Formation mechanisms of unconformity-related uranium deposits: insights from numerical modeling

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Formation mechanisms of unconformity-related uranium deposits: insights from numerical modeling

by

Tao Cui

A Dissertation
Submitted to the Faculty of Graduate Studies
through the Department of Earth and Environmental Sciences
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the Degree of Doctor of Philosophy at the
University of Windsor

Windsor, Ontario, Canada
2012
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Formation mechanisms of unconformity-related uranium deposits: insights from numerical modeling

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I. Co-Authorship Declaration

I hereby declare that this thesis incorporates material that is the result of joint research, as follows:

The collaboration occurs in Chapter 2, 3 and 4 of this dissertation. These chapters contain material that has been published or submitted for publication, and are coauthored with Jianwen Yang and Iain M. Samson. Drs. Yang and Samson provided advice on experiment designs, result analyses and paper writing. The key ideas, primary contributions, experimental designs, and data interpretation, however, were performed by the author of this thesis.

I am aware of the University of Windsor Senate Policy on Authorship and I certify that I have properly acknowledged the contribution of other researchers to my thesis, and have obtained written permission from each of the co-author(s) to include the above material(s) in my thesis.

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II. Declaration of Previous Publication

This thesis includes 3 original papers that have been previously published/submitted for publication in peer-reviewed journals, as follows:

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Chapter 4

Cui, T., Yang, J., and Samson, I. M., 2012, Uranium transport across basement/cover interfaces by buoyancy-driven thermohaline convection: implications for the formation of unconformity-related uranium deposits: ready to submit to American Journal of Science

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Abstract

A series of numerical experiments based on the finite element and finite difference modelings have been carried out to investigate ore-forming fluid systems related to uranium mineralization. Conceptual models were constructed by integrating important hydrogeological features shared by the Athabasca, Thelon and Kombolgie basins. Based on these conceptual models, various numerical scenarios were designed to investigate the interaction among fluid flow, heat transport, topographic relief and tectonic deformation. Equations governing these processes were solved by FEFLOW and FLAC. The modeling suggests that buoyancy-driven thermohaline convection develops in thick sandstone sequences at any geothermal gradient of 25 to 35 °C/km during periods of tectonic quiescence. Thermohaline convection may penetrate into the basement for up to 1-2 km below the basal unconformity when typical hydrological parameters for these Proterozoic hydrogeological units are used. Fluid flow velocities in the sandstone sequence are several orders of magnitude larger than those in the basement. If a uranium source is assumed to be located in the center of the basin below the unconformity, uranium is able to gradually spread into the sandstone through thermohaline convection. The location of the uranium source also affects the solute transport efficiency. Given appropriate hydrological conditions, thermohaline convection could have caused widespread interaction of basinal brines with basement rocks or basement-derived fluids in uranium-bearing Proterozoic basins, and that enough uranium could have been leached from the uranium-rich basement to form large, high-grade unconformity-related uranium deposits. Reactivation of preexisting basement structures and the generation of new faults suppress free convection and lead to deformation-dominated fluid flow or mixed convection,
depending on strain rates. During compressive deformation, reduced brines in the basement may be forced out along fractured zones and encounter uranium-bearing fluids in the clastic sequence to form sandstone-hosted deposits. By contrast, basement-hosted deposits are likely to form during extension, when oxidized basinal brines flow into faulted structures to interact with reduced minerals or fluids in the basement. Thus, the combined effect of thermohaline convection and tectonic deformation leads to the development of unconformity-related uranium deposits at intersections of the basal unconformity with faults or shear zones.
Acknowledgements

This study would not have been accomplished without the help of many individuals, whom I would like to thank.

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Table of Contents

Declaration of Co-Authorship / Previous Publication..........................................................III
Abstract.............................................................................................................................V
Acknowledgements..........................................................................................................VII
List of tables.....................................................................................................................XII
List of figures..................................................................................................................XIII
List of abbreviations.......................................................................................................XVI
List of symbols..............................................................................................................XVII

Chapter 1 Introduction........................................................................................................1
  1.1 Introduction.............................................................................................................1
  1.2 Background and literature review..........................................................................6
      1.2.1 Uranium transport and deposition in aqueous systems..................................6
      1.2.2 Geological setting of URU deposits...............................................................9
  1.3 Methodology and principles....................................................................................14
  1.4 Organization of thesis............................................................................................17
References.......................................................................................................................22

Chapter 2 Numerical modeling of hydrothermal fluid flow in the Paleoproterozoic
  Thelon basin, Nunavut, Canada (thermally driven free convection model)..................30
  2.1 Introduction.............................................................................................................30
  2.2 Development of conceptualized model.................................................................31
      2.2.1 Integration and analysis of existing data.......................................................31
      2.2.2 Paleo-hydrostratigraphic model.................................................................35
  2.3 Numerical experiments and results........................................................................37
      2.3.1 Basic model..................................................................................................37
      2.3.2 Effect of unconformity..................................................................................38
      2.3.3 Effect of stratigraphic heterogeneity...........................................................41
      2.3.4 Effect of topography.....................................................................................41
      2.3.5 Effect of faults.............................................................................................43
Chapter 3 Tectonic deformation and fluid flow: implications for the formation of unconformity-related uranium deposits (thermo-hydro-mechanical model)........53

3.1 Introduction......................................................................................................53
3.2 Geological framework.....................................................................................55
  3.2.1 The Athabasca basin.........................................................................55
  3.2.2 The Thelon basin...............................................................................58
  3.2.3 The Kombolgie basin........................................................................59
  3.2.4 Far-field orogenic events and fault reactivation...............................61
3.3 Conceptual model............................................................................................62
3.4 Theoretical background...................................................................................66
3.5 Results..............................................................................................................74
  3.5.1 Extensional deformation only (model 1) .........................................74
  3.5.2 Compressive deformation only (model 2) .......................................75
  3.5.3 Thermally-driven convection (model 3) ..........................................76
  3.5.4 Coupling of thermally-driven convection and extensional
deformation (model 4) ..................................................................78
  3.5.5 Coupling of thermally-driven convection and compressive
deformation (model 5) ..................................................................79
3.6 Discussion........................................................................................................81
  3.6.1 Driving mechanisms of fluid flow....................................................81
  3.6.2 Implications for uranium sources and deposit genesis.....................85
  3.6.3 Limitations........................................................................................88
3.7 Conclusions......................................................................................................89
References..............................................................................................................92

Chapter 4 Uranium transport across basement/cover interfaces by buoyancy-driven
termohaline convection: implications for the formation of unconformity-related
uranium deposits....................................................................................................100

4.1 Introduction.................................................................................................100
4.2 A brief introduction of URU deposits and their geological setting.........102
List of Tables

Table 1.1. Aqueous complexes of uranium.................................................................7

Table 1.2. Equations governing groundwater flow, heat transport and reactive mass transport.................................................................................................................18

Table 2.1. Major hydrogeological parameters of various units in Chapter 2..............36

Table 2.2. Maximum average linear velocity for various numerical cases.................39

Table 3.1. Major hydrogeological parameters of various units in Chapter 3..............63

Table 4.1. Major hydrogeological parameters of various units in Chapter 4...............111
List of Figures

Fig. 1.1. Two end-member genetic models for the unconformity-related uranium deposits....................................................................................................................3
Fig. 1.2. Regional geological map showing the locations of the Athabasca and Thelon basins and major uranium deposits in them............................................................4
Fig. 1.3. Simplified geological map showing important structures surrounding the Kombolgie basin and sites of major uranium deposits...........................................5
Fig. 1.4. The flow chart of solving a scientific problem by numerical modeling.................................................................16
Fig. 2.1. Map of the Thelon basin with locations of seismic exploration sites and generalized paleocurrent directions...........................................................................32
Fig. 2.2. Paleo-hydrostratigraphic model for the Thelon basin.................................................................33
Fig. 2.3. Modeling results for the basic case..............................................................................................38
Fig. 2.4. Fluid-flow patterns when a high-permeability unconformity zone is considered..............................................................................................................40
Fig. 2.5. Modeling results when diagenetic aquitards are incorporated..............................................42
Fig. 2.6. Streamline configuration and Darcy flux field for water table slopes of 0.0005 and 0.001 m/m.........................................................................................................43
Fig. 2.7. Temperature field and free convection patterns when three faults penetrate the basin fill...........................................................................................................45
Fig. 3.1. Regional geological map showing locations of the Athabasca and Thelon basins and major uranium deposits........................................................................56
Fig. 3.2. Simplified paragenetic relationships of the Athabasca, Thelon and Kombolgie basins with major tectonic events........................................................................57
Fig. 3.3. Simplified geological map showing important structures surrounding the Kombolgie and locations of major uranium deposits...........................................60
Fig. 3.4. Idealized hydrogeological model with the most important features of basins that host URU deposits.........................................................................................63
Fig. 3.5. Fluid-flow patterns for different amounts of extensional strain at a strain rate of $10^{-13}$ s$^{-1}$ .........................................................................................................................75
Fig. 3.6. Fluid-flow patterns for different amounts of compressive strain at a strain rate of $10^{-13}$ s$^{-1}$...................................................................................................................77
Fig. 3.7. Fluid-flow patterns and thermal field for thermally-driven convection ........78
Fig. 3.8. Fluid flow patterns and thermal field for different amounts of extensional strain at a strain rate of $10^{-13}$ s$^{-1}$ ........................................................................................................80
Fig. 3.9. Fluid flow patterns and thermal field for different amounts of compressive strain at a strain rate of $10^{-13}$ s$^{-1}$ ............................................................................................... 82
Fig. 3.10. Fluid pore-pressure profiles through the centre of the conceptual model at different strain rates for 1.0 percent strain. ...........................................................84
Fig. 3.11. Schematic diagrams illustrating an idealized genetic model for URU deposits..................................................................................................................91
Fig. 4.1. Regional geological map, and a simplified lithostratigraphic architecture in the Athabasca basin...................................................................................................103
Fig. 4.2. Conceptual model showing the main hydrogeological units.............................108
Fig. 4.3. Initial values of background salinity and its evolution during modeling at 0.1, 1.0 and 5.0 m.y........................................................................................................110
Fig. 4.4. Fluid flow patterns and temperature field of the basic model.........................112
Fig. 4.5. Temperature field overlapped with the Darcy flux "bullets" of two sections.................................................................................................................115
Fig. 4.6. Uranium concentration distribution at different modeling times when the uranium source is located in the center of the basement.................................117
Fig. 4.7. Uranium concentration distribution at different modeling times when the uranium source is located in the basement.........................................................119
Fig. 4.8. Uranium concentration distribution at different modeling times when the uranium source is located in the center of the sandstone unit............................. 121
Fig. 4.9. Uranium concentration distribution at different modeling times when the uranium source is located in the sandstone unit close to the right basin margin..................................................................................................................122
Fig. 4.10. Hydraulic conductivity of the crystalline basement rocks versus depth.................................................................................................................................125
Fig. 4.11. Fluid flow patterns and uranium concentration distribution at different modeling times, with a hydraulic conductivity of $10^{-9}$ m/s assigned to the uppermost 2 km of the basement ................................................................. 126

Fig. 4.12. Uranium concentration at 1 m.y. modeling time .............................................. 128

Fig. 4.13. Fluid flow patterns of the models with relatively low basement hydraulic conductivities of $3 \times 10^{-15}$ m/s and $3 \times 10^{-13}$ m/s ................................................................. 129

Fig. 4.14. Simulated fluid flow, temperature field and uranium transport with a sandstone $K = 1 \times 10^{-8}$ m/s at 1 m.y. modeling time ....................................................... 131

Fig. 4.15. Uranium concentration distribution at different modeling times with the uranium source located in the center of the basement. Uranium concentrations of source area fluids are free to change in this model .............................................. 132
**List of Abbreviations**

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>AP</td>
<td>aluminum phosphate</td>
</tr>
<tr>
<td>APS</td>
<td>aluminum sulfate phosphate</td>
</tr>
<tr>
<td>FLAP</td>
<td>fluorapatite</td>
</tr>
<tr>
<td>H0</td>
<td>primary hematite in the paleoweathered regolith</td>
</tr>
<tr>
<td>H1, H2</td>
<td>early diagenetic hematite in basal red mudstone beds</td>
</tr>
<tr>
<td>H3</td>
<td>pervasive hematite</td>
</tr>
<tr>
<td>H4</td>
<td>dark and intense hematite cement</td>
</tr>
<tr>
<td>M1</td>
<td>the Trans-Hudson Orogeny</td>
</tr>
<tr>
<td>M2</td>
<td>the orogenic events associated with the accretion of Nena</td>
</tr>
<tr>
<td>M3</td>
<td>the Mackenzie Dikes</td>
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<tr>
<td>M4</td>
<td>the Grenville Orogeny</td>
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<tr>
<td>M5</td>
<td>the breakup of Rodinia</td>
</tr>
<tr>
<td>Q1, Q2, Q3</td>
<td>quartz cement</td>
</tr>
<tr>
<td>URU</td>
<td>unconformity-related uranium</td>
</tr>
<tr>
<td>U1</td>
<td>primary uranium mineralization</td>
</tr>
<tr>
<td>XEN</td>
<td>xenotime (a rare earth phosphate mineral)</td>
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**List of Symbols**

