solutions. Extremely small grid size and time step are beneficial for the accuracy of numerical results but dramatically increase the running time so that decrease the calculation efficiency. Therefore, to ensure both the accuracy and efficiency, tests were conducted by running models with a variety of grid size and time steps within a relatively short period of time (200,000 years). It was found that modeling results can show good convergences once the maximum grid size of sandstone units was less than 100 m and time step size was less than 2.0 year. Thus, the grid size of sandstone aquifers was set to be 50×50×50 m³; the basement and aquitard that are not geochemically concerned were assigned to have a larger grid size (100×50×100 m³) in order to reduce the running time; The time step size was assigned to be 1.0 year. Automated time step adjustment enables the code to run in a “quasistationary states” (QSS), which allows the program to choose larger time steps (greater than 1.0 year) to reduce the running time (Lichtner, 1988). If the QSS is activated, time step is automatically adjusted by the code depending on the convergence of the problem. If the problem system converges properly, the next time step size will increase; otherwise it will decrease automatically (Xu et al., 2004). In this study, we have activated this function to reduce running time and the upper limit of time step size was assigned to be 2 years.
Chapter 4: Petrographic Studies of Compaction and Cementation

4.1 Petrography of detrital and authigenic components

4.1.1 Detrital components

Detrital quartz

Detrital quartz is the most common clastic component in the Athabasca Basin. Detrital quartz in this study can be separated into monocrystalline quartz and polycrystalline quartz, and most of quartz grains are monocrystalline. Most of monocrystalline quartz grains exhibit uniform extinction while the others exhibit undulated extinction (Fig. 4.1A). Polycrystalline quartz is a granular aggregation of many optically different quartz crystal units and looks like a single grain (Fig. 4.1A). Polycrystalline quartz grains are mechanically less stable than monocrystalline quartz and are occasionally observed to be crushed into several pieces of small quartz fragments as the result of mechanical compaction. Hematite coatings covering detrital grains mark the boundary between quartz grains and authigenic quartz overgrowths. Due to significant secondary dissolution, quartz is partially dissolved at point contacts or at the edge of grains.
**Feldspar**

Detrital feldspar is absolutely absent in the sandstones of the Athabasca Group. It is inferred that detrital feldspar grains have been thoroughly leached out due to strong chemical interaction between water and minerals during a prolonged diagenetic history of the Athabasca Basin.

**Lithic Fragments**

Lithic fragments were absolutely altered to become argillaceous pseudomatrix (Fig. 4.1B). Compared with rigid detrital quartz grains, altered lithic fragments are ductile components that can be significantly deformed due to mechanical compaction.

**Muscovite**

Muscovite fragments are commonly seen within the Athabasca Group but is generally of very little quantities (<1%). Muscovite fragments are generally deformed and bent due to significant mechanical compaction as the result of overburden pressure (Fig. 4.1C).

**Zircon**

Detrital zircon grains generally occur as anhedral to subhedral clasts with remarkable high reliefs and dark outlines. Detrital zircon can survive from pressure solution and be
preserved together with insoluble argillaceous materials within stylolites.

**Matrix**

Matrixes are very fine-grained argillaceous materials that deposit together with detrital grains rather than being chemically precipitated from interstitial pore water. It can be observed that a majority of matrixes are partially recrystallized and many irregular-shaped, tiny crystallized fragments are seen to be extensively distributed on uncrystallized argillaceous materials (Fig. 4.1C). Interstitial matrixes generally serve as soft cushions that protect quartz grains from being significantly compacted, and in this case, grain contacts of the matrix-supported sandstones are mostly null contacts or point contacts.
4.1.2 Authigenic components

Syntaxial quartz overgrowth

Quartz cement is one of the most common authigenic phase in the sandstones of the Athabasca Group. Most of quartz cements occur in the morphology of syntaxial quartz overgrowths which can be separated from detrital quartz with the help of dark-colored rims of hematite coatings. Syntaxial quartz overgrowths generally have thicknesses ranging from 20 μm to 100 μm. Edges of quartz overgrowths is observed to be partially dissolved or replaced by clay minerals and thus present an irregular, cloudy-shaped or even embayment-shaped boundaries. Percentage abundance of syntaxial quartz cements is closely interrelated with the compaction characteristics. Quartz cement can only fill the open pore remained after compaction, and in turn quartz cement retards subsequent compaction by consolidating grains framework. Therefore, sandstones which are pervasively cemented by quartz overgrowths usually have large numbers of point contacts and null-contacts, with concave-convex and sutured contacts being rarely seen (Fig. 4.2A). However, sandstones that are poorly cemented by syntaxial quartz overgrowth tend to be significantly compacted, which is characterized by having large numbers of concave-convex and sutured contacts (as seen in Fig. 4.2B). The intimate relationship between quartz cementation and compaction will be discussed in detail in the following part of the thesis.
**Illite cement**

Illite that has high first-order interference colors occurs in intergranular or secondary pore space as hair-like filamentous aggregation. It is very common that illite coexists with other types of authigenic minerals (Fig. 4.1D), for example, illite intergrows with kaolinite/dickite aggregations in the remaining interstitial pore after quartz cementation, which indicates the paragenetic relation that authigenic illite was precipitated after the major quartz cementation.

**Kaolinite/dickite cement**

Aggregations of kaolinite and dickite have very similar crystallized morphology that is microscopically characterized with “book” textures (Fig 4.1E). It is difficult to distinguish kaolinite from dickite using optical microscope, and therefore kaolinite and dickite were labeled together as kaolinite/dickite for petrographic description in this study.

**Hematite cement**

Hematite is iron oxide/hydroxide that mainly occurs as dark-colored grain coatings that separate detrital quartz grains and syntaxial quartz overgrowths, or appears as interstitial pore-filling materials (Fig. 4.2C).
Fig. 4.1 Photomicrographs of typical compositions in the Athabasca Group. (A) Grains of monocrystalline quartz (MCQ) and polycrystalline quartz (PCQ). Sample 01GC303, 54.4 m depth, LL. (B) Altered lithic fragments. Sample 11GC013, 440.4 m depth, LZ. (C) Sedimentary matrix and mica fragments. Sample 01GC317, 299.0 m depth, WP. (D) Authigenic illite. Sample 11GC341, 809.8 m depth, MF. (E) Authigenic kaolinite/dickite. 11GC016, 693.6 m depth, MF. (E) Argillaceous materials and tiny quartz fragments in mudstone. 01GC315, 258.7 m, WP.
Fig. 4.2 Photomicrographs of typical texture in the Athabasca Group. (A) Quartz grains floating in quartz overgrowth. Sample 1401, 5.8 m, LZ. (B) Significantly-compacted texture. Sample 11GC030, 1103.4 m, MF. (C) Sandstones of a low degree of compaction. Sample 1425, 576.5 m, MF. (D) Stylolite. Sample 01GC356, 1139.6 m, MF. (E) Quartz wacke. Sample 01GC322, 390.1 m, LZ. (F) Sharp boundary between quartz arenite and quartz wacke. Sample 01GC317, 299 m, WP.
4.1.3 Paragenetic relationships