\( a_i \) = equilibrium activity of the \( i \)th species [dimensionless]

\( a_{aq} \) = subscript referring to the aqueous phase

\( C_{\rho} \) = solute concentration [kg/m\(^3\)]

\( C \) = cohesive strength of the porous rocks [Pa]

\( C_T \) = effective bulk specific heat [J/(kg·°C)]

\( C_w \) = fluid specific heat [J/(kg·°C)]

\( C_v \) = the mean specific heat capacity of the mineral grains [J/(kg·°C)]

\( C_{v}^{W} \) = groundwater volumetric heat capacity [J/(m\(^3\)·°C)]

\( C_{v}^{S} \) = solid volumetric heat capacity [J/(m\(^3\)·°C)]

\( D \) = hydrodynamic dispersion tensor [m\(^2\)/s]

\( D_m \) = molecular diffusivity [m\(^2\)/s]

\( d_m \) = the mean diameter of granular particles [m]

\( E \) = effective thermal dispersion tensor [W/(m·°C)]

\( G \) = shear modulus of porous media [Pa]

\( g \) = gravitational acceleration [m/s\(^2\)]

\( H \) = thickness [m]

\( h \) = hydraulic head [m]

\( K \) = hydraulic conductivity [m/s]

\( K^T \) = effective thermal conductivity of the saturated porous material [W/(m·°C)]

\( K_e \) = equilibrium constant or solubility product [dimensionless] (Chapter 1)

\( = \) bulk modulus of drained porous media [Pa] (Chapter 3)
\( K_s \) = bulk modulus of rock particles [Pa]
\( K_s^* \) = thermal conductivity of solid matrix [W/(m·K)]
\( K_w \) = bulk modulus of fluid [Pa]
\( K_w^* \) = thermal conductivity of pore fluids [W/(m·K)]
\( K_s \) = horizontal hydraulic conductivity [m/s]
\( K_z \) = vertical hydraulic conductivity [m/s]
\( k \) = permeability [m²]
\( k_0 \) = initial permeability [m²]
\( M \) = mass concentration [mol/m³] (Chapter 1)
\( M_t \) = total mass concentration [mol/m³]
\( n \) = porosity [dimensionless]
\( \vec{n} \) = downward unit vector
\( n_0 \) = initial porosity [dimensionless]
\( p, P \) = fluid pressure [Pa]
\( q \) = Darcy flux of pore fluid flow [m/s]
\( q^T \) = heat-flux vector [W/m²]
\( q^*_v \) = volumetric heat source intensity [W/m³]
\( Ra_T \) = the thermal Rayleigh number [dimensionless]
\( Ra_S \) = the solute Rayleigh number [dimensionless]
\( S \) = degree of saturation [dimensionless]
\( s_0 \) = solid surface exposed to fluids per unit volume of solid material [m⁻¹]
\( sol \) = subscript referring to the solid phase
\( T \) = temperature [°C]
\( T_0 \) = reference temperature [°C]
\( t \) = time [s]
\( \dot{u}_i \) = velocity [m/s]
\( V, V_0 \) = the initial and current volumes of an element [m³]
\( V_i \) = stoichiometric reaction coefficient for the \( i \)th species [dimensionless]
\( v \) = average linear velocity vector [m/s]
\( x_{ij} \) = composition coefficient (moles of component \( i \) per formula weight \( j \)) [dimensionless]
\( x_{ik} \) = composition coefficient (moles of component \( i \) per formula weight \( k \)) [dimensionless]
\( Z \) = elevation [m]
\( z \) = depth [m]
\( \alpha \) = the Biot coefficient [dimensionless]
\( \beta \) = thermal expansion coefficient [K⁻¹]
\( \gamma \) = compressibility [Pa⁻¹]
\( \beta_g \) = thermal expansion coefficient of rock grains [°C⁻¹]
\( \beta_w, \beta_w^T \) = thermal expansion coefficient of fluids [°C⁻¹]
\( \delta_{ij} \) = the Kronecker delta [dimensionless]
\( \varepsilon \) = volumetric strain of porous media [dimensionless]
\( \theta \) = angle of friction [°]
\( \kappa \) = thermal diffusivity [m²/s]
\( \lambda \) = thermal conductivity [W/(m·K)]

\( \mu, \mu_w, \mu_f \) = fluid dynamic viscosity [kg/(m·s)]

\( \mu_0 \) = reference dynamic viscosity [kg/(m·s)]

\( \mu_r \) = dynamic viscosity ratio = \( \frac{\mu_0}{\mu} \) [dimensionless]

\( \sigma_{ij} \) = stress tensor [Pa]

\( \sigma_n \) = normal stress [Pa]

\( \tau \) = maximum shear stress [Pa]

\( \rho \) = groundwater density [kg/m\(^3\)] (Chapter 1)

= bulk density [kg/m\(^3\)] (Chapter 3)

\( \rho_s \) = dry rock density [kg/m\(^3\)]

\( \rho_w, \rho_f \) = fluid density [kg/m\(^3\)]

\( \rho_w^0, \rho_f^0 \) = fluid reference density [kg/m\(^3\)]

\( \Phi \) = porosity [dimensionless]

\( \nabla T \) = temperature gradient [K/m]
Chapter 1

Introduction

1.1 Introduction

Uranium is a commodity that is not only linked to economic development, but also to political issues. Following the initial upsurge in uranium exploration in the 1950s, which was mainly driven by military ventures, studies of uranium deposits declined dramatically in the latter half of the 20th century. Since the beginning of the 21st century, the uranium industry has been experiencing a strong revival due to its potential role in reducing greenhouse gas emissions and slowing global climate change (Vujić et al., 2012). The cycles of exploration and mining of uranium deposits are also reflected by the rise and fall of uranium oxide prices, from US$40 per pound in 1979, to about US$7 per pound in the 1980s, to about US$10 per pound in 2000, to more than US$130 per pound in 2007, and to US$53 per pound at the time of writing this chapter\(^1\). As of August 2011, there are 432 nuclear power plants in operation in the world, which generate about 20% of the world's electricity, and another 65 plants are under construction\(^2\). The current and projected needs of uranium require more accurate genetic models of uranium deposits.

The OECD (the Organization of Economic Co-operation and Development) Nuclear Energy Agency and the International Atomic Energy Agency (IAEA)(2008)


assigned the world uranium resources into thirteen types of deposits based on their geological setting (e.g., host rock type and orebody morphology). Among these deposits, unconformity-related uranium (URU) deposits are the most important type of sediment-hosted uranium deposits on the basis of economic significance. They contribute approximately 33% of the world's uranium reserves, and are mainly located in Paleoproterozoic basins in Canada and Australia\(^3\). They are located within or around basal unconformities between Proterozoic basin fill and underlying Archean granitoid gneiss and Paleoproterozoic metamorphosed sedimentary rocks, where reductants and faults exist (Jefferson et al., 2007). Many models have been proposed for their formation, which include supergene, metamorphic-hydrothermal, magmatic-hydrothermal, and multiphase models (Plant et al., 1999). Currently, the diagenetic model (Hoeve et al., 1980; Hoeve and Quirt, 1984), in which ore formation and host rock alteration are directly linked to stages of basin diagenesis and evolution, is widely accepted and has been subsequently refined (Cuney et al., 2003). Furthermore, the diagenetic model includes two end-member genetic models (e.g., Kyser et al., 2000; Kyser and Cuney, 2009). In one model, the sedimentary fill serves as a source of both the uranium and oxidizing mineralizing fluids (Fig. 1.1A) (e.g., Kyser et al., 2000); uranium minerals precipitate when oxidizing uranium-bearing brines mix with basement-derived reducing fluids or react with reducing minerals in the basement. It has been proposed that thermal convection was responsible for moving fluids around to leach uranium (Raffensperger and Garven, 1995). The other model (Fig. 1.1B) considers the basement as the source of

---

Fig. 1.1. Two end-member genetic models for unconformity-related uranium deposits (Kyser and Cuney, 2009). (A) The basin model involving a sandstone sequence as the source of uranium and oxidized brines; (B) The basement model in which a sandstone sequence provides oxidizing brines, and basement rocks supply uranium.
the uranium that was leached by the oxidizing basinal fluids when they penetrated the basement (Cuney et al., 2003). Despite this general understanding, the fluid flow patterns, the driving forces of fluid flow, the uranium sources, and the relative roles of various factors (e.g., unconformity zones and faults) are still not fully understood.

Fig. 1.2. Regional geological map showing the locations of the Athabasca and Thelon basins (highlighted by two rectangles) and their major uranium deposits (modified from Jefferson et al., 2007). The rectangle in the index map indicates the location of the study area in Canada.

Unconformity-related uranium deposits are currently mainly mined within the Athabasca and Kombolgie basins in Canada and Australia, respectively (Figs. 1.2, 1.3) (Hiatt et al., 2010). According to the data compiled by Gandhi (2007), 587,063 t of
uranium has been identified in the Athabasca basin, and 283,304 t in the Kombolgie basin. The Thelon basin in Canada only hosts two areas with known uranium mineralizations: the Boomerang Lake prospect and the Kiggavik deposit, which lie at the eastern and western margins of the basin, respectively (Fig. 1.2). The basin, however, is believed to hold substantial potential for uranium and other metal deposits due to its similarity to the Athabasca basin in terms of geological setting, paragenesis, and tectonic evolution (Jefferson et al., 2007). Although some URU deposits are present in other basins, such as the Cuddapah basin, India (Banerjee, 2005) and the Pasha-Ladoga basin, Russia (Velichkin et al., 2005), this modeling focused on the three basins with the largest deposits.

Fig. 1.3. Simplified geological map showing important structures surrounding the Kombolgie basin (highlighted by the rectangle) and sites of major uranium deposits (modified form Lindsay, 2001). The rectangle in the index map indicates the location of the study area in Australia.
Numerical modeling can provide considerable insights into what the controlling factors might be in the formation of mineral deposits, and how these factors interact. Because ore-forming processes usually happen slowly in time and widely in space, it is almost impossible to reproduce these processes entirely in the laboratory. Numerical modeling, however, has the ability to test, compare, and contrast various geological models through integrating all types of data. A number of numerical studies have been carried out to investigate the genetic models of ore deposits related to hydrothermal systems (e.g., Garven, 1985, 1994; Raffensperger and Garven, 1995; Yang et al., 2004a; Yang et al., 2004b; Oliver et al., 2006; Feltrin et al., 2009; Yang et al., 2010a, b), and have provided valuable knowledge to mineral exploration and scientific research. In this research, numerical modeling was applied to integrate existing geological, geophysical and geochemical data to study the formation of URU deposits. Four basic questions have been addressed: 1, what are the major driving forces for mineralizing fluid flow in URU deposit-bearing sedimentary basins?; 2, what are the key factors focusing the ore-forming fluids into favorable districts?; 3, what are the essential conditions for the formation of unconformity-related uranium deposits?; and 4, what is the source of uranium?.

1.2 Background and literature review

1.2.1 Uranium transport and deposition in aqueous systems

Uranium transport and deposition in aqueous systems is one of the most important controlling factors in the formation of major uranium deposits. Uranium in minerals exists predominantly in the hexavalent ($\text{U}^{6+}$) state in relatively oxidizing environments, such as near-surface groundwater systems. Tetravalent uranium ($\text{U}^{4+}$), however, is the
dominant uranium ion in reducing conditions that are more typical of magmatic and metamorphic environments (Skirrow et al., 2009). More than 150 uranium-containing minerals have been identified in nature. They have been generally classified into three groups: 1) uranium ore minerals, such as U^{4+}-bearing uraninite and pitchblende, and U^{6+}-bearing oxyhydroxides and phosphates; 2) accessory minerals, such as apatite, zircon and monazite; and 3) complex ore minerals, such as davidite and euxenite in which U^{4+} substitution occurs.

The hexavalent and tetravalent uranium ions are both hard acids, and thus tend to form complexes with hard bases, such as F^-, OH^-, NO_3^-, CO_3^{2-}, HCO_3^-, SO_4^{2-}, HSO_4^-, PO_4^{3-}, H_2PO_4^- and H_2PO_4^-. Chloro complexes with U^{4+} also form in highly saline fluids. Table 1.1 shows some common aqueous complexes of uranium (Skirrow et al., 2009).

Table 1.1. Aqueous complexes of uranium

<table>
<thead>
<tr>
<th>Complex Type</th>
<th>Uranium Species</th>
</tr>
</thead>
<tbody>
<tr>
<td>Simple, oxy, and hydroxy</td>
<td>U^{3+}, U^{4+}, U(OH)^{3+}, U(OH)_2^{2+}, U(OH)_3^{1+}, U(OH)_4,</td>
</tr>
<tr>
<td></td>
<td>U(OH)_5^{1-}, U_2(OH)_3^{3+}, UO_2^{1+}, UO_2^{2+}, UO_2(OH)^{1+},</td>
</tr>
<tr>
<td></td>
<td>UO_2(OH)_2, UO_2(OH)_3, (UO_2)_2(OH)_2^{2+}, (UO_2)_3(OH)_5^{3+}</td>
</tr>
<tr>
<td>Carbonate (uranil)</td>
<td>UO_2CO_3, UO_2(CO_3)_2^{2-}, UO_2(CO_3)_3^{4-}</td>
</tr>
<tr>
<td>Phosphate (uranous and uranyl)</td>
<td>UHPO_4^{2-}, U(HPO_4)_2^{2-}, U(HPO_4)_3^{2-}, U(HPO_4)_4^{4-},</td>
</tr>
<tr>
<td></td>
<td>UO_2(HPO_4), UO_2(HPO_4)_2^{2-}, UO_2(H_2PO_4)^{2-}, UO_2(H_2PO_4)_2,</td>
</tr>
<tr>
<td></td>
<td>UO_2(H_2PO_4)_3;</td>
</tr>
<tr>
<td>Sulfate (uranous and uranyl)</td>
<td>U(SO_4)_2, UO_2(SO_4), UO_2(SO_4)_2^{2+}, USO_4^{2+}</td>
</tr>
<tr>
<td>Fluoride</td>
<td>UF^{3+}, UF_2^{2+}, UF_3^{3+}, UF_4, UF_6^{6-}</td>
</tr>
<tr>
<td>Chloride</td>
<td>UCl^{3+}, UO_2Cl^{+}</td>
</tr>
</tbody>
</table>

The geochemical properties of uranium in aqueous systems at temperatures below 100°C are relatively well-known based on a wide range of experiments (Kyser and Cuney, 2009). Uranium characteristics above 100°C, however, are not well-constrained by experimental data, and are mainly derived from theoretical calculations, for example the data from the revised Helgeson-Kirkham-Flowers equation of state (Shock et al., 1997). The generalized and favourable conditions for uranium transport in low to moderate temperature (< 200 °C) are listed below according to the summary by Skirrow et al. (2009).

1. Only geofluids with an oxygen fugacity (fO₂) value above the hematite-magnetite buffer conditions are able to transport uranium in significant quantities (U concentration > 0.01-1 ppm).

2. Uranium solubility generally increases with increasing fO₂, decreasing pH in the acidic regime, increasing temperature (except at very high fO₂), and increasing concentrations of aqueous phosphate, bicarbonate, sulfate, chloride and fluoride ions.

3. Uranyl phosphate complexes are dominant at intermediate pH conditions (~4 to 8), although uranium transport may be mainly in the form of bicarbonate, oxy-hydroxide, or sulfate complexes where aqueous phosphate concentrations are low relative to other potential ligands.

Uranium precipitation from aqueous fluids is controlled by changes in the redox state, temperature, pH, and ligand composition and concentration (Skirrow et al., 2009). Uranium species are present dominantly as relatively movable hexavalent-uranium
bearing uranyl complexes in oxidizing fluids. Reductants are consequently required to
deposit tetravalent-uranium bearing minerals, such as uraninite, coffinite, and brannerite.
The most common reductants include reducing carbon (e.g., organic matter,
hydrocarbons and graphite), Fe$^{2+}$ ions in minerals or aqueous species, and reducing sulfur
in sulfide minerals (e.g., H$_2$S gas and aqueous H$_2$S). Deposition may occur during fluid-
rock interaction and/or fluid mixing. It is believed that uranium mineral precipitation
from redox reaction was responsible for the formation of Proterozoic URU deposits
(Kyser and Cuney, 2009). If chloride and fluoride complexes of U$^{6+}$ and U$^{4+}$
predominate, a temperature decrease and pH increase may cause uranium deposition from
high temperature solutions regardless of redox conditions. The change in pH and/or
increase in the concentration of other ions, such as Ca, K, and V are able to destabilise
uranyl uranium aqueous complexes; redox reactions are not involved in this process. In
addition, uranium precipitation may occur through adsorption and bacteria-related
activities at low temperatures (Fredrickson et al., 2000; Sherman et al., 2008).

Skirrow et al. (2009) proposed six key parameters that control hydrothermal
uranium mineralization, based on the mineral systems paradigm of source-transport-deposition (Wyborn et al., 1994) and the mineral system approach developed by the
predictive mineral discovery Cooperative Research Centre (pmd*CRC) in Australia
(Walshe et al., 2005). These parameters include fluid sources, uranium sources, fluid
flow drivers, fluid flux over time, fluid pathway and depositional parameters.

1.2.2 Geological setting of URU deposits
Unconformity-related uranium deposits comprise pods, veins, and semi-massive accumulations of uranium-bearing minerals (mainly uraninite) around basal unconformities below thick continental siliciclastic sequences. Most are hosted by Paleoproterozoic sedimentary basins, namely the Athabasca, Thelon and Kombolgie basins, in Canada and Australia. The major geological setting and diagenetic characteristics are summarized below.

The Athabasca basin

The Athabasca sedimentary basin (Fig. 1.2) covers about 100,000 square kilometers in northern Saskatchewan and northeastern Alberta, and has a maximum formation age of ca. 1730 Ma (Alexandre et al., 2009). The basement is composed of Archean to Paleoproterozoic rocks that were metamorphosed during the Trans-Hudson Orogeny (ca. 1800 Ma) (Lewry and Sibbald, 1980). The overlying sedimentary basin fill is represented by the Athabasca Group (1-2 km thick), which mainly comprises quartz-rich sandstone and conglomerate from alluvial, fluvial and upper-shore sedimentary environments. Four major sequences, which are bounded by basin-wide unconformities, have been recognized in this group (Hiatt and Kyser, 2007). The basal sequence of the Athabasca Group (Manitou Falls and Fair Point formations) consists of hematite-rich conglomerate and sandstone. Overlying the basal sequence is an arkosic formation of less-permeable marine sandstone, siltstone, and mudstone. In turn, this formation is capped by shale (Douglas Formation) and stromatolitic dolomites (Carswell Formation) (Jefferson et al., 2007).
The diagenetic history of the Athabasca basin has been investigated using petrography, stable isotope analysis, geochronology, and fluid inclusion analysis of altered and unaltered rocks, which are proximal and distal to the uranium deposits, respectively (Hiatt et al., 2010). Early diagenesis of the Athabasca basin caused quartz overgrowth and hematite precipitation at 1600-1750 Ma. Fluid inclusions from the quartz overgrowths have homogenization temperatures of 150-170°C and a salinity of ~25 wt percent NaCl equiv. Subsequently, the basinal brines underwent an increase in temperature, salinity and $\delta^{18}O$ as a consequence of burial. Finally, U/Pb dates from uraninite and Ar/Ar dates from syn-ore illite indicate that the major uranium mineralizing event began at 1600 Ma. Several later remobilization events have also been identified, which are interpreted to have occurred in response to far-field tectonic events at ca. 1400, 1270, 1100 and 850 Ma (Hiatt et al., 2010). The uranium deposits in the Athabasca basin represent the largest and highest-grade uranium deposits in the world, and comprise the only uranium district in Canada that is actively being mined (Jefferson et al., 2007).

*The Thelon basin*

The Paleoproterozoic Thelon basin straddles the border between Nunavut and the Northwest Territories of Canada (Fig. 1.2). This basin is underlain by Archean to Paleoproterozoic basement rocks that were affected by the Hudsonian orogeny and intruded by post-tectonic Hudsonian granites. The overlying sediments include three sequences, which are, from oldest to youngest, the Thelon Formation, Kuungmi Formation and Lookout Point Formation. A sequence stratigraphic framework for the Thelon Formation, involving three third-order sequences, has been presented by Gall et
al. (1992) and Hiatt et al. (2003). The unimodal westward paleocurrent direction in the Thelon Formation and other surrounding Paleoproterozoic basins indicates that the succession was initially deposited over a large portion of the Churchill Province after the Trans-Hudson orogeny (ca. 1.85 Ga) (Palmer et al., 2004). The present depth of the basement interface under the Thelon Formation has been determined using 1D seismic reflection data from Overton (1979). Palmer et al. (2004) concluded that the source areas for the Thelon Formation evolved from proximal to distal, and back to proximal, based on zircon geochronology and oxygen isotope data from detrital quartz. The detrital material in the lower third-order sequence was derived predominantly from the proximal Neoarchean and Paleoproterozoic metamorphosed sedimentary rocks surrounding the eastern Thelon basin. Zircon in the middle third-order sequence has a range of ages (3.94-1.78 Ga).

Petrographic and fluid inclusion data indicate that the Thelon basin experienced a protracted diagenetic history ( > 600 m.y.) (Renac et al., 2002; Hiatt et al., 2010). The diagenetic evolution has been divided into three stages: early diagenesis (1720-1650 Ma), peak diagenesis (1650-1000 Ma) and late diagenesis (after 1000 Ma), based on the nature of diagenetic mineral assemblages and diagenetic conditions (e.g., temperature). Peak diagenesis and primary mineralization began when fluid temperatures reached approximately 200°C (Renac et al., 2002). The Thelon basin only hosts two known occurrences of uranium mineralizations, the Boomerang Lake prospect, and the Kiggavik deposit, which lie at the eastern and western margins of the basin, respectively. The basin, however, holds substantial potential for uranium and other metal deposits due to its
similarity with the Athabasca basin in terms of geological setting and diagenetic evolution (Jefferson et al., 2007).

**The Kombolgie basin**

Unconformity-related uranium deposits contribute 20 percent of Australia’s total uranium resources. Most of the uranium mined in Australia since 1980 is from the Kombolgie basin, which forms the northern part of the larger McArthur basin (Fig. 1.3) (Hiatt et al., 2007). The Kombolgie basin is located in the Northern Territory of Australia, and formed at about 1793 Ma, after the Barramundi Orogeny (1890-1870 Ma) and the Top End Orogeny (1863-1847 Ma) (Hiatt et al., 2007). The Kombolgie basin is floored by Archean to Paleoproterozoic gneisses and metasedimentary rocks; the latter are domed by pre-orogenic granitic intrusions and the Zamu Dolerite (1884 Ma). Overlying the steeply-dipping basement is the flat-lying Kombolgie Subgroup (1-2 km), which is mainly composed of sandstone and conglomerate with interlayered volcanic units. Hiatt et al. (2007) divided the Kombolgie Subgroup into three sequences. The lowermost sequence evolved from a proximal high-energy braided facies (coarse-grained sandstone and conglomerate) to a distal low-energy braided-stream environment (coarse-to medium-grained quartz arenite). The middle sequence is interpreted as a coarse-grained fluvial facies, overlain by a distal fluvial and interbedded marine and eolian facies. The uppermost sequence is composed of distal fluvial and marine sedimentary rocks that suggest marine transgression.

Several stages of sandstone diagenesis have been identified for the Kombolgie Subgroup (Kyser et al., 2000; Polito et al., 2006), and the following is a summary of the
work of these authors. The early stage is characterized by the formation of quartz overgrowths at 80-130°C from low-salinity (< 10 wt percent) NaCl fluids. The next stage is marked by the precipitation of illite and chlorite at temperatures greater than 200°C at 1650 ± 80 Ma. Quartz vein formation represents the third stage of diagenesis. Fluid inclusions from these veins have homogenization temperatures of 200-400 °C and a salinity of about 22 wt percent NaCl equiv. The final stage of alteration is represented by widespread kaolinite precipitation in the Kombolgie Group that penetrates at several hundred meters depth away from the surface along fractures. Uranium-Pb and $^{207}$Pb/$^{206}$Pb ratios of uraninite in the Jabiluka uranium deposit indicate that primary uraninite precipitated at 1680 Ma, which was followed by remobilization at 1300, 1190 and 800 Ma. Syn-ore illite crystallinity data and chlorite chemistry suggest a mineralization temperature range of between 150 and 250°C. The Nabarlek deposit is associated with a reverse fault or shear zone and primary mineralization formed from 200°C basinal brines at ca. 1640 Ma.