The sequence of diagenetic phases/processes that control the sandstones petrology of the Athabasca Basin was defined based on the textural paragenetic relationships among the diagenetic minerals relative to deposition and compaction using the optical microscope. The paragenetic sequences of the sandstones are the reflection of a progressive burial history that sandstones have undergone, and has been investigated in detail by many former studies (e.g. Kotzer and Kyser, 1995; Kyser et al., 2000). Soon after deposition, alteration and dissolution of unstable detrital components (feldspar, lithic fragments and magmatic heavy minerals) may have occurred through the reaction with oxidant and actively circulating meteoric waters. It is likely that all the less stable species of detrital heavy minerals were leached out and their dissolution resulted in the precipitation of iron oxides (hematite). Nearly all hematite coatings which separate detrital quartz and syntaxial quartz overgrowth develop at this period of time (Fig. 4.3A). With the overlying sediments accumulating, the mechanical compaction of sandstones became the dominant diagenetic process that strongly impacted the textural characteristics (proximity of detrital grains) and physical properties (porosity and permeability). Quartz cementation Q1 occurred during mechanical compaction before chemical compaction and significantly retarded the subsequent compaction, in which case large numbers of null contacts and point contacts can be observed (Fig. 4.3B).
Chemical compaction is more conspicuous in the sandstones of the Athabasca Basin. Intergranular pressure solution resulted in the deformation of grain contacts. Long, concave-convex and sutured contacts have been extensively detected throughout the stratigraphy, except in the sandstones that were pervasively cemented by syn-compaction quartz overgrowth Q2. As the incidence of intergranular pressure solution of quartzose sandstones occurs normally only deeper than 2–3 km, the significant intergranular pressure solution at grain contacts indicate that sandstones of the Athabasca Basin may have been subjected to great burial depth during the burial history. While another interpretation for the significant chemical compaction is that there might be a substantially higher-than-normal thermal disturbance that heated the basinal sediments and accelerated the chemical compaction process. Many syn-compaction quartz overgrowth developed at this period of chemical compaction, which is reflected by the petrographic characteristics that quartz overgrowths coexist with tight grain contacts (long, concave-convex and sutured) contacts (Fig. 4.3C). At a deeper burial depth, the precipitation of illite and dickite within the remaining intergranular pore space and the dissolution of quartz occurred (Fig. 4.3D). The diagenetic paragenesis relative to deposition and compaction is illustrated in Fig. 4.4.
Fig. 4.3 Photomicrographs showing paragenetic sequences of sandstones. (A) Hematite coatings occur between detrital quartz grains and syntaxial quartz overgrowth, while interstitial authigenic hematite fills the remaining pore after quartz cementation. Sample 1452, 250.8 m, LZ. (B) Detrital quartz grains having null or point contacts with each other, indicating pre-compaction quartz cementation Q1 occurred before the completion of mechanical compaction. Sample 1401, 5.8 m, LZ. (C) Detrital quartz grains having point or long-contacts with each other, indicating syn-compaction quartz cementation Q2 occurred during the stage of chemical compaction. Sample 1425, 576.5 m, MF. (D) Illite occurs at the remaining intergranular pore space after chemical compaction and quartz cementation. Sample 1439, 911.7 m, MF.
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Fig. 4.4 Diagram showing mineral diagenetic paragenesis (relative to deposition and compaction) of sandstones in the Athabasca Basin.
4.2 Point counting results

4.2.1 Component-related point counting results

Component-related point counting data is tabulated in Table 4.1. All selected samples for point counting are well- to moderately-well sorted, matrix-poor quartz arenite. The sorting index and mean grain size of these samples are calculated with the point counting data of the maximum diameters of 100 target grains, and the results are plotted in Fig. 4.5. Mean grain size of the samples ranges from 0.68 φ to 2.18 φ with an average of 1.35 φ, and the sorting index ranges from 0.41△φ to 0.59△φ with an average of 0.51△φ.

Quartz cement is the most important authigenic phase in this study and can be found at any given depth interval. It is found that the distribution of authigenic quartz is not homogeneous with respects to burial depth (as seen from Fig. 4.6), which has been revealed from the plotting diagrams of Fig. 4.6. Specifically, in the Rumple Lake drill core, above the Wolverine Point Formation, sandstones in the Locker Lake Formation contain a very little amount of quartz cement with average percentage abundance less than 3.3%, while below the Wolverine Point Formation, quartz cement abundance at the Lazenby Lake Formation can be up to 22.7%, which sharply decreases to nearly 0% at the sandstones of the Read Formation near the basement. In the WC-79-01 drill core,
quartz cement abundance decreases from 21.7% at the Lazenby Lake Formation to about 2% at the Manitou Falls Formation near the basement, which is similar to the trend of the Rumpel Lake drill core below the Wolverine Point Formation. In the BL-08-01 drill core, quartz cement abundance decreases from 17.3% at the Wolverine Point Formation, increases and decreases sporadically at the Manitou Falls Formation and finally reaches 3.7% at the Read Formation near the basement. In the DV10-001 drill core, quartz cement abundance does not present a conspicuous decreasing or increasing trend with respect to depth. The maximum value is about 8.7% at the Lazenby Lake Formation and the minimum value is about 0.3% at the Manitou Falls Formation near the basement. However, as for the other three types of authigenic minerals (illite, kaolinite/dickite and hematite), monotonous decreasing or increasing trends with depth are not found.

Percentage abundance of authigenic illite fluctuates throughout the stratigraphy with an average of about 1.2% (Fig. 4.7). Percentage abundance of authigenic kaolinite/dickite has an average of about 2.4%, which does not present an observable trend with depth (Fig. 4.8). Primary intergranular pores that have been significantly diminished by compaction or occluded by authigenic minerals are rarely seen, and nearly all pore space recognized is secondary pore space that mainly forms from the dissolution of detrital quartz or authigenic quartz cement. The porosity of this study can be roughly classified into intergranular porosity, intragranular porosity and ‘artificial’ porosity. The artificial
porosity was created during thin section making and therefore were excluded.

Intergranular porosity is important for calculating IGV which is an indicator of compaction, and it is observed that intergranular porosity generally ranges from 0 to 5.3% with an average of 1.8%, while the percentage abundance of intragranular porosity ranges from 0 to 6% with an average of 1.3%, which do not show a systematic trend with depth.
Table 4.1 Point counting results sandstone components from the Rumpel Lake, WC-79-1, BL-08-01 and DV10-001 drill cores.

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Fm. = formation, Qtz Cement = quartz cement abundance, Int Por = intergranular porosity, Intra Por = intragranular porosity, Tot Por = total porosity, IGV = intergranular volume.
Fig. 4.5 The range of grain size and sorting of the samples covered in this study.
Note there is a slightly negative correlation between grain size and sorting. Most samples are medium-grained (1<φ<2), and some are coarse-grained (0.5<φ<1) and fine-grained (2<φ<3). All samples studied are well sorted (Δφ<0.5) or moderately well-sorted (0.5<Δφ<0).
Fig. 4.6 Quartz cement abundance (syntaxial quartz overgrowth) versus depth of the four drill cores (DV10-001, BL-08-01, Rumpel Lake and WC-79-1). Note the systematic decreasing trend of quartz cement abundance with depth below the Wolverine Point Formation in the BL-08-01, Rumpel Lake and WC-79-01 drill cores. However, there is no conspicuous trend in the DV10-001 drill core.
Fig. 4.7 Authigenic illite abundance versus depth of the four drill cores (DV10-001, BL-08-01, Rumpel Lake and WC-79-1).

Note there is no systematic trend with respect to depth.
Fig. 4.8 Authigenic kaolinite/dickite abundance versus depth of the four drill cores (DV10-001, BL-08-01, Rumpel Lake and WC-79-1). Note there is no systematic trend with respect to depth.
Fig. 4.9 Intergranular porosity abundance versus depth of the four drill cores (DV10-001, BL-08-01, Rumpel Lake and WC-79-1). Note there is no systematic trend with respect to depth.
Fig. 4.10 Hematite abundance versus depth of the four drill cores (DV10-001, BL-08-01, Rumpel Lake and WC-79-1). Note there is no systematic trend with respect to depth.
Fig. 4.11 Matrix abundance versus depth of the four drill cores (DV10-001, BL-08-01, Rumpel Lake and WC-79-1). Note there is no systematic trend with respect to depth.
4.2.2 Compaction index point counting results

Contact index and tight packing index

Contact index (CI), tight packing index (TPI) and intergranular volume (IGV) were calculated from the point counting data which are tabulated in Table 4.2. The calculated results of CI and TPI were plotted together with the current depth for trend observation as illustrated in Fig. 4.12. It is observed that below the Wolverine Point Formation, the values of CI and TPI increase systematically with depth, which suggests that sandstones become more and more compacted from the Lazenby Lake Formation towards the Manitou Falls/Read Formation near the basement. However, the values of CI and TPI at the Locker Lake Formation above the Wolverine Point Formation are much higher than those at the Lazenby Lake Formation below, which suggests that the Locker Lake Formation is significantly compacted, as indicated by the dominance of concave-convex and sutured grain contacts.

Intergranular volume

Intergranular volume (IGV) is the sum of intergranular porosity, intergranular cement, and depositional matrix (Houseknecht, 1987; Paxton et al., 2002) and is an effective gauge of compaction degree by evaluating the proximity of grain frameworks (Houseknecht, 1987; Paxton et al., 2002). Paxton et al (2002) reported that the minimum
IGV is about 26% for well-sorted, quartz-rich sandstones that are compacted solely by mechanical compaction, while the additional decreasing to below 26% is mostly likely due to chemical compaction (intergranular pressure solution) that pushes framework grains to approach each other by dissolving materials at grain contacts.

As illustrated in IGV versus depth diagram (Fig. 4.13), the values of IGV of samples at the Locker Lake Formation are typically <20% (below 26%), suggesting that the Locker Lake Formation has been subjected to significant mechanical and chemical compaction. However, the values of IGV at the Wolverine Point and Lazenby Lake formations generally range from 25 to 30%, which indicates that chemical compaction (pressure solution) are not significantly developed in these sandstones, as petrographically indicated by the dominance of point- to null- contacts between detrital quartz grains. Below the Wolverine Point Formation, the values of IGV decrease systematically with depth towards the basement, and the sandstones near the basement typically have the lowest values of IGV generally below 15%, for example, at the Manitou Falls and Read formations, at which sandstones have undergone significant mechanical and chemical compaction.

Compaction porosity loss (COPL) and cementation porosity loss (CEPL) can be
calculated using the IGV and other parameters for evaluating the significance of compaction and cementation in reducing sandstone porosity using the following equations (Lundegard, 1992):

\[
\text{COPL} = P_i - \left( \frac{100 - P_i}{100 - P_{mc}} \right) P_{mc}
\]
\[
\text{CEPL} = (P_i - \text{COPL}) \left( \frac{C}{P_{mc}} \right)
\]

Where \( P_i \) is the initial porosity which is assumed to be 45% (Lundegard, 1992), \( P_{mc} \) is the intergranular volume, \( C \) is the total cement volume. The plot of COPL versus CEPL suggests that the porosity loss mainly results from compaction, with an average of COPL about 32.1% compared with an average of CEPL about 9.9% (Fig. 4.13). Besides, the IGV has a positive relationship with quartz cement abundance as illustrated in Fig. 4.14, which suggests that quartz cementation occurred before significant chemical compaction.
Table 4.2 Point counting data of sedimentary texture and compaction-related parameters of sandstone samples from the Rumpel Lake, WC-79-1, BL-08-01 and DV10-001 drill cores.

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<th>Grain Size (φ)</th>
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<th>IGV</th>
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<td>36.8</td>
<td>4.6</td>
<td></td>
</tr>
</tbody>
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Fm. = formation, Std Dev = standard deviation of the maximum diameters of the target grains, representing the sorting index, CI = contact index, TPI = tight packing index, IGV = intergranular volume, COPL = compaction porosity loss, CEPL = cementation porosity loss.
Fig. 4.12 Contact index & tight packing index versus depth of the four drill cores (DV10-001, BL-08-01, Rumpel Lake and WC-79-1). Note the systematic increase of CI and TPI with depth in the BL-08-01, Rumpel Lake and WC-79-01, whereas similar trend is not discerned from the DV10-001. The Locker Lake Formation above the Wolverine Point Formation is significantly compacted as illustrated by the high values of the CI and TPI.
Fig. 4.13 Intergranular volume versus depth of the four drill cores (DV10-001, BL-08-01, Rumpel Lake and WC-79-1). Note the systematic increase with depth in the BL-08-01, Rumpel Lake and WC-79-1 drill cores. Whereas the similar trend is not discerned in the DV10-001. The Locker Lake Formation above the Wolverine Point Formation is significantly compacted as indicated by the low values of intergranular volume.
Fig. 4.14 (A) Plot of compaction porosity loss (COPL) versus cementation porosity loss (CEPL); (B) Plot of intergranular volume versus quartz cement abundance. Data from all selected samples used in this study.
Fig. 4.15 Diagenetic well correlation panel through the studied four drill cores. Note that pervasive quartz cement is typically found at the Lazenby Lake Formation just below the mud-rich Wolverine Point Formation, while significant compaction is generally observed right above the basement and mud-rich Wolverine Point Formation. (Fm = Formation, Litho = lithology, QC = quartz cement abundance, IGV = intergranular volume, CI = contact index, TPI = tight packing index)
4.3 Relationships between quartz cementation and compaction

By plotting the results of compaction-related index and quartz cement abundance within the lithostratigraphic frameworks (Fig. 4.15), it was observed that pervasive quartz cementation occurs at the Lazenby Lake Formation, where sandstones are at a low degree of compaction as indicated by the low values of CI and TPI and high values of IGV. The samples that are poorly cemented by syntaxial quartz overgrowth were generally found at the Locker Lake, Manitou Falls and Read formations, where sandstones are at a high degree of compaction as indicated by the high values of CI, TPI and low values of IGV. The quartz cementation versus compaction pattern has the following characteristics: 1) the sandstones at the Lazenby Lake Formation below the mud-rich Wolverine Point Formation are pervasively cemented (>15%) by syntaxial quartz overgrowth, and these samples are the least compacted. 2) The samples at the Locker Lake, Manitou Falls and Read formations right above the low-permeability, mud-rich Wolverine Point Formation are poorly cemented (<5%) by authigenic quartz, and these samples are the most compacted. 3) Sandstones at the middle of the confined sandstones are moderately cemented by syntaxial quartz overgrowth. 4) Significant quartz cementation does not develop in matrix-rich sandstones, which are usually observed in the samples from DV10-001 drill core.
Chapter 5: Numerical Modeling Results