1.3 Methodology and principles

An improved understanding of ore deposit genesis in sedimentary basins has been gained by integrating various data sets into numerical models. A topographically-driven flow model has been accepted as the most reasonable explanation for MVT (Mississippi Valley Type) deposits, in large part as a result of extensive numerical modeling studies (e.g., Deming and Nunn, 1991; Garven et al., 1993; Makhoukhi et al., 2000; Cathles and Adams, 2005). Various numerical experiments also have been designed to evaluate the roles of the major driving forces (compaction, gravity and buoyancy) for the formation of SEDEX deposits (e.g., Druschel et al., 2002; Garven et al., 2003; Yang et al., 2004b;
Yang, 2006; Yang et al., 2006; Feltrin et al., 2009). By comparison, limited numerical studies have been made to simulate hydrothermal fluid flow associated with the formation of URU deposits. Raffensperger and Garven (1995) examined the role of thermally-driven free convection as a mechanism for driving regional groundwater flow systems and URU ore formation in a highly hypothetical 2D hydrostratigraphic model constrained by some common features of the Athabasca (Canada) and McArthur (Australia) basins. De Veslud et al. (2009) investigated the relationship between the development of breccias and mineralization using numerical modeling, and proposed that quartz dissolution created space for fluid mixing. Other numerical modeling studies concerning the URU deposits in the eastern part of the Athabasca basin have been carried out by CSIRO (Commonwealth Scientific Industry Research Organization) and SRC (Saskatchewan Research Council) between May 2000 and March 2003, as a sub-project of EXTECH IV (EXploration science and TECHnology initiative)(Jefferson et al., 2007). No results, however, have been published. Although some improved insights into the formation of URU deposits have been obtained from the published studies mentioned above, some important controlling factors have been ignored: 1) fluid flow patterns in the basement and the relevant determining factors; 2) the role of the basal unconformity as a preferential flow pathway due to the presence of highly permeable weathered regolith; 3) the impact on fluid flow of a tectonic stress field and the resultant deformation; and 4) uranium transport through thermohaline convection.

This study followed the general numerical modeling framework for addressing scientific problems (Fig. 1.4) (Cimbala and Cengel, 2006). The goal is to simulate fluid flow related to URU deposits in the geological framework of three representative
Paleoproterozoic sedimentary basins. A typical unconformity-related uranium system contains five important elements: 1) a basal unconformity where paleo-regolith can be found; 2) a metamorphosed basement containing graphitic rocks, usually beneath the unconformity; 3) a flat-lying fluvial sandstone cover; 4) domes of Archaean granite-gneiss; and 5) first-order basement faults that were reactivated as a result of far-field tectonic events. The hydrological properties associated with these components, such as porosities, permeabilities, and thermal conductivities, are important variables controlling fluid flow. These properties were determined based on data used in similar numerical modeling studies and on published compilations (e.g., Raffensperger and Garven, 1995;
Yang et al., 2004b), and corresponding improvements have been made according to the geological data of these Paleoproterozoic basins. Fluid parameters (salinity, temperature and composition) were derived from fluid inclusion data. The relevant physical laws controlling fluid flow include Darcy's law, Fourier's law, and the laws of conservation of energy and mass. Governing equations derived from these laws for fluid flow, heat and reactive mass transport are listed in Table 1.2 (equations 1.1 to 1.6). These equations were solved with FEFLOW in Chapters 2 and 4. FEFLOW is a software package for simulating groundwater flow, mass transport and heat transfer in porous media, and uses the finite element method to solve groundwater flow and other equations. When tectonic deformation was coupled into the modeling in Chapter 3, the finite-difference code FLAC was employed. The viscosity of fluid phase \( \mu \) is a function of temperature \( T \) and solute concentration \( C_\rho \) as follows,

\[
f_\mu = \frac{\mu}{\mu(C_\rho, T)} = \frac{1 + 1.85\omega(C_\rho = C_{\rho 0}) - 4.1\omega^2(C_\rho = C_{\rho 0}) + 44.5\omega^3(C_\rho = C_{\rho 0})}{1 + 1.85\omega - 4.1\omega^2 + 44.5\omega^3} \times \frac{1 + 0.7063\xi = 0.04832\xi^3}{1 + 0.7063\xi(T = T_0) = 0.04832\xi^3(T = T_0)}, \tag{1.7}
\]

with \( \omega = \frac{C_\rho}{\rho_f} \), \( \xi = \frac{(T - 150)}{100} \).

This function was used in Chapters 2 and 4. A constant fluid viscosity of \( 1.0 \times 10^{-3} \) kg/(m·s) was assumed in Chapter 3 when tectonic deformation was coupled into the modeling.

1.4 Organization of thesis

This thesis is structured as follows.

Chapter 2 mainly deals with numerical experiments related to thermally-driven free convection. A paleo-hydrostratigraphic model was developed for the time of uranium
Table 1.2. Summary of equations governing groundwater flow, heat transport and reactive mass transport (Garven and Raffensperger, 1997)

Conservation of groundwater mass:
\[ \nabla \cdot (\rho q) = -\frac{\partial \rho \phi}{\partial t} \quad (1.1) \]

Conservation of momentum (Darcy’s law):
\[ q = -K \mu_r (\nabla h + \rho, \nabla Z) \quad (1.2) \]

Conservation of thermal energy:
\[ \nabla \cdot (E \nabla T) - C_v^w q \cdot \nabla T = \left[ \phi C_v^w + (1-\phi) C_v^s \right] \frac{\partial T}{\partial t} \quad (1.3) \]

Conservation of solute mass in flow system:
\[ \nabla \cdot (D \nabla \phi M_{t,i,(aq)}) - v \cdot \nabla \phi M_{t,i,(aq)} = \frac{\partial \phi M_t}{\partial t} \quad (1.4) \]

| a_i | Equilibrium activity of the i\textsuperscript{th} species |
| aq | Subscript referring to the aqueous phase |
| C_v^w, C_v^s | Groundwater and solid volumetric heat capacity |
| D | Hydrodynamic dispersion tensor |
| E | Effective thermal dispersion tensor |
| h | Equivalent freshwater hydraulic head |
| K_e | Equilibrium constant or solubility product |
| K | Hydraulic conductivity tensor |
| M | Mass concentration (molarity) |
| M_t | Total mass concentration (molarity) |
| q | Darcy velocity or specific discharge vector |
| sol | Subscript referring to the solid phase |
| T | Temperature |
| v | Average linear velocity vector |
| Z | Elevation |
| M_{t,i,(aq)} | Mass balance in chemical subsystem: |
| \[ M_{t,i,(aq)} = \sum_j x_{ij} M_{j,(aq)} \] (1.5a) |
| M_{t,i,(sol)} | Mass balance in chemical subsystem: |
| \[ M_{t,i,(sol)} = \sum_k x_{ik} M_{k,(sol)} \] (1.5b) |
| M_{t,i} | Mass action in chemical subsystem: |
| \[ M_{t,i} = \sum_j x_{ij} M_{j,(aq)} + \sum_k x_{ik} M_{k,(sol)} \] (1.5c) |
| K_e | \[ K_e = \prod_i a_i^{r_i} \] (1.6) |

| \( \mu \) | Dynamic viscosity of groundwater |
| \( \mu_0 \) | Reference dynamic viscosity |
| \( \mu_r \) | Dynamic viscosity ratio = \( \mu_0 / \mu \) |
| \( V_i \) | Stoichiometric reaction coefficient for the i\textsuperscript{th} species |
| \( \rho \) | Groundwater density |
| \( \rho_0 \) | Reference groundwater density |
| \( \rho_r \) | Relative groundwater density = \((\rho - \rho_0) / \rho_0 \) |
| \( \phi \) | Porosity |
| x_{ij} | Composition coefficient (moles of component i per formula weight j) |
| x_{ik} | Composition coefficient (moles of component i per formula weight k) |
mineralization through integrating existing geological, geophysical and geochemical data for the Thelon basin, especially the thirty-five available 1D seismic profiles. Based on this model, a number of scenarios were designed to examine the roles of various factors in controlling ore-forming fluid flow. The modeling suggests that free convection develops throughout the Thelon Formation given typical hydrological parameters (Fig. 2.3). Changes of the assumed geothermal gradients in the range of 25-35 °C/km, and the presence of high-permeability basal unconformities and/or diagenetic aquitards do not modify the fluid flow patterns significantly at the basin scale. The highest fluid fluxes were found at the basal unconformity, which supports the fact that uranium is concentrated within or around basal unconformities. Three relatively low-permeability units with a thickness of 200 m were added into the model domain to reflect its heterogeneity; consequently, convection cells are partially separated and fluid flow velocities are slightly reduced. Variation of water table slopes leads to complicated streamline configurations; however, it is not the dominant driving force for fluid flow due to the random distribution of URU deposits along basin margins. The importance of faults depends on their scales, spatial distribution densities, relationship to the aquifer and their connectivity with the water table. More details about the role of faults in the formation of uranium deposits will be provided in the following section. Fluids in the basement are not observed to flow upward through the basement faults as was expected, probably because active faulting processes were not taken into account in these models.

The influence of tectonic deformation was coupled into the modeling in Chapter 3. In this chapter, a relatively simple conceptual model at local scale was designed in order to accommodate the long computing time required by the thermo-hydro-mechanical
modeling. Although a simplified model, it includes the most important elements for a typical URU deposit in the Athabasca, Thelon and Kombolgie basins. The interaction among fluid flow, heat transport, topographic relief and tectonic deformation was analyzed on the basis of this local model. The FLAC2D code was selected to solve all the equations related to these coupled processes. The modeling suggests that thermally-driven free convection develops in the sandstone sequence during periods of tectonic quiescence, given a typical geothermal gradient of 30 °C/km. Reactivation of preexisting basement structures and the generation of new faults, resulting from tectonic deformation, suppress free convection and lead to deformation-dominated fluid flow or mixed convection, depending on strain rates. During compressive deformation, reduced brines in the basement may be squeezed out along fracture zones and encounter uranium-bearing fluids in the clastic sequence to form sandstone-hosted URU deposits. By contrast, basement-hosted deposits are likely to form during extension, when oxidized basinal brines flow into faulted structures to interact with reduced minerals or fluids in the basement. Maximum flow rates are estimated to be up to 7 m/yr and 2.5 m/yr in the relatively high-permeability fracture zone for compressive and extensional deformation, respectively, given a strain rate of $10^{-13}$ s$^{-1}$ and a geothermal gradient of 30 °C/km. In addition, the rate of pressure accumulation and dissipation is different in various geological units depending on their hydrological and mechanical properties even during the same tectonic event; this difference is the main factor determining fluid interactions along faults across adjoining geological units.

Although fluid migration along faults during tectonic deformation explains, to some degree, the fluid-fluid and fluid-rock interactions in and around the intersections of
faults and basal unconformities, an understanding of the pervasive interaction straddling the basal unconformity remains unresolved. In Chapter 4, in order to address this question, FEFLOW 6.0 was used to investigate thermohaline convection, and the new data visualization technique of FEFLOW 6.0 was used to analyze the modeling results. The built-in equation of state (EOS) for fluids in FEFLOW is only applicable to fluids ranging between 0 and 100 °C (Diersch, 2009). To satisfy the wide range of temperature (20-320°C), pressure and salinity (0-340,000 mg/l) relevant to URU deposits, the original EOS was replaced according to the method of Magri (2010). The results show that buoyancy-driven convection may penetrate into the basement for up to 1-2 km below the unconformity, where typical hydrological parameters for these Proterozoic hydrogeological units are used. Fluid flow velocities in the sandstone sequence are several orders of magnitude larger than those in the basement. If a uranium source (a pore fluid with 500 mg/l U) is assumed to be located in the center of the basin below the unconformity, uranium would gradually spread into the sandstone aquifer. The uranium concentration of basinal fluids above the uranium source approaches 15 and 24 mg/l after 1 and 5 m.y. of modeling time, respectively. If the uranium source is initially located at the centre of the aquifer, a uranium plume develops and percolates down to up to 2 km below the unconformity at 5 m.y. The location of the uranium source also affects the solute transport efficiency. A uranium source located around the sloping basal unconformity, either in the basin fill or basement, close to the basin margin, leads to a wider uranium plume than if it is located near the center of the basin. Given appropriate hydrological conditions, buoyancy-driven convection could have caused widespread interactions of basinal brines with basement rocks or basement-derived fluids in
Proterozoic basins where RUR deposits have developed; enough uranium could have been leached from the uranium-rich basement and transported to favourable depositional sites to form large, high-grade RRU deposits.
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Chapter 2

Numerical modeling of hydrothermal fluid flow in the Paleoproterozoic Thelon basin, Nunavut, Canada (thermally driven free convection model)

2.1 Introduction

The Paleoproterozoic Thelon basin straddles the border between Nunavut and the Northwest Territories of Canada. It hosts the Kiggavik uranium deposit on the eastern periphery of the basin and the Boomerang Lake prospect on the western periphery, and represents the area with the most potential for new discoveries of unconformity-type uranium deposits in Canada (Jefferson et al., 2007). It is generally believed that unconformity-type uranium deposits may form by the mixing of an oxidizing fluid circulating in basinal sandstones with a reducing fluid emanating from basement rocks, at or near the unconformity (Kyser and Cuney, 2009). Despite this general understanding, the mechanisms of fluid flow, and the heat and mass transport required for ore formation are still not fully understood.