5.1 Scenario 1 - SiO₂ dissolution versus precipitation in confined sandstone unit at shallow burial, elevated thermal gradient conditions – one confined sandstone aquifer

The scenario 1 was modeled using the physical model 1 and was used to test the hypothesis that fluid convection occurred at a shallow burial, elevated thermal gradient condition, and controlled the development of quartz dissolution and precipitation pattern. The initial condition of this scenario is shown in Fig. 5.1. Modeling results show that thermal convection only prevailed within the confined sandstones, taking about 0.3 Ma to approach a steady state. At the steady state, eight evenly-spaced convection cells developed in the confined sandstones; each cell had an approximate width of 0.6 km as illustrated in Fig. 5.2. The direction of fluid flow is indicated by arrows orientation and the velocity is reflected by arrows size. Arrow arrangement indicates that fluid velocities were greatest near the boundaries between the confined aquifer and confining layers (the cover and basement), as well as at the center of the upwelling and downwelling plumes. The fluid velocity was about 1.5 m/year at the maximum, which decreased gradually from the boundary to the center of the cell. Flowing arrows indicate that convection flow cannot freely penetrate into the low-permeability cover and basement, indicating that
fluid exchange between the aquifer and aquitard can only maintain a very slow rate. Fluid velocities in the basement and cover were less than 0.01 m/year. The temperature regime of the sandstone unit was elevated due to fluid convection. Several high temperature and low temperature plumes developed due to the upwelling and downwelling flows. At a steady state, the concentration of SiO$_2$(aq) in sandstones ranged from 0.0019 mol/L at the top to 0.0032 mol/L at the bottom of the confined sandstones as illustrated in Fig. 5.2. Quartz dissolution versus precipitation pattern at 1 Ma was illustrated in Fig. 5.2. Pervasive quartz cementation occurred at the top of the confined sandstone compared to significant quartz dissolution at the bottom above the basement. The volumetric change of quartz with time (from 0 to 1 Ma time interval) due to dissolution and precipitation within representative grids (from grid A to F) are shown in Fig. 5.13. At the grid A located at the top of the upwelling zone, about 1.8% volumetric fraction of quartz precipitated. At the grid B located at the center of the upwelling zone, about 0.5% volumetric fraction of quartz precipitated. At the grid C at the bottom of the upwelling zone, about 4% volume fraction of quartz was dissolved. At the grid D at the top of the downwelling zone, 1.8% volume fraction of quartz precipitated. At the grid E at the center of the downwelling zone, 0.45% volume fraction of quartz was dissolved. At the grid F at the bottom of the downwelling zone, about 2.8% volume fraction of quartz was dissolved.
Fig. 5.1 Initial conditions of the physical model 1. (a) thermal profile; (b) SiO$_2$(aq) concentration profile. (c) quartz dissolution versus precipitation pattern. Note geothermal gradient was assigned to be 57°C/km and the permeability of the confined sandstone was assigned to be 100 md. Note also that positive values in (c) indicate quartz precipitation (cementation), and negative values indicate dissolution; the initial condition (shown by yellow color) is neither precipitation nor dissolution.
Fig. 5.2 Results of scenario 1 after 1 Ma of fluid convection. (a) thermal profile. (b) SiO$_2$(aq) concentration profile. (c) quartz dissolution versus precipitation pattern. Note also that positive values in (c) indicate quartz precipitation (cementation), and negative values indicate dissolution; the initial condition (shown by yellow color) is neither precipitation nor dissolution.
Fig. 5.3 Cell history plots of the observation grids illustrating quartz dissolution or precipitation at different parts of the model of scenario 1. At grid A, B and D, quartz precipitated and the amount of cement increases with time, whereas at C, E and F, quartz is dissolved and the amount of cement decreases with time.
5.2 Scenario 2 – SiO₂ dissolution versus precipitation at deep burial, normal thermal gradient conditions – one confined sandstone aquifer

The scenario 2 was designed to model quartz dissolution versus precipitation process at a deep (5 to 7 km) burial, normal thermal gradient (30°C/km) condition, coincident with the conventional diagenetic-hydrothermal U mineralization model (Hoeve and Sibbald, 1978). The initial condition of this scenario is shown in Fig. 5.4. Modeling results show that fluid convection took about 0.2 Ma to approach a steady state, and six evenly-spaced convection cells developed within the permeable, confined quartz-rich sandstones. Each convection cell had an approximate width of 0.8 km (Fig. 5.5). Maximum fluid velocity was about 1 m/year at the center of the upwelling and downwelling plumes as well as at the boundaries between the confined sandstones and the underlying and overlying confining layers. The concentration of SiO₂(aq) ranged from 0.00384 mol/L at the bottom to 0.0028 mol/L at the top of the aquifer as seen in Fig. 4.20. The volumetric change of quartz due to dissolution and precipitation in the six grids during the time period of 0 to 1 Ma is illustrated in Fig. 5.6.
Fig. 5.4 Initial conditions of physical model 2. (a) thermal profile; (b) SiO2(aq) concentration profile. (c) quartz dissolution versus precipitation pattern. Note geothermal gradient was assigned to be 30°C/km and the permeability of the in-between sandstone was assigned to be 50 md. Note also that positive values in (c) indicate quartz precipitation (cementation), and negative values indicate dissolution; the initial condition (shown by yellow color) is neither precipitation nor dissolution.
Fig. 5.5 Results of scenario 2 after 1 Ma fluid convection (a) thermal profile. (b) SiO$_2$(aq) concentration profile. (c) quartz dissolution versus precipitation pattern. Note also that positive values in (c) indicate quartz precipitation (cementation), and negative values indicate dissolution.
Fig. 5.6 Cell history plots of the observation grids illustrating quartz dissolution and precipitation in scenario 2. At grid A, B and D, quartz precipitates and the amount of cement increases with time, whereas at C, E and F, quartz is dissolved and the amount of cement decreases with time.
5.3 Scenario 3 – SiO\textsubscript{2} dissolution versus precipitation at shallow burial, elevated thermal gradient conditions – two confined sandstone aquifers

The scenario 3 was to model quartz dissolution versus precipitation at a deeply buried condition using the physical model 3 that was established based on the shallow burial model of unconformity-related U mineralization (Chi et al., 2018). The initial condition is shown in Fig. 5.7. Thermal convection prevailed within the two confined sandstone units and took about 0.1 Ma to approach a steady state. Eleven evenly-spaced convection cells developed and each convection cell had an approximate width of 0.45 km as illustrated in Fig. 5.8. The direction of fluid flow is indicated by arrows orientation and fluid velocities are indicated by the arrow size. Fluid velocities were greatest at the boundaries between the confined sandstones and confining layers (cover and basement), as well as in the middle center of the upwelling and downwelling plumes. The maximum fluid velocity was about 1.5 m/year at the lower aquifer and about 0.8 m/year at the upper aquifer. The fluid velocity decreased gradually from the boundary to the center of the cell. The quartz dissolution versus precipitation pattern developed in both the upper and lower aquifers as illustrated in Fig. 5.8. The temperature at both the upwelling and downwelling plumes, and the slope of thermal gradient became steep as the result of the development of thermal convection (Fig. 5.9).
Fig. 5.7 Initial conditions of physical model 3. (a) thermal profile, (b) SiO$_2$(aq) concentration profile, (c) quartz dissolution versus precipitation pattern. Note also that positive values in (c) indicate quartz precipitation (cementation), and negative values indicate dissolution; the initial condition (shown by yellow color) is neither precipitation nor dissolution.
Fig. 5.8 Results of scenario 3 after 1 Ma of fluid convection (a) thermal profile. (b) SiO$_2$(aq) concentration profile. (c) quartz dissolution versus precipitation pattern. Note geothermal gradient was assigned to be 57°C/km and the permeability of the lower sandstone was assigned to be 100 md and the upper sandstone was assigned have permeability of 200 md. Note also that positive values in (c) indicate quartz precipitation (cementation), and negative values indicate dissolution.
Fig. 5.9 Temperature profile obtained from reactive mass transport modeling of scenario 3. Grey line indicates the initially assigned temperature gradient. The blue line indicates the temperature gradient at the center of downwelling plume. The red line indicates the temperature gradient at the center of the upwelling plume. Note temperature is elevated and the slope of thermal gradient becomes steep at both upwelling and downwelling plumes.
5.4 Scenario 4 – SiO$_2$ dissolution versus precipitation at deep burial, normal thermal gradient conditions – two confined sandstone aquifers