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A variety of numerical models have been developed to explain the connection between fluid flow and ore deposit genesis (e.g., Deming and Nunn, 1991; Raffensperger and Garven, 1995; Zhao et al., 2001; Yang et al., 2004), especially for MVT and SEDEX deposits. By comparison, little computational effort has been directed to simulate hydrothermal fluid flow associated with unconformity-type uranium deposits (Raffensperger and Garven, 1995). To date, no computational studies are available for the subsurface hydrothermal flow systems that are likely to control ore-forming processes of unconformity-type deposits in the Thelon basin. This paper helps to fill this gap.

In this study, the hydrostratigraphic framework of the Thelon basin is first constructed at the time of uranium mineralization (ca.1650 Ma). In the context of this reconstruction, the governing equations of fluid flow and heat transport are then solved numerically using the finite-element software package FEFLOW (Diersch and Kolditz, 1998). This paper is concerned in particular with thermally- and topography-driven fluid flow and related controlling factors without taking into account possible salinity variations. Tectonic deformation is not considered due to the limits of FEFLOW, although it has been proposed to be responsible for driving ore-forming fluid flow in some environments (e.g., Oliver et al., 2006).

2.2 Development of conceptualized model

2.2.1 Integration and analysis of existing data

The Thelon basin is underlain by Archean to Paleoproterozoic basement rocks that were affected by the Hudsonian orogeny and intruded by post-tectonic Hudsonian
granites. The overlying sedimentary package includes three sequences, which are, from oldest to youngest, the Thelon Formation, Kuungmi Formation and Lookout Point Formation. As shown in Fig. 2.1, the unimodal westward paleocurrent directions in the Thelon Formation (e.g., the flow direction of braided streams when the Thelon Formation was accumulating) and in other surrounding Paleoproterozoic basins indicate that the succession was initially deposited over a large portion of the Churchill Province after the Trans-Hudson orogeny (ca. 1850 Ma). Detrital zircon geochronology and detrital quartz oxygen isotope data indicate that the sediments in the Thelon Formation were derived predominantly from the east (Palmer et al., 2004).

Fig. 2.1. Map of the Thelon basin with locations of seismic exploration sites and generalized paleocurrent directions (modified after Overton, 1979; Hiatt et al., 2003). Sequence distribution and drill core locations are also shown. The black square in the index map indicates the location of the study area.
Thirty-five 1D seismic profiles indicating the general distribution and thickness of the Thelon Formation have been published and interpreted by Overton (1979). From these interpreted seismic profiles, three subsidence centers are identified for the preserved Thelon Formation (Fig. 2.1). A cross-section through the whole basin from NE to SW was constructed by connecting seismic sites (Fig. 2.2). The basement depth between sites was determined by interpolation. In this section, the basin is asymmetrical, with a basal unconformity that is shallower-dipping on the eastern side, compared to the western side (Fig. 2.2).

Fig. 2.2. Paleo-hydrostratigraphic model for the Thelon basin. Various hydraulic units are indicated by rock properties. Three faults are named Fault 1, Fault 2 and Fault 3 from left to right. The cross-section location is shown in Fig. 2.1.

Uranium-lead dating of authigenic fluorapatite indicates that the Thelon Formation formed prior to 1720 Ma. The youngest age for the last magmatic event in the basement is 1753 Ma, which limits the beginning of Thelon Formation sedimentation (Miller et al., 1989). Argon-argon ages of peak-diagenetic (highest temperature) illite reveal a peak diagenesis period of ca. between 1690 and 1000 Ma (Renac et al., 2002). Kyser (2008, personal communication) considers the primary mineralization age in the
Thelon basin to be between 1670 and 1400 Ma, and most likely near 1600 Ma, as is the case in the Athabasca basin.

The Thelon Formation comprises flat-lying conglomerate and sandstone with a thickness of 1-2 km. The overlying Kuungmi Formation is represented by an aphanitic basalt flow with a thickness of less than 10 m and a limited spatial distribution (Fig. 2.1). This volcanic unit is overlain by the stromatolitic and dolomitic Lookout Point Formation with a thickness of less than 40 m. The Thelon basin experienced a protracted diagenetic history (>600 Ma), as evidenced by petrographic and fluid inclusion data (Renac et al., 2002). The Thelon Formation has been divided into three third-order sequences (Hiatt et al., 2003). The uppermost part of every third-order sequence in the Thelon Formation consists of more compositionally and texturally mature sediments than the underlying units. The absence of a clay matrix in these uppermost cycles resulted in a high initial porosity, allowing quartz cementation to develop in these units, converting them into aquitards (i.e., diagenetic aquitards) prior to uranium mineralization. In contrast, a higher clay content in the lower parts of these third-order sequences inhibited the development of quartz cementation. Consequently, the units with initially low porosity ultimately acted as conduits for mineralizing fluids during the peak diagenesis phase (Hiatt et al., 2003). Peak diagenesis and primary mineralization began when fluid temperature reached ~200 °C (Renac et al., 2002).

Reactivated faults, intersecting both basement and the cover clastic sequence, which acted as the conduits for reduced fluids rising from the basement and/or for the descent of uranium-bearing fluids, are an essential structural component in the
development of unconformity-related uranium deposits (Kyser and Cuney, 2009). The Kiggavik deposit (Fig. 2.1) is hosted within the crystalline basement, close to the faulted margin of the north-eastern Thelon basin (Hiatt et al., 2003). The existence of growth faults has been confirmed by high-resolution seismic surveys as part of the EXTECH IV project in the Athabasca basin (Jefferson et al., 2007).

2.2.2 Paleo-hydrostratigraphic model

The previous work summarized above provides constraints for the development of the hydrostratigraphic model. Some conjecture, however, is required as the upper part of the basin fill has been eroded. Assuming that the geothermal gradient was 30 °C/km for the Thelon basin, the sedimentary fill should have had a thickness of 6-7 km in order to attain the peak diagenetic temperatures. The existence of stromatolite-bearing dolomites and evaporites overlying the clastic sequence in the western portion of the basin was interpreted to represent the evolution of the basin from an alluvial-fluvial environment to a stable continental platform or peritidal environment (Hiatt et al., 2003). Thus, it is assumed, for modeling purposes, that there was a shale sequence (0.5-1 km) and a thicker carbonate sequence (1-1.5 km) above the conglomerate-sandstone sequence of the Thelon Formation (Fig. 2.2), and which have subsequently been eroded. Additionally, the basin was assumed to be covered by seawater when mineralization took place (cf. Raffensperger and Garven, 1995). The exact distribution and size of basement faults are uncertain, but it is presumed that they are located where basement relief changes considerably. The hydrostratigraphic model constructed on the basis of published data
and these assumptions (Fig. 2.2) served as the basis for the numerical experiments described below.

In order to describe the geological model exactly, triangular meshes were employed with a maximum resolution of 20 m in the vertical direction. Initially, the three third-order sequences of the Thelon Formation were considered as a single sandstone layer with no faults. For this basic case, the whole section is represented by 39,573 mesh elements and 20,218 mesh nodes.

Table 2.1. Major physical parameters of various hydrogeological units for numerical models

<table>
<thead>
<tr>
<th>Hydrogeological unit</th>
<th>$K_x$ (m/yr)</th>
<th>$K_z$ (m/yr)</th>
<th>$\lambda$ (J/m/s/°C)</th>
<th>$n$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Carbonate</td>
<td>1</td>
<td>0.01</td>
<td>3.0</td>
<td>0.15</td>
</tr>
<tr>
<td>Shale</td>
<td>1</td>
<td>0.01</td>
<td>2.5</td>
<td>0.15</td>
</tr>
<tr>
<td>Sandstone</td>
<td>100</td>
<td>1</td>
<td>3.5</td>
<td>0.20</td>
</tr>
<tr>
<td>Basement</td>
<td>0.001</td>
<td>0.001</td>
<td>2.5</td>
<td>0.10</td>
</tr>
<tr>
<td>Faults</td>
<td>100</td>
<td>100</td>
<td>3.5</td>
<td>0.30</td>
</tr>
</tbody>
</table>

Notes: some of these parameters are changed in this study, see text for details, $K_x$ is horizontal hydraulic conductivity, $K_z$ is vertical hydraulic conductivity, $\lambda$ is thermal conductivity, $n$ is effective porosity.

Various rock properties (Table 2.1) were determined based on data used in similar numerical modeling studies and on published compilations (e.g., Raffensperger and Garven, 1995; Yang et al., 2004). The upper boundary is treated as a water table with a fixed hydraulic head of 0 m. In some models, this condition is adjusted in order to investigate the role of topography. The side and bottom boundaries are assumed to be impermeable to fluid flow. For heat transport, the top boundary is maintained at 20 °C.
The bottom boundary is assigned a temperature value based on, for most cases, a geothermal gradient of 30 °C/km. The side boundaries are insulated to heat.

2.3 Numerical experiments and results

2.3.1 Basic model

The basic case corresponds to the conditions presented in Table 2.1. The temperature field is modified by the moving fluids and convection cells develop in the permeable sandstone layer (Fig. 2.3). The temperature is elevated significantly by advective heat transport around the upwelling centers with the highest values (200 °C) attained at the base of the basin. The magnitude of the temperature perturbations is proportional to the velocity of fluid flow.

Streamlines describe the paths of variable-density fluid flow and the density of the streamlines is proportional to the fluid flux. Convection cells develop throughout most of the sandstone sequence. At the peripheries of the sequence, where the aquifer thickness decreases to ca. 1 km, convection is restricted. Fluid velocities are the greatest at the boundaries between the aquifer and the confining layers, with a maximum value of $2.42 \times 10^{-3}$ m/d (Fig. 2.3B, Table 2.2). In a given convection cell, the average linear velocity decreases gradually towards the center of the cell to ca. $3.03 \times 10^{-4}$ m/d. Fluid velocities in the basement and in the overlying confining layers are $1.96 \times 10^{-9}$ m/d or less.
Fig. 2.3. Modeling results for the basic case. (A) Temperatures are indicated by isotherms and colors. (B) Fluid-flow patterns are shown by streamlines and fluid-flow rates are represented by colors. Arrows indicate fluid-flow directions.

2.3.2 Effect of unconformity

A basal unconformity is the most favorable exploration target for unconformity-related uranium deposits, although the role of the unconformity in the formation of these deposits remains uncertain. The thickness of the unconformity zone can range from a few

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2 The density of streamlines between two neighboured convection cells is increased due to the compression to the figure along the horizontal axis. Without the influence of the compression, streamlines at the boundary between the sandstone and confining units have the highest density.
Table 2.2. Maximum average linear velocity for various numerical cases

<table>
<thead>
<tr>
<th>Scenarios</th>
<th>Maximum average linear velocity (m/d)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.3.1 Basic model</td>
<td>2.42×10⁻³</td>
</tr>
<tr>
<td>2.3.2 Effect of unconformity</td>
<td></td>
</tr>
<tr>
<td>$K_x = 200$ m/yr</td>
<td>4.76×10⁻³</td>
</tr>
<tr>
<td>$K_x = 400$ m/yr</td>
<td>7.72×10⁻³</td>
</tr>
<tr>
<td>$K_x = 1000$ m/yr</td>
<td>1.34×10⁻²</td>
</tr>
<tr>
<td>2.3.3 Effect of stratigraphic</td>
<td>2.21×10⁻³</td>
</tr>
<tr>
<td>heterogeneity</td>
<td></td>
</tr>
<tr>
<td>2.3.4 Effect of topography</td>
<td></td>
</tr>
<tr>
<td>Slope=0.0001 m/m</td>
<td>2.95×10⁻³</td>
</tr>
<tr>
<td>Slope=0.0005 m/m</td>
<td>4.69×10⁻³</td>
</tr>
<tr>
<td>Slope=0.001 m/m</td>
<td>2.49×10⁻³</td>
</tr>
<tr>
<td>2.3.5 Effect of faults</td>
<td></td>
</tr>
<tr>
<td>Size=400×1500m</td>
<td></td>
</tr>
<tr>
<td>10,000 yrs</td>
<td>2.96×10⁻³</td>
</tr>
<tr>
<td>100,000 yrs</td>
<td>2.89×10⁻³</td>
</tr>
<tr>
<td>Size=500×4000m</td>
<td></td>
</tr>
<tr>
<td>10,000 yrs</td>
<td>1.06×10⁻²</td>
</tr>
<tr>
<td>100,000 yrs</td>
<td>5.25×10⁻³</td>
</tr>
</tbody>
</table>

centimeters up to 220 m (Jefferson et al., 2007). In this preliminary model, in order to assess the effect of the unconformity, it is assumed there is an unconformity zone with a thickness of 250 m at the base of the Thelon Formation. Three horizontal hydraulic conductivities (200 m/yr, 400 m/yr and 1000 m/yr) have been used to examine the effect of different permeabilities in this unconformity zone. Varying this parameter has almost no effect on the temperature field in the vertical direction, and the maximum temperature remains at about 200 °C at the base of the sequence. In the horizontal direction, the
Isotherms have a tendency to become flatter with increasing hydraulic conductivity in the unconformity zone. The convection cells still develop in the sandstone sequence, in a similar manner to the previous case (see Fig. 2.3). The horizontal dimensions of the cells, however, grow when the horizontal hydraulic conductivity of the unconformity is increased (Fig. 2.4). For the models where $K_x = 400$ m/yr and $K_x = 1000$ m/yr, some convection cells cross the boundary between the sandstone sequence and overlying layers. Additionally, a few of the convection cells penetrate into the basement. The maximum velocities for the three hydraulic conductivities are, in increasing conductivity magnitude, $4.76 \times 10^{-3}$ m/d, $7.72 \times 10^{-3}$ m/d, and $1.34 \times 10^{-2}$ m/d (Table 2.2).