The scenario 4 was used to model quartz dissolution versus precipitation within two confined sandstone packages at a deep burial, normal thermal gradient condition, corresponding to the conventional deep-burial model of diagenetic-hydrothermal conditions. The initial condition is shown in Fig. 5.10. Free convection cells developed both at the upper and lower confined sandstone units and it took about 0.1 Ma to approach a steady state. At the steady state, five evenly-spaced convection cells developed in the lower sandstone units, while six convection cells developed in the upper sandstone units (Fig. 5.11). The direction of fluid flow is indicated by the orientation of arrows and fluid velocities are indicated by the size of arrows. Fluid velocities were greatest at the boundaries between the confined aquifer and at the middle center of the upwelling and downwelling plumes. The maximum velocity at the lower aquifer was about 1 m/year, and it was about 0.5 m/year at the upper aquifer. Modeling results show that the quartz dissolution versus precipitation patterns in both the upper and lower confined sandstones had the similar geometric characteristics. Temperature were elevated due to thermal convection, and the slope of thermal gradient became much steeper (Fig. 5.12).
Fig. 5.10 Initial conditions of physical model 4. (a) thermal profile. (b) SiO$_2$(aq) gradient. (c) quartz dissolution versus precipitation pattern. Note geothermal gradient was assigned to be 30°C/km and the permeability of the lower sandstone unit was assigned to be 50 md and that of the upper sandstone unit is assigned to be 100 md. Note also that positive values in (c) indicate quartz precipitation (cementation), and negative values indicate dissolution; the initial condition (shown by yellow color) is neither precipitation nor dissolution.
Fig. 5.11 Results of physical model 4 after 100 Ma fluid convection. (a) thermal profile. (b) SiO\textsubscript{2}(aq) concentration profile. (c) quartz dissolution versus precipitation pattern. Note also that positive values in (c) indicate quartz precipitation (cementation), and negative values indicate dissolution.
Fig. 5.12 Temperature profile obtained from reactive mass transport modeling of scenario 4. Grey line indicates initially assigned temperature gradient. Blue line indicates temperature gradient at the center of downwelling plume. Red line indicates temperature gradient at the center of the upwelling plume. Note temperature is elevated and the slope of thermal gradient becomes steep at both upwelling and downwelling plumes.
5.5 Sensitivity study

Sensitivity study is to test how various sources of uncertainty in the mathematical model contribute to the overall uncertainty of the modeling results. In this study, the most important input that was uncertain and might significantly affect the results is the permeability. Thus, the influence of the permeability must be carefully evaluated.

Permeability, as an important factor influencing the value of the Rayleigh number, may determine whether thermal convection can occur or not in porous media (Turcotte and Schubert, 2002). However, the permeability of the confined sediments of the Athabasca Group is not a constant value, which cannot be accurately estimated due to its prolonged burial and diagenetic history. Two basic principles have been considered when assigning the permeability to the physical models during the sensitivity study, and that are: 1) Relatively shallow-buried sediments do not compact as much as the deeply-buried counterparts, and therefore should be assigned with a higher permeability. 2) Under normal circumstances, quartz-dominated sandstones have higher porosity and permeability due to the higher resistance to mechanical and chemical compaction compared with feldspatic sandstones and lithic sandstones. So a series of permeability values were tested to ensure the validity of the quartz dissolution versus precipitation patterns yielded from reactive transport modeling. Hereby, another six modeling results
of temperature regime and quartz dissolution versus precipitation pattern are shown below. These modeling results were yielded by assigning different permeability values to the sandstone aquifers with two contrasting thermal gradients that are normal thermal gradients of 30°C/km and elevated thermal gradients of 58 °C/km. Comparing the model results with different assignment of permeability and thermal gradient enables researchers to understand the influence of permeability and thermal gradients on geometry of convection cells, fluid flowing rate, and quartz dissolution versus precipitation pattern.
Fig. 5.13 Modeling results after 1 Ma when initial thermal gradient was assigned to be 57 °C/km, the permeability of the basement and mud-rich aquitard was assigned to be 0.01 md, and the permeability of the confined sandstone aquifer was assigned to be 25 md. (a) thermal profile. (b) quartz dissolution versus precipitation pattern. Note that positive values in (b) indicate quartz precipitation (cementation), and negative values indicate dissolution; the initial condition (shown by yellow color) is neither precipitation nor dissolution.
Fig. 5.14 Modeling results after 1 Ma years when initial thermal gradient was assigned to be 30 °C/km, the permeability of the basement and mud-rich aquitard was assigned to be 0.01 md, and the permeability of the confined sandstone aquifer was assigned to be 25 md. (a) thermal profile. (b) quartz dissolution versus precipitation pattern. Note that positive values in (b) indicate quartz precipitation (cementation), and negative values indicate dissolution; the initial condition (shown by yellow color) is neither precipitation nor dissolution.
When the initial thermal gradient was assigned to be 57°C/km, the permeability of the basement, confined aquifer and aquitard was assigned to be 0.01 md, 25 md and 0.01 md separately (compared with scenario 1 with the permeability of the aquifer assigned to be 100 md and other conditions to be the same), thermal convection occurred and the associated quartz dissolution versus precipitation pattern developed as illustrated in Fig. 5.13. Six convection cells developed in the confined aquifer with major quartz cementation occurring at the top and dissolution at the bottom (Fig. 5.13). Compared with scenario 1 in which eight convection cells developed, though thermal convection still developed, the decrease of the permeability of the confined aquifer from 100 md to 25 md results in the decrease of the number of convection cells from eight to six, and the flow rate and quartz dissolution and/or precipitation rate slow down. Once the thermal gradient decreases to 30°C/km, thermal convection occurred and five convection cells developed in the confined aquifer with major quartz cementation occurring at the top and dissolution at the bottom (Fig. 5.14). Compared with the results of Fig. 5.13 in which six convection cells developed, though thermal convection and the associated quartz dissolution versus precipitation pattern still developed, the decrease of the initial thermal gradient from 57°C/km to 30°C/km results in the decrease of the number of convection cells from six to five and the slowdown of the rate of fluid flow and quartz dissolution and precipitation.
Fig. 5.15 Modeling results after 1 Ma years when initial thermal gradient was assigned to be 57°C/km, the permeability of the basement and mud-rich aquitard was assigned to be 0.01 md, and the permeability of the confined sandstone aquifer was assigned to be 200 md. (a) thermal profile. (b) quartz dissolution versus precipitation pattern. Note that positive values in (b) indicate quartz precipitation (cementation), and negative values indicate dissolution; the initial condition (shown by yellow color) is neither precipitation nor dissolution.
Fig. 5.16 Modeling results after 1 Ma years when initial thermal gradient was assigned to be 30 °C/km, the permeability of the basement and mud-rich aquitard was assigned to be 0.01 md, and the permeability of the confined sandstone aquifer was assigned to be 200 md. (a) thermal profile. (b) quartz dissolution versus precipitation pattern. Note that positive values in (b) indicate quartz precipitation (cementation), and negative values indicate dissolution; the initial condition (shown by yellow color) is neither precipitation nor dissolution.
When the initial thermal convection was assigned to be $57^\circ$C/km, the permeability of the basement, confined sandstone aquifer and aquitard was assigned to be 0.01 md, 200 md and 0.01 md (compared with scenario 1 with the permeability of the aquifer assigned to be 100 md and other conditions to be the same). It is observed that eight convection cells developed in the confined aquifer with major quartz cementation occurring at the top and dissolution at the bottom (Fig. 5.15). Due to the increase of the permeability of the confined aquifer from 100 md to 200 md, the flow rate of the thermal convection and the quartz dissolution/precipitation rate speed up. Once thermal gradient was decreased to $30^\circ$C/km with other parameters staying unchanged, thermal convection occurred as illustrated in Fig. 5.16. It is observed that convection cells developed in the confined aquifer with major quartz cementation occurring at the top and dissolution at the bottom (Fig. 5.16). Compared with Fig. 5.15, due to the decrease of the initial thermal gradient from $57^\circ$C/km to $30^\circ$C/km, the flow rate of the thermal convection and the quartz dissolution/precipitation rate decrease.
Fig. 5.17 Results after 1 Ma years when initial thermal gradient was assigned to be 30 °C/km, the permeability of the basement and mud-rich aquitard was assigned to be 0.01 md, and the permeability of the confined sandstone aquifers was assigned to be 5 md. (a) thermal profile. (b) quartz dissolution versus precipitation pattern. Note that no quartz dissolution and precipitation pattern is discerned.
Fig. 5.18 Results after 1 Ma years when initial thermal gradient was assigned to be 30 °C/km, the permeability of the basement and mud-rich aquitard was assigned to be 0.01 md, and the permeability of the confined sandstone aquifers was assigned to be 10 md. (a) thermal profile. (b) quartz dissolution versus precipitation pattern. Note that no quartz dissolution and precipitation pattern is discerned.
When the initial thermal convection was assigned to be 30°C/km, the permeability of the basement, confined sandstone aquifer, aquitard was assigned to be 0.01 md, 5 md and 0.01 md, thermal convection did not occur as illustrated in Fig. 5.17. Compared with scenario 2 in which six convection cells developed, thermal convection and the quartz dissolution versus precipitation pattern did not develop in Fig. 5.17 due to the decrease of the permeability of the confined aquifer from 50 md to 5 md, which might result in the decrease of the Rayleigh number to less than the critical Rayleigh number. When the permeability of the confined sandstone aquifers was increased to be 10 md which is a relatively low permeability for quartz arenite that is generally considered to have a high permeability, obvious thermal convection only occurred in the lower confined aquifer as illustrated in Fig. 5.18. Significant quartz dissolution and precipitation pattern did not well develop within 1 Ma time, which may be due to the relatively low velocity of convection flow.

The sensitivity study show that thermal convection cannot occur if the permeability of the confined sandstone aquifer is too low, and as long as thermal convection occurs, the quartz dissolution versus precipitation will inevitably develop. Higher thermal gradient and higher permeability of confined sandstone aquifers are favourable for the development of the quartz cementation and dissolution pattern.
Chapter 6: Discussion and conclusions

6.1 The controlling factors of compaction and cementation of sandstones in the Athabasca Group

Sedimentary fabric and composition, represented by sorting and matrix abundance, may have exerted a strong control on the compaction and cementation characteristics of sandstones as mentioned by other case studies (e.g. Paxton et al., 2002). Thus, poorly-sorted and matrix-rich sandstones are not good indicators of compaction as discussed in section 3.1.5. After excluding poorly-sorted and matrix-rich samples from the point counting study, based on the petrographic observation and point counting results, a compaction and cementation pattern that is similar to the quartz dissolution versus precipitation pattern yielded from reactive mass transport modeling of SiO$_2$ was recognized. The similarity between the petrographic characteristics and modeling results suggests that the compaction and cementation characteristics of the quartozse sandstones in the Athabasca basin may be fundamentally controlled by a large-scale fluid convection that occurred in the confined sandstone aquifers. Another petrographic observation is that most grain contacts are point contacts at the Lazenby Lake Formation (at the top of the confined aquifer) with quartz cement abundance up to 30%. It indicates that quartz cement was introduced at a shallow burial depth before the sandstones were significantly
compacted, because deeply buried sediments are unlikely to have such high pre-cemented porosity. On the other hand, the poorly-cemented sandstones that mainly occur at the Locker Lake/Manitou Falls/Read formations (at the bottom of the confined aquifer) are significantly compacted as indicated by the dominance of sutured grain contacts (highly compacted characteristics). The IGV versus depth plots show a similar pattern as illustrated in Fig. 4.15. In theory, the IGV decreases as the result of mechanical and chemical compaction gradually with the burial depth increasing. For quartz-dominated, matrix-less and well-sorted sandstones, mechanical compaction reduces intergranular volume from initially 40% to 26% (Lundegard, 1992, Paxton et al., 2002). It is assumed that 26% is the minimum value that can be approached by pure mechanical compaction and the further decrease to below 26% is caused by chemical compaction. In this study, the measured IGVs that are less than 30% for all samples suggest that the all the sandstones have been subjected to major mechanical compaction. For samples with IGV < 26% (or even <20%), significant chemical compaction have occurred as shown by a sutured fabric which indicates a high degree of diagenetic maturity. The IGV versus depth plots show a similar trend in the four drill cores. That is, IGV decreases systematically from relatively high values (mainly >26%) at the Lazenby Lake Formation to extremely low values (<20% or even <15%) at the Manitou Falls/Read formations (Fig. 4.15). Correspondingly, quartz cement abundance versus depth plots show a trend that quartz
cement abundance decreases with depth from an extremely high value (mainly >20%) at the Lazenby Lake Formation to an extremely low value (mainly <5%) at the Manitou Falls/Read formations (Fig. 4.15).