![Fig. 2.4. Fluid-flow patterns when the unconformity zone is considered. Different hydraulic conductivities are employed for the unconformity, $K_x = 200$ m/yr (A) and $K_x = 400$ m/yr (B).](image)
2.3.3 Effect of stratigraphic heterogeneity

In the models described above, it was assumed that the entire sandstone sequence was homogeneous. In order to examine the effect of heterogeneity in the sandstone sequence, it was subdivided into three layers. In each of the three units, the topmost 200 m (the diagenetic aquitard) was assigned a hydraulic conductivity of $K_x = 1 \text{ m/yr}$, compared to $K_x = 100 \text{ m/yr}$ for the lower part of each sandstone unit. Consequently, the resulting free convection is partially confined by the introduction of these heterogeneities (Fig. 2.5). Convection cells are still observed, and smaller convection cells develop within the third-order sequences. Maximum fluid flow rates are slightly reduced to $2.21 \times 10^{-3} \text{ m/d}$.

2.3.4 Effect of topography

Although carbonates overlie the sandstone sequence in the Thelon basin, there is no direct evidence that suggests the basin was covered by seawater when mineralization occurred. Also, the close spatial connection between unconformity uranium deposits and faults indicates that tectonic events accompanied the mineralizing process. Thus, it is possible that gravity-driven flow developed due to tectonic uplift, and influenced mineralization.

The various water table slopes were implemented by modifying the hydraulic head on the top boundary. The isotherms become smoother when the water table slope is increased from 0.0001 m/m to 0.001 m/m. The temperature profile is not modified
Fig. 2.5. Modeling results when diagenetic aquitards are involved. (A) Temperatures are indicated by isotherms and colors. (B) Fluid patterns are shown by streamlines and fluid-flow rates are represented by colors.

Significantly when the slope is increased from 0.0001 m/m to 0.0005 m/m. The maximum temperature, however, is reduced to ca. 180 °C, given a slope of 0.001 m/m. At low water table slopes (e.g., 0.0001 m/m), free convection dominates the flow system. Here, some convection cells disappear at the margins of the aquifer. For a slope of 0.0005 m/m, mixed convection develops (Fig. 2.6A). When the slope reaches 0.001 m/m, forced convection (topography-driven flow) dominates the whole basin (Fig. 2.6B). The impact on the velocity field is somewhat complicated. For slopes of between 0.0001 m/m and 0.0005 m/m, the maximum fluid rate increases from $2.95 \times 10^{-3}$ m/d to $4.69 \times 10^{-3}$ m/d.
(Table 2.2). The maximum average linear velocity drops to $2.49 \times 10^{-3}$ m/d when forced convection dominates fluid flow at a slope of 0.001 m/m, although this is still higher than the $2.42 \times 10^{-3}$ m/d observed in the basic model.

2.3.5 Effect of faults

First, three faults with a width of 400 m and depth of 1500 m were added. No significant changes are observed in the temperature field between the steady state and 10,000 yrs and 100,000 yrs after faulting. Free convection retains the same configuration at the basin scale. The only difference is that the maximum flow rate becomes slightly

Fig. 2.6. Streamline configuration and Darcy flux field for water table slopes of 0.0005 m/m (A) and 0.001 m/m (B).

2.3.5 Effect of faults

First, three faults with a width of 400 m and depth of 1500 m were added. No significant changes are observed in the temperature field between the steady state and 10,000 yrs and 100,000 yrs after faulting. Free convection retains the same configuration at the basin scale. The only difference is that the maximum flow rate becomes slightly
elevated to $2.96 \times 10^{-3}$ m/d (10,000 yrs) and $2.89 \times 10^{-3}$ m/d (100,000 yrs) (Table 2.2). The size of the faults was then modified to 500×4000 m. In this case, the temperature in and around the faults is reduced considerably, with a decrease of 15 °C at 100,000 yrs after the appearance of faults. The introduction of faults causes the development of a few convection cells in the overlying shale and carbonate layers. The influence, however, gradually becomes weaker over time and almost disappears at 100,000 yrs. The maximum fluid flow rate reaches $1.06 \times 10^{-2}$ m/d at 10,000 yrs, dropping to $5.25 \times 10^{-3}$ m/d at 100,000 yrs. When these faults are allowed to penetrate the carbonate unit, Fault 2 and Fault 3 (Fig. 2.2) are connected by the aquifer to form convection cells. The temperature is reduced in and around Faults 1 and 2, whereas it is elevated around Fault 3 (Fig. 2.7). A few convection cells involving both the aquifer and the overlying confining layers develop.

A series of models were also designed to examine the coupled effect of topographic relief and faults. The modeling results for a water table slope 0.0005 m/m were selected to serve as the initial conditions. Based on this, three faults were added into the hydrostratigraphic model. In this regard, Faults 2 and 3 serve as the major recharge channels of fluid flow, whereas Fault 1 is the discharge locus for the fluid system.

2.4 Discussion and conclusions

Although changes to the unconformity and faults result in fluctuations in the temperature and velocity fields, the free convection patterns remain similar at the basin scale. The introduction of a basal unconformity into the model modifies the streamline
configuration and fluid-flow rate field. Significant changes to the fluid-flow patterns, however, require a horizontal hydraulic conductivity of 400 m/yr or greater. Previous studies have suggested that it is unreasonable for a basal sandstone unit to have horizontal hydraulic conductivities of 300 m/yr or higher (Raffensperger and Garven, 1995).

Paleoregoliths may have a horizontal hydraulic conductivity of 182-1825 m/yr (Tindimugaya, 1995) at a depth of about 30 m, but these values will be lowered as a result of burial and diagenesis. Thus, it is reasonable to assume that the unconformity had a horizontal hydraulic conductivity of less than 300 m/yr. Given such values, the
existence of a basal unconformity does not change the free convection patterns. Modeling shows that fluids within the unconformity zone move at the greatest velocity, which presumably relates to the association of uranium deposits with the unconformity. The results also show that free convection still develops in the Thelon basin even if the three diagenetic aquitards proposed by Hiatt et al. (2003) are added to the model.

Free convection diminishes with an increase in the water table slope, which is consistent with the study of Raffensperger and Garven (1995). In this study, a slope of 0.001 m/m almost shuts-off free convection (Fig. 2.6B). Although topographically driven flow usually produces higher fluid-flow velocities than free convection, due to the existence of confining cap rocks, the maximum average linear velocity obtained (2.486×10^{-3} m/d) at a slope of 0.001 m/m is only slightly higher than that in the basic model. Topographically driven fluid flow will lead to the development of an asymmetrically-distributed temperature field in the basin. Temperatures are reduced on the recharge side, and elevated in the discharge area. Because the thermal regime is an important factor for the formation of deposits, if topography driven flow dominates in the basin, deposits should be asymmetrically distributed, or at least have significantly different characteristics along the fluid flow direction. The two known uranium occurrences are located at the two ends of the Thelon basin. In the Athabasca basin, the deposits are located in the vicinity of the basin edge where faults occur. Thus, unlike MVT deposits, fluid properties do not appear to have changed gradually along a potential topographically driven fluid-flow path, suggesting that topographic relief is not the dominant driving force for fluid flow in the formation of unconformity-related uranium deposits.
deposits. Thus, free convection remains an attractive mechanism for the leaching and redistribution of uranium, and would involve the entire sandstone sequence.

Faults are expected to play an important role in focusing fluids into mineralized zones due to their high hydraulic conductivities. When three faults with dimensions of 500×4000 m were added to the model, the temperatures within and around the faults are reduced over time. When these faults penetrate the basin fill, temperature deviations from previous models are enhanced, and Faults 2 and 3 are connected through the aquifer to form convection cells, as recharge and discharge channels, respectively (Fig. 2.7). An interesting observation is that, although the temperature of the lower part of Fault 3, below the aquifer, is reduced due to the downwelling of water, it is elevated for the upper part, as this is the discharge conduit (Fig. 2.7), unlike in Faults 1 and 2. A similar phenomenon for Fault 1 is observed when faults and topographic relief are coupled. In this case, fluids flow out through the upper part of Fault 1 (above the basement), while no significant fluids are derived from the basement through these basement faults. In relation to this, fluid inclusion studies at the Rabbit Lake deposit in the Athabasca basin (Derome et al., 2003) indicate that the temperature and pressure adjacent to the unconformity decreased from 220 °C and 1.25 kbar (peak diagenesis), respectively, to 140-160 °C and 0.6 kbar (mineralization stage). Thus, observations are consistent with the presented results. The occurrence of faults that penetrate out of the basin fill will result in a decrease in pressure and the introduction of cold water from surface.

The results are consistent with the model of Kyser et al. (2000) who proposed that the sedimentary basin is the source of both uranium and fluids. When these oxidized
uranium-bearing brines flow down into the basement along faults or shear zones and encounter organic matter and/or reduced minerals, uraninite or other uranium minerals will precipitate (basement-hosted unconformity-related uranium deposits, such as the Kiggavik deposit). However, given the scenarios modeled in this study, it is more difficult to explain the formation of deposits in the sandstone sequence, such as the Boomerang Lake prospect, because no reduced fluids were observed to flow out from the basement. In addition, mechanical deformation accompanying tectonic events and solute transport have been ignored in this modeling, although it will be considered in the following research.

In summary, the numerical experiments support the following conclusions:

1. Free convection may have developed in the Paleoproterozoic Thelon basin across most parts of the Thelon Formation sandstone sequence. In the presented model, zones adjacent to the boundary between the aquifer and the confining layers are the loci of relatively high fluid-flow rates, with a maximum value $2.42 \times 10^{-3}$ m/d for the basic case at steady state. Through free convection, uranium could have been leached and transported extensively throughout the basin.

2. The introduction of a basal unconformity zone with a thickness of 250 m has little influence on the basin-scale pattern of free convection. Fluid-flow velocity and fluid flux, however, are enhanced in the unconformity zone. This characteristic makes the unconformity zone favorable for deposit formation.
3. The occurrence of diagenetic aquitards in the Thelon Formation reduces the maximum fluid flow rate to $2.21 \times 10^{-3}$ m/d. Nevertheless, free convection still develops.

4. Complicated fluid-flow patterns occur depending on the scale of topographic relief. A water table slope of 0.001 m/m almost eliminates free convection, with a maximum average linear velocity of $2.49 \times 10^{-3}$ m/d. The lack of an asymmetrical distribution of unconformity-associated uranium deposits in the Thelon basin and Athabasca basin suggests that topography is not the dominant driving force in these basins.

5. The ability of faults to localize fluids depends on their size, spatial distribution frequency and their relationship to the aquifer. Isolated and small-scale faults do not have an influence on basin-scale free convection. These numerical experiments can explain the formation of basement-hosted uranium deposits. However, without considering deformation-driven flow, there are insufficient volumes of reduced fluids derived from the basement to account for the basin-hosted deposits.
References


Chapter 3

Tectonic deformation and fluid flow: implications for the formation of unconformity-related uranium deposits (thermo-hydro-mechanical model)

3.1 Introduction

Unconformity-related uranium (URU) deposits, which are mainly hosted by Paleoproterozoic basins in Canada and Australia, currently supply 30 percent of global uranium (Hiatt et al., 2009). They are located within or around basal unconformities between the Proterozoic basin fill and the underlying Archean granitoid gneisses and Paleoproterozoic metamorphosed sedimentary rocks, where reductants and faults exist (Jefferson et al., 2007). Although significant research has been carried out on these deposits, certain aspects of their formation are still not clearly understood. Many genetic models have been proposed that have included a range of processes, from supergene, to metamorphic-hydrothermal, to magmatic-hydrothermal (Plant et al., 1999). The diagenetic model, which was originally suggested by Hoeve et al. (1980) and Hoeve and Quirt (1984), is now widely accepted and has undergone considerable refinement (e.g., Kyser and Cuney, 2009). According to this model, the uranium ores formed as a result of the interaction of oxidizing basinal brines and reductants derived from the basement.

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1 This chapter has been published in a slightly modified form, "Cui, T., Yang, J., and Samson, I. M., 2012, Tectonic deformation and fluid flow: Implications for the formation of unconformity-related uranium deposits: Economic Geology, v. 107, p. 147-163".
Previous numerical modeling has indicated that thermally-driven convection may have been responsible for the leaching of uranium in the diagenetic model (Raffensperger and Garven, 1995; Cui et al., 2010). In these studies, however, convective fluid flow was restricted to the high-permeability basinal siliciclastic sequence, and therefore cannot explain the flow of such fluids into the basement sequence, where they could have been reduced, or the flow of reduced basement-derived fluids into the basin-fill sequence, where fluid mixing could have occurred. They also do not account for the intimate spatial association between deposits and faults, which were presumably important for trans-unconformity fluid flow.

Tectonic deformation has been considered to be an effective driving force for the migration of mineralizing fluids in various geological settings (e.g., Cox et al., 2001; Sibson, 2001), and a series of numerical studies have been carried out to investigate the interaction of fluid flow and tectonic deformation (e.g., Ord and Oliver, 1997b; McLellan et al., 2004; Zhang et al., 2008; Feltrin et al., 2009). Most notably, Oliver et al. (2006) concluded that extensional deformation may cause fluid flow across basement-cover interfaces during basin-related mineralization based on numerical modeling. Although new insights have been obtained from these studies, the influence of strain rate and permeability variations during deformation have been ignored in most simulations. Additionally, with the exception of an abstract by Schaub and Fisher (2009), no research on the role of deformation-driven flow in the formation of URU deposits has been published. The aim of this study is to bridge these gaps for the geological setting typical of URU deposits, and to examine the implications of the results for the formation of these deposits. First, a simplified hydrogeological model was constructed by integrating
existing geological, geophysical and geochemical data from the Athabasca, Thelon and Kombolgie basins, which host most of the Earth’s known URU deposits. Based on this model, a series of numerical experiments were carried out to study the relationship between tectonic deformation, thermal structure and mineralizing-fluid flow for URU deposits. The new constraints derived from this study for genetic models are then discussed, for example on the relationship between deposit style and tectonic setting.

3.2 Geological framework

3.2.1 The Athabasca basin

The Athabasca basin (Fig. 3.1) covers about 100,000 square kilometers in northern Saskatchewan and northwestern Alberta, and has a maximum formation age of ca. 1730 Ma (Alexandre et al., 2009). The basement is composed of Archean to Paleoproterozoic rocks that were metamorphosed during the Trans-Hudson Orogeny (ca. 1800 Ma) (Lewry and Sibbald, 1980). The overlying sedimentary basin fill is represented by the Athabasca Group (1-2 km thick), which mainly comprises quartz-rich sandstone and conglomerate from alluvial, fluvial and upper-shore sedimentary environments. Four major sequences, which are bounded by basin-wide unconformities, have been recognized in this group (Hiatt and Kyser, 2007). The basal sequence of the Athabasca Group (Manitou Falls and Fair Point Formations) consists of hematite-rich conglomerate and sandstone. Overlying the basal sequence is an arkosic formation of less-permeable marine sandstone, siltstone, and mudstone. In turn, this formation is capped by shale (Douglas Formation) and stromatolitic dolomites (Carswell Formation) (Jefferson et al., 2007).
Fig. 3.1. Regional geological map showing locations of the Athabasca and Thelon basins (highlighted by two rectangles) and major uranium deposits in them (modified from Jefferson et al., 2007). The rectangle in the index map indicates the location of the study area in Canada.

The diagenetic history of this basin (Fig. 3.2) has been investigated using petrography, stable isotope analysis, geochronology, and fluid inclusion analysis of the altered and unaltered rocks, which are proximal and distal to the uranium deposits, respectively (Hiatt et al., 2009). Early diagenesis of the Athabasca basin caused quartz overgrowth and hematite precipitation at 1600-1750 Ma (Fig. 3.2). Fluid inclusions from the quartz overgrowths have homogenization temperatures of 150-170°C and a salinity of ~25 wt percent NaCl equiv. Subsequently, the basinal brines underwent an increase in temperature, salinity and $\delta^{18}O$ as a consequence of burial. Finally, U/Pb dates from
Fig. 3.2. Simplified paragenetic relationships of the Athabasca, Thelon and Kombolgie basins with major tectonic events, which may have stimulated mineralizing fluid flow, associated with Laurentia (modified form Jefferson et al., 2007; Betts et al., 2008; Hiatt et al., 2009). U1 represents primary mineralization, AP = aluminum phosphate, APS = aluminum sulfate phosphate, FLAP = fluorapatite, H0 = primary hematite in the paleoweathered regolith, H1 and H2 = early diagenetic hematite in basal red mudstone beds, H3 = pervasive hematite, H4 = dark, intense hematite cement, Q1, Q2, and Q3 = quartz cement, XEN = xenotime. M1 is the Tans-Hudson Orogeny, M2 represents the orogenic events associated with the accretion of Nena, M3 is the Mackenzie Dikes, M4 is the Grenville Orogeny, and M5 marks the breakup of Rodinia.

Uraninite and Ar/Ar dates from syn-ore illite imply that the major uranium mineralizing event began at 1600 Ma (Fig. 3.2). Several later remobilization events have also been identified, which are interpreted to have occurred in response to far-field tectonic events at ca. 1400, 1270, 1100 and 850 Ma (Fig. 3.2) (Hiatt et al., 2009). The uranium deposits in the Athabasca basin represent the largest and highest-grade uranium deposits in the
world, and comprise the only uranium district in Canada that is actively being mined (Jefferson et al., 2007).

3.2.2 The Thelon basin

The Paleoproterozoic Thelon basin straddles the border between Nunavut and the Northwest Territories of Canada (Fig. 3.1). The following details about its evolution and sedimentary fill are based on the work of Palmer et al. (2004) and Rainbird and Davis (2007). This basin began to form after the Trans-Hudson orogeny (ca. 1.85 Ga) and is underlain by Archean to Paleoproterozoic metamorphosed rocks that were also affected by the Hudsonian orogeny and intruded by post-tectonic Hudsonian granites. The basinal succession includes three sequences, which, from oldest to youngest, are the Thelon Formation, Kuungmi Formation and Lookout Point Formation. The Thelon Formation comprises flat-lying conglomerate and sandstone with a thickness of 1-2 km. The overlying Kuungmi Formation is represented by an aphanitic basalt flow with a thickness of less than 10 m and a limited spatial distribution. This volcanic unit is overlain by the stromatolitic and dolomitic Lookout Point Formation, which has a thickness of less than 40 m.

Petrographic and fluid inclusion data indicate that the Thelon basin experienced a protracted diagenetic history (> 600 m.y.) (Renac et al., 2002; Hiatt et al., 2009). The diagenetic evolution (Fig. 3.2) has been divided into three stages: early diagenesis (1720-1650 Ma), peak diagenesis (1650-1000 Ma) and late diagenesis (after 1000 Ma), based on the nature of diagenetic mineral assemblages and on diagenetic conditions (e.g., temperature). Peak diagenesis and primary mineralization began when fluid temperature
reached about 200°C (Renac et al., 2002). The Thelon basin only hosts two areas with known uranium mineralization: the Boomerang Lake prospect and the Kiggavik deposit, which lie at the eastern and western margins of the basin, respectively (Fig. 3.1). The basin, however, holds substantial potential for uranium and other metal deposits due to its similarity with the Athabasca basin in terms of geological setting, characteristics, and tectonic evolution (Jefferson et al., 2007).

3.2.3 The Kombolgie basin

Unconformity-related uranium deposits contribute 20 percent of Australia’s total uranium resources. Most of the uranium mined in Australia since 1980 is from the Kombolgie basin, which forms the northern part of the larger McArthur basin (Hiatt et al., 2007). This basin is located in the Northern Territory of Australia (Fig. 3.3), and formed at about 1793 Ma, after the Barramundi Orogeny (1890-1870 Ma) and the Top End Orogeny (1863-1847 Ma) (Hiatt et al., 2007). The Kombolgie basin is floored by Archean to Paleoproterozoic gneisses and metasedimentary rocks; the latter are domed by pre-orogenic granitic intrusions and the Zamu Dolerite (1884 Ma). Overlying the steeply-dipping basement is the flat-lying Kombolgie Subgroup (1-2 km), which is mainly composed of sandstone and conglomerate with interlayered volcanic units. Hiatt (2007) divided the Kombolgie Subgroup into three sequences. The lowermost sequence evolved from proximal high-energy braided facies (coarse-grained sandstone and conglomerate) to distal low-energy braided-stream environments (coarse- to medium-grained quartz arenite). The middle sequence is interpreted as a coarse-grained fluvial facies, overlain by distal fluvial and interbedded marine and eolian facies. The uppermost sequence is
composed of distal fluvial and marine sedimentary rocks that suggest marine transgression.

![Geological Map](image)

Fig. 3.3. Simplified geological map showing important structures surrounding the Kombolgie basin (highlighted by the rectangle) and locations of major uranium deposits (modified from Lindsay, 2001). The rectangle in the index map indicates the location of the study area in Australia.

Several stages of sandstone diagenesis (Fig. 3.2) have been identified for the Kombolgie Subgroup (Kyser et al., 2000; Polito et al., 2006), and the following is a summary of the work of these authors. The early stage is characterized by the formation of quartz overgrowths (Fig. 3.2) at 80-130°C from low-salinity (< 10 wt %) NaCl fluids. The next stage is marked by the precipitation of illite and chlorite at temperatures of greater than 200°C at 1650 ± 80 Ma (Fig. 3.2). Quartz vein formation represents the third stage of diagenesis. Fluid inclusions from these veins have homogenization temperatures
of 200-400 °C and a salinity of about 22 wt percent NaCl equiv. The final stage of alteration is represented by widespread kaolinite precipitation (Fig. 3.2) in the Kombolgie group that penetrates several hundreds of meters depth away from the surface along fractures. Uranium-Pb and $^{207}\text{Pb}/^{206}\text{Pb}$ ratios of uraninite in the Jabiluka uranium deposit imply that primary uraninite precipitated at 1680 Ma, which was followed by remobilization at 1300, 1190 and 800 Ma (Fig. 3.2). Syn-ore illite crystallinity data and chlorite chemistry suggest a mineralization temperature range of between 150 and 250°C. The Nabarlek deposit is associated with a reverse fault/shear zone and primary mineralization formed from 200°C basinal brines at ca. 1640 Ma.

3.2.4 Far-field orogenic events and fault reactivation

Although the Athabasca and Thelon basins are categorized as intracratonic basins, their formation mechanisms are still enigmatic. Ramaekers and Catuneanu (2004) inferred that they formed by escape tectonics (the lateral extrusion of fault-bounded geological units resulting from tectonic compression) driven by far-field stresses from ongoing continent convergence. The basement rocks of these basins were affected significantly by the Trans-Hudson orogeny, and, as a result, a number of faults and shear zones formed in the basement rocks (Mercadier et al., 2010). These structures may have served as weak zones for fault reactivation and new fault generation during subsequent tectonic events. A series of accretionary orogenic events have been identified along the southern margin of Laurentia between 1800 and 500 Ma (Betts et al., 2008) (Fig. 3.2). These events may have induced far-field tectonic stresses that led to fault reactivation and the stimulation of fluid flow. Similarly, in the Kombolgie basin, paleomagnetic data...
indicate that there were several tectonic events that may have stimulated uranium mineralization and remobilization (Hiatt et al., 2009). Thus, these tectonic events provide geological constraints for the construction of the conceptual model and deformation-driven fluid flow modeling.

3.3 Conceptual model

In order to construct a numerical grid, it was necessary to create a simplified conceptual model that incorporates the most important features of these basins. Unfortunately, these basins have been subjected to significant erosion since they formed, and consequently complete present-day cross sections are not available. Nonetheless, previous studies have shown that these deposits, and the basins that contain them, have some common characteristics that provide the constraints necessary for the numerical experiments: 1. The deposits are located within and around an unconformity between low-permeability metamorphosed basement rocks and high-permeability thick hematite-rich sandstone units; 2. They are spatially associated with basement faulted structures and reduced lithologies; 3. The overlying high-permeability sandstone sequences are covered by relatively low-permeability shallow marine sedimentary facies; 4. Primary mineralization generally occurred at temperatures of about 200°C, 50-250 m.y. after basin formation; 5. The ore-hosting sedimentary basins experienced protracted fluid histories (e.g., more than 600 m.y. for the Thelon basin); and 6. A series of remobilization events occurred after the initial primary mineralizing event (Raffensperger and Garven, 1995; Cuney, 2005; Hiatt et al., 2009). An additional constraint is that the sedimentary fill must have been 6-7 km thick in order to attain the mineralizing
temperatures of around 200°C, assuming that the geothermal gradient was 30°C/km, which represents a typical intracontinental geothermal gradient (Fridleifsson et al., 2008).

Fig. 3.4. Idealized hydrogeological model with the most important features of basins that host unconformity-related uranium deposits. The hydraulic and mechanical properties for the various units are listed in Table 3.1.