Based on the petrographic observation and point counting data, the compaction degree of sandstones are strongly related to both the quartz cementation and depth in the Athabasca Basin. To interpret the relationship between compaction, cementation and depth, a diagenetic model that incorporates mechanical compaction, chemical compaction, quartz cementation and quartz dissolution have been established. This diagenetic model is able to explain the quartz cementation process, including silica supply, silica transport and silica precipitation, and their control on compaction characteristics. In terms of silica supply, various sources of silica including clay minerals transformation (smectite to illite transformation), feldspar alteration, surface weathering of unstable silicate, intergranular pressure solution and stylolitization are potential candidates (McBride, 1991). In the study area, clay minerals are lacking in most fluvial/alluvial sandstones with the exception of the marine-face Wolverine Point Formation. The Wolverine Point Formation that is composed of large amounts of clay may be able to produce considerable amount of silica through smectite illitizationion, but it is unlikely that silica produced within the clay-rich Wolverine Point Formation can be largely transported outward into the adjacent
sandstones because the mud-rich Wolverine Point Formation is a relatively closed system which is unfavourable for the expelling of silica. Though several literatures suggest that shale diagenesis can be an open system (Lynch, 1997; Lynch et al., 1997), the viewpoint that the Wolverine Point Formation is the major source of silica cannot account for the contrasting cementation features of the underlying Lazenby Lake (well-cemented) and the overlying Locker Lake (poorly-cemented) formations. The possibility that silica was derived from surficial weathering of unstable silicate is also ruled out, as supported by the results of the previous fluid inclusion analysis of Chu and Chi (2016); the high temperature (>120°C) and high salinity (>20 wt.%) indicated by fluid inclusion studies on syntaxial quartz overgrowth suggest that the diagenetic fluid responsible for quartz cementation is originated from diagenetic brines rather than meteoric water. Feldspar dissolution may be a contributor of silica supply, but the contribution cannot be quantified; Oversized dissolution pores are not observed, which may indicate that the feldspar leaching is not the major source of silica. Therefore, the most probable silica source is intergranular pressure solution and stylolites as supported by the petrographic observation discussed in the above chapter. In terms of precipitation, nucleation sites are abundantly available for the precipitation of quartz cement so that it is not an important control on the precipitation of quartz cement within matrix-less fluvial/alluvial quartz arenite. While for matrix-rich quartz wacke, detrital clay matrix covering the surface of
grains significantly reduces the nucleation areas and thus inhibits quartz cementation, which is a commonly seen phenomenon in the DV10-001 drill cores.

Given the extremely low solubility of quartz in diagenetic environments (< 250°C), large volumes of fluid flux are necessary for the transport of dissolved quartz (Blatt, 1979). In relatively deep burial (> 2 km) condition after significant mechanical compaction, the mechanisms that are available to generate fluid circulation and transport large volumes of silica are rare. Meteroric water flux is able to recharge sufficient amounts of fluid into sandstone reservoirs (Bjørlykke, 1994), but this mechanism cannot account for the quartz cementation in the Athabasca Basin, because the origin of diagenetic fluids within quartz overgrowth is high-salinity (>20%) brine rather than meteoric water (Chu and Chi, 2016). Therefore, considering the geological environment of the Athabasca Basin and numerical modeling results from reactive mass transport modeling, the most probable mechanism for silica transport within the Athabasca Basin is fluid convection. Fluid convection may have driven the circulation of high-salinity brines within the basin at a specific geological period, and at that time the quartz cementation and dissolution pattern developed. It can be envisaged that within a confined sandstone aquifer, quartz dissolution occurs at the bottom compared with quartz precipitation happen at the top during fluid convection. The reactive mass transport modeling suggests that quartz dissolution versus precipitation
process in the Athabasca Basin might be under a local chemical equilibrium condition, and in this case the precipitation and dissolution of quartz in each small grid are controlled by net silica import. If silica import was positive (as exemplified in the upwelling plumes), pore fluid within the grid became oversaturated with silica, resulting in quartz precipitation. In the opposite situation, if silica import was negative (as exemplified in the downwelling plumes), pore fluids became undersaturated with silica, resulting in quartz dissolution and the enhancement of chemical compaction. Modeling results also suggest that there may have existed two thermal convection systems within the known stratigraphic framework. The lower confined sandstone unit that is composed of the Lazenby Lake, Manitou Falls and Read formations may have accommodated a convection system that is confined by the overlying Wolverine Point Formation and the underlying basement; at stratigraphically higher parts of the basin, the confined sandstone unit that is composed of Locker Lake and Otherside formations may have accommodated another convection system that is confined by the overlying Douglas Formation and the underlying basement. The sandstones at the Locker Lake Formation are significantly compacted, which indicates that the Locker Lake Formation was once located at the bottom of the upper convection system. The Otherside Formation which is interpreted to be located at the top of the upper convection system has been fully eroded at the studied area. Finally, a schematic diagenetic model has been established to interpret the formation
mechanisms of the compaction and cementation characteristics of sandstones in the Athabasca Basin. It shows that the compaction and cementation characteristics of sandstones are fundamentally controlled by a large-scale fluid convection (Fig. 6.1).
Fig. 6.1 Schematic illustrating the compaction and cementation of sandstones are associated with large-scale fluid convection based on petrographic observation and reactive mass transport modeling. Note that pervasive quartz cement occurs at the Lazenby Lake Formation below the mud-rich Wolverine Point Formation. By contrast, the Locker Lake and Read formations at the bottom of the convection cells are significantly compacted as indicated by large numbers of sutured grain-grain contacts.
6.2 Implications for large-scale fluid flow, temperature regime and hydrological evolution in the Athabasca Basin

Proterozoic basins, as exemplified by the Athabasca Basin, that are predominantly filled by alluvial/fluvial sandstones and gravels with very few muddy sediments are fundamentally different from Phanerozoic counterparts with respect to sedimentological and stratigraphic characteristics as the result of the lack of terrestrial plants on earth surface at that time (Schumm, 1968; Hiatt et al., 2007). Proterozoic alluvial and fluvial sediments reaching thicknesses up to even more than a thousand meters can be regarded as large-thickness, high-permeability groundwater reservoirs. On the other hand, the extensively-distributed, muddy marine sediments serve as low-permeability aquitards that confine the large-thickness, fluvial/alluvial sandstone packages. This type of close/semi-closed hydrological configuration is favorable for the occurrence of free thermal convection during stable tectonic settings as supported by previously published numerical modeling results (Raffensperger and Garven, 1995; Cui et al., 2012; Li et al., 2016).

This study firstly proposed that the quartz dissolution versus precipitation pattern recognized in this study may be fundamentally controlled by thermal convection as
mentioned in other studies (Wood and Hewett, 1982; Wood, 1986). Though some scholars argued that large-scale fluid convection cannot be a dominant process controlling the diagenesis of the sedimentary rocks due to the frequent occurrence of mudstone layers which baffle or even prohibit fluid convection, except local-scale convection that may happen near igneous intrusions, salt domes or active faults (Bjørlykke et al., 1988), the Athabasca Basin and its Proterozoic analogies are composed of nearly 100% quartz-rich sandstones with very few mud interstitial materials, and in which fluid convection may have prevailed and controlled the diagenetic process. In addition, the marine-face sediments of low permeability are generally extensively developed and may serve as spatially and temporally stable confining layers. Thus, this specific type of hydrological configuration enables fluid convection to take place in high-permeability sandstone aquifers, which is supported by this study and previous numerical modeling studies (Raffensperger and Garven, 1995; Cui et al., 2012; Li et al., 2016). In the marine-face Wolverine Point and Douglas formations that contain high abundance of sedimentary matrix, convection fluid flow cannot penetrate into these low-permeability barriers as illustrated by the flow arrows in the numerical modeling results, in which layers heat and mass transport can only take place through thermal conduction and molecule diffusion. Given thermal convection as a much more efficient method compared with thermal conduction in transporting heat from the deep part to the
shallow part of the basin, which can elevate the temperature of the whole basin and result in the development of anomalously steep thermal gradients as recorded within the fluid inclusions (Chu and Chi, 2016). Besides, due to temperature difference within convection cells, minerals with solubility as a function of temperature, e.g. quartz or calcite, are inevitably dissolved at the hot areas, transported by convection flow, and then precipitate at other parts of the basin, which results in the formation of specific mineral dissolution versus precipitation patterns (Wood and Hewett, 1982).

As for the hydrological properties evolution, the top part of of the confined sandstones might be blocked by quartz overgrowth at the early stage of the fluid convection; significant chemical compaction promoted by dissolution occurred at the bottom. The dissolved silica at grain-grain contacts was removed by convection flow so that no quartz cement precipitated in situ, by which way the sandstones that are poorly cemented and significantly compacted formed.
6.3 Timing of compaction, quartz cementation and fluid convection

Compaction degree of sandstones that are cemented by syntaxial quartz overgrowth indicates the relative time (burial depth) of quartz cementation (Molenaar, 1986; Bjørlykke and Egeberg, 1993; Hiatt, 2007). A small quantity of (5%) syntaxial quartz overgrowth can significantly retard or even absolutely prevent subsequent compaction because quartz cement bonds detrital grains and enhances the mechanical stability of detrital frameworks (Molenaar, 1986; Bjørlykke and Egeberg, 1993; McBride, 1989; McBride et al., 1991; Hiatt, 2007). Assuming that quartz cementation is associated with thermal convection, the low-degree of compaction of the Lazenby Lake Formation indicates thermal convection responsible for quartz cementation occurred when the sandstones at the Lazenby Lake Formation is not significantly compacted (prior to chemical compaction stage). Whereas, the sandstones that are significantly compacted at the bottom of the confined aquifer (Locker Lake, Manitou Falls or Read formations) may develop due to the coupled compaction and dissolution associated with thermal convection, which indicates the maximum burial depth of the Athabasca Basin may be >5-7 km. Solely based on this study, an accurate maximum burial depth of the basin can not be given and the questions about the maximum burial depth of the basin are in need of further investigation.
6.4 Implications for the unconformity-related U mineralization

Unconformity-related U mineralization in the Athabasca Basin is generally considered to be associated with protracted circulation of U-rich brines between the basin and the basement because brine circulation is critical for uranium leaching, transportation and concentration (Kyser et al., 1990; Kotzer et al., 1995). Thermal convection due to normal geothermal gradient has been proposed as the driving force of brine circulation responsible for U mineralization by several numerical modeling studies (Raffensperger and Garven, 1995; Cui et al., 2012; Li et al., 2015; Chu and Chi, 2016). However, there is no unequivocal geological evidence supporting the validity of these numerical modeling results. This study characterized the compaction and cementation pattern of sandstones and succeed in linking the petrographic pattern with thermal convection. Combined with previous fluid inclusion studies reporting that the homogenization temperature of fluid inclusions at syntaxial quartz overgrowths generally ranges from 130 to 200°C (Pagel, 1975; Kotzer and Kyser, 1995; Chu and Chi, 2016), this study suggests that fluid convection may have occurred at a shallow burial (<3 km) diagenetic environment with elevated geothermal gradients. In combination with the study that reported a discovery of U-rich fluid inclusions at syntaxial quartz overgrowths (Chi et al., 2019), this study further suggests that basinal quartz cementation, fluid convection and uranium
mineralization may be genetically related.

All in all, this study supported the shallow-burial genetic model of URU mineralization in which thermal convection have driven fluid circulation within quartzose successions, leached uranium from U-rich minerals that were distributed throughout the basin (Hoeve and Sibbald, 1978; Kotzer and Kyser, 1995; Komninou and Sverjensky, 1996; Fayek and Kyser, 1997), or infiltrated into the basement and extracted uranium from there (Hetch and Cuney, 2000; Derome et al., 2005; Richard et al., 2010), and finally transported U-rich brines to specific parts of the basin where favourable precipitation conditions were met. The formation of the world-class uranium deposits in the Athabasca Basin may be genetically linked to a geodynamic event that provided an anomalously higher-than-normal thermal disturbance from the deep part of the earth and heated the whole basin to an anomalously high (>200 °C) temperature at the time of shallow burial depth of the Athabasca Basin (Chi et al., 2018).
6.5 Conclusions

This study means to reveal the diagenetic background (especially the burial depth) of the Athabasca Basin via petrographic analysis and reactive mass transport modeling. Though compaction and cementation pattern of sandstones with respect to depth is characterized from the petrographic study, the maximum burial depth of the Athabasca Basin cannot be estimated solely based on this study. However, other scientific implications have been obtained. Firstly, this study firstly points out the potential linkage between the compaction and cementation pattern and fluid convection. That is, the sandstones that are pervasively cemented by quartz overgrowth at the top of the confined sandstone layers (Lazenby Lake Formation) are interpreted to be located at the top of a convection cell, while the sandstone at the bottom of confined aquifers (Locker Lake Formation, Manitou Falls and Read formations) are poorly cemented by authigenic quartz cement. In combination with the previous fluid inclusion studies reporting that quartz overgrowths host uraniferous and high-salinity brines, this study further suggests that fluid convection may have driven the uraniferous brine circulation during the shallow-burial (<3 km) stage of the Athabasca basin and made significant contributions to uranium leaching from either the basin or the basement.
6.6 *Recommendations for future studies*

Firstly, this study has given a petrographic evidence, a compaction and cementation pattern of quartzose sandstones, to support the hypothesis that fluid convection have taken place and contributed to the uranium mineralization in the Athabasca Basin. Future studies can be done to find more geological evidence (e.g. clay mineral assemblages, trace elements profiles at the syntaxial quartz overgrowth) to further support this hypothesis. As a significant mark of alteration halos enveloping uranium deposits, quartz dissolution and precipitation were assumed to be genetically associated with uranium deposits (Jefferson et al., 2007). Therefore, comparing the quartz dissolution and precipitation pattern at both mineralization areas and barren places may give more detailed information about hydrodynamic mechanisms and geochemical reactions associated with the unconformity-related uranium mineralization.

Secondly, this study proposes that major quartz cementation occurred when the basin was at a shallow burial depth, which indicates that fluid movement at the time of thermal convection were mainly controlled by the primary hydrological configuration. Therefore, the further refinement of three-dimensional models of sedimentary lithofacies may be important to understand and reestablish the hydrological system at the time of
unconformity-related U mineralization.

Thirdly, this study has a universal significance to the study on subsurface geofluid systems. The significance of large-scale thermal convection is usually underestimated in Phanerozoic basins considering the frequently-occurred mud-rich layers that significantly decrease the vertical permeability critical to thermal convection (Bjørlykke et al., 1988). However, it is not sure whether thermal convection occurred in some basins that have undergone significant thermal disturbance, e.g. rifting events. Potential candidates should be further studied in detail in terms of diagenetic patterns because understanding whether thermal convection has occurred or not is significant for understanding the transport and distribution of subsurface fluid of economic significance, such as underground water and hydrocarbon. Besides, understanding the mechanism of thermal convection may broaden the current knowledge of reservoir development, ore deposits formation and geothermal resources exploration in sedimentary basins.
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