Table 3.1. Major physical parameters of the various hydrological units

<table>
<thead>
<tr>
<th>Property</th>
<th>Cover</th>
<th>Sandstone</th>
<th>Basement</th>
<th>Fault</th>
<th>Water</th>
</tr>
</thead>
<tbody>
<tr>
<td>Density (kg·m⁻³)</td>
<td>2400</td>
<td>2500</td>
<td>2650</td>
<td>2400</td>
<td>1000</td>
</tr>
<tr>
<td>Bulk modulus (Pa)</td>
<td>2.8 × 10¹⁰</td>
<td>3.2 × 10¹⁰</td>
<td>4.95 × 10¹⁰</td>
<td>2.33 × 10¹⁰</td>
<td>2 × 10⁹</td>
</tr>
<tr>
<td>Shear modulus (Pa)</td>
<td>1.7 × 10¹⁰</td>
<td>4.0 × 10⁹</td>
<td>2.9 × 10¹⁰</td>
<td>3.0 × 10⁷</td>
<td></td>
</tr>
<tr>
<td>Cohesion (Pa)</td>
<td>3 × 10⁶</td>
<td>3 × 10⁶</td>
<td>4 × 10⁶</td>
<td>2 × 10²</td>
<td></td>
</tr>
<tr>
<td>Friction angle</td>
<td>30°</td>
<td>30°</td>
<td>30°</td>
<td>30°</td>
<td></td>
</tr>
<tr>
<td>Dilation angle</td>
<td>4°</td>
<td>4°</td>
<td>3°</td>
<td>4°</td>
<td></td>
</tr>
<tr>
<td>Effective porosity</td>
<td>0.15</td>
<td>0.2</td>
<td>0.1</td>
<td>0.2</td>
<td></td>
</tr>
<tr>
<td>Permeability (m²)</td>
<td>1.0 × 10⁻¹⁴</td>
<td>3.0 × 10⁻¹³</td>
<td>3.0 × 10⁻¹⁶</td>
<td>1.0 × 10⁻¹²*</td>
<td></td>
</tr>
<tr>
<td>Thermal conductivity (W·m⁻¹·°C⁻¹)</td>
<td>2.5</td>
<td>3.5</td>
<td>2.5</td>
<td>4.0</td>
<td>0.6</td>
</tr>
<tr>
<td>Thermal expansivity (°C⁻¹)</td>
<td>8 × 10⁻⁶</td>
<td>10 × 10⁻⁶</td>
<td>8 × 10⁻⁶</td>
<td>10 × 10⁻⁶</td>
<td>1.85 × 10⁻³</td>
</tr>
<tr>
<td>Specific heat capacity (J·kg⁻¹·°C⁻¹)</td>
<td>803.0</td>
<td>803.0</td>
<td>803.0</td>
<td>803.0</td>
<td>4185.0</td>
</tr>
</tbody>
</table>

* The permeability of faults is changed in some cases.
An idealized sandwich-like model has been constructed to reflect these shared geological features (Fig. 3.4). In the model, the Archean to Paleoproterozoic metasedimentary rocks and gneisses are represented by a low-permeability basement layer, and the overlying alluvial and fluvial sedimentary facies by a homogeneous sandstone layer, which is covered by a relatively low-permeability layer that represents the shallow marine sedimentary rocks. Heterogeneity and anisotropy with respect to permeability are ignored in the same layer. The actual faults with which URU deposits are associated have a variety of characteristics; they can be dextral or sinistral, extensional or transpressional, and ductile or brittle (Jefferson et al., 2007). Offsets on these faults vary significantly, depending on the local tectonic setting, and range from tens of meters to hundreds of meters, or in some cases, tens of kilometers. Despite this, a study of the influence of the scale of fault offsets on fluid flow is not attempted in the modeling reported here. The dimensions of different ore zones bounded by faults are also variable. The well known Koongarra ore body in Australia has a width of 50-100 m and extends up to 450 m along strike (Snelling, 1990). Another typical URU deposit, the Cigar Lake deposit in the Athabasca basin, shows an elongated body with an extending length of 2150 m and a maximum width of 100 m (Bruneton, 1987). For experiments involving deformation, a fault zone with representative dimensions of 300 m wide and 1000 m high is assumed to be located in the central part of the model, straddling the unconformity (Fig. 3.4). In order to reduce the influence of boundary conditions on fluid flow around the fault zone, the distance from the fault center to either boundary should be 10 times the half width of the fault zone (Itasca, 2005), which corresponds to a horizontal dimension of 6 km for the conceptual model. Despite the highly simplified nature of the
ore district-scale model, it will help us understand fluid flow around a fault zone during tectonic deformation.

In order to focus on the mineralizing zones around the basal unconformity, the topmost 3 km is not included in the modeling. The upper boundary was treated as having a fixed pore pressure (30 MPa assuming hydrostatic conditions) and stress plane (76 MPa from the overlying rocks and pore fluids). The side and bottom boundaries were assumed to be impermeable to fluid flow for models with the fault zone. A number of experiments without the fault zone were also conducted to investigate the pressure evolution during geo-mechanical deformation at various strain rates; permeable side boundaries for fluid flow were used in these models. For heat transport, the top boundary was maintained at 90°C (3 km depth), whereas the bottom was assigned a value of 240°C based on a geothermal gradient of 30 °C/km. The two side boundaries were insulated to heat transport. The base of the model was fixed vertically, but was free to move horizontally to account for tectonic deformation. The sides had a horizontal velocity during deformation that depends on the strain rate. The top was completely free to deform in both the vertical and horizontal directions to reflect the fact that topography changes during tectonic deformation. Hydrostatic pressure and a thermal field with a gradient of 30 °C/km were the initial conditions assigned to the system. This hydrostratigraphic model served as the basis for the numerical experiments described below. Various rock properties (Table 3.1) were determined based on data used in similar numerical modeling and published compilations (e.g., Raffensperger and Garven, 1995; McLellan et al., 2004; Yang et al., 2004; Oliver et al., 2006). According to the compilations of Pfiffner and Ramsay (1982) and Campbell-Stone (2002), most geologic strain rates range from $10^{-11}$
to $10^{-17}$ s$^{-1}$. No previous studies have investigated strain rates in these Proterozoic basins during uranium mineralization or deformation. Thus, a mid-range strain rate of $10^{-13}$ s$^{-1}$ was assigned at the side boundaries for most scenarios in the modeling. The interior of the modeled domain deformed according to its mechanical properties and boundary conditions.

3.4 Theoretical background

The modeling reported in this contribution focuses on the interaction between fluid flow and tectonic deformation. In some models, thermally-driven free convection was also coupled, but not for the influence of salinity distribution. Thermally-driven free convection is governed by a series of partial differential equations, including Darcy’s law, Fourier's law, mass conservation, energy conservation and equations of state (for details refer to Garven and Freeze, 1984a). In general, it is more difficult to describe the interaction between fluid flow and rock deformation due to the complicated deformation behavior of porous rocks.

The ore geology and petroleum geology communities view the role of faults in fluid flow differently. Faults are usually considered one of the most important fluid flow conduits for hydrothermal ore-forming processes (e.g., Oliver and Bons, 2001; Sibson, 2001; Cox, 2005), whereas, in petroleum systems, fault zones may act as an efficient seal for hydrocarbon traps or pathways for secondary oil migration (e.g., Aydin and Antonellini, 1994; Aydin, 2000). This difference may be attributed to the variable ways in which rocks mechanically deform, depending on porosity, temperature, pressure and lithology. Faults in low-porosity rocks typically include a thin low-permeability core that
is surrounded by a wider, high-permeability damage zone (Caine et al., 1996). As a result, these faults act as conduits for fluid flow in the direction parallel to fault planes, but tend to inhibit fluid flow across the faults. By contrast, faults in high porosity rocks usually comprise low-permeability deformation bands and, sometimes, discrete slip surfaces (Aydin, 2000; Rawling et al., 2001; Fossen et al., 2007). Barnicoat et al. (2009) proposed that the faults in sedimentary rocks at depths shallower than ~3 km do not show brittle behavior due to high porosity or poor lithification, except where the sediments are cemented abnormally early, or undergo tectonic inversion (uplift) after initial burial. For URU deposits, the mineralizing temperatures imply a burial depth of around 6 km. In addition, field observations show that mineralization occurred in brittle fractures (Kotzer and Kyser, 1995; Lorilleux et al., 2002), indicating that faults are characterized by discrete fractures and acted as fluid-flow pathways in the mineralizing process.

Elastic-plastic deformation of porous rocks in the middle and upper crust may be described approximately by the Mohr-Coulomb model. According to this model, rocks under loading deform elastically up to a yield point, after which they deform in a plastic manner. In the elastic deformation range, pore pressure varies due to the pore space change caused by the imposed stress. Yield occurs when the following relationship is satisfied,

\[ |\tau| = C - \sigma_n \tan \theta, \]

(3.1)

where \( \tau \) and \( \sigma_n \) are the maximum shear stress and the normal stress (Pa), respectively, across an arbitrary plane within the earth material, \( C \) is the cohesive strength of the porous rocks (Pa), and \( \theta \) is the angle of friction (Hobbs et al., 1976). Plastic deformation
after the yield point causes an increase of localized pore volume due to the frictional sliding of microfractures and grain boundaries (Ord and Oliver, 1997b). Such deformation-induced dilatancy is determined by a dilatancy angle in the Mohr-Coulomb model. For most common rocks, the dilation angle is between $+10^\circ$ and $+20^\circ$. For some high-porosity rocks, or rocks confined by extremely high pressure, plastic deformation, however, will reduce localized porosity, for which a negative dilatancy angle must be employed (Ord, 1991).

FLAC2D, a finite difference code ("FLAC," 2005), is capable of simulating the interaction of fluid flow, tectonic deformation and heat transport based on various material constitutive models. It has been applied successfully to investigate the ore-forming fluid evolution in a number of previously published studies (e.g., Ord and Oliver, 1997a; Oliver et al., 1999; Rutqvist et al., 2002; Feltrin et al., 2009). All of the presented numerical experiments in this chapter have been calculated using FLAC2D. The major governing equations of coupled mechanical-thermal-fluid flow processes are formulated within the framework of the quasi-static Biot theory for fully saturated rocks as follows ("FLAC," 2005).

1) Darcy’s law,

$$q = -\frac{k}{\mu_w} [\nabla P + \rho_w g \hat{n}],$$  \hspace{1cm} (3.2)

where $q$ is the Darcy flux of pore fluid flow (m/s), $k$ is the permeability (m$^2$), $\mu_w$ is the fluid dynamic viscosity (kg/(m·s)), $P$ is the fluid pressure (Pa), $\rho_w$ is the mass density of
the fluid (kg/m³), \( g \) is the gravitational acceleration (m/s²), and \( \mathbf{n} \) is the downward unit vector.

2) Fourier's law,

\[
q^T = -K^T \nabla T, \tag{3.3}
\]

where \( q^T \) is the heat-flux vector (W/m²), \( \nabla T \) is the temperature gradient (K/m), and the effective thermal conductivity of the saturated porous material \( K^T \) is defined as

\[
K^T = (1 - n)K_s^T + nK_w^T, \tag{3.4}
\]

where \( n \) is the effective porosity, \( K_s^T \) and \( K_w^T \) are the thermal conductivity of the solid matrix and the pore fluid, respectively (W/(m·K)).

3) The fluid mass conservation equation,

\[
\frac{\partial p}{\partial t} = M \left[ -\nabla \cdot (q) - \alpha \frac{\partial \varepsilon}{\partial t} + \left( n\beta_w + (1 - n)\beta_g \right) \frac{\partial T}{\partial t} \right], \tag{3.5}
\]

where \( \varepsilon \) is the volumetric strain of the porous medium, \( t \) is the time (s), \( \beta_w \) and \( \beta_g \) are the thermal expansion coefficient of the fluid and the rock grains (°C⁻¹), respectively, and \( M \) is the Biot modulus that is defined by

\[
M = \frac{K_w}{n + (a - n)(1 - a)\frac{K_w}{K_e}} \tag{3.6}
\]

where \( K_w \) and \( K_e \) are the bulk modulus of the fluid and the drained porous medium, respectively (Pa), and the Biot coefficient \( a \) is defined by
\[ \alpha = 1 - \frac{K_e}{K_s}, \]  

(3.7)

where \( K_s \) is the bulk modulus of rock particles (Pa).

4) The momentum balance equation,

\[ \frac{\partial \sigma_{ij}}{\partial x_i} + \rho g_i = \rho \frac{du_i}{dt} \text{ and } \rho = (1 - n) \rho_s + n \rho_w, \]  

(3.8)

where \( \sigma_{ij} \) is the stress tensor (Pa), \( \rho \) is the bulk density (kg/m\(^3\)), \( u_i' \) is the velocity (m/s), and \( \rho_s \) is the dry rock density (kg/m\(^3\)).

5) The energy balance equation,

\[ \rho C^T \frac{\partial T}{\partial t} + \nabla \cdot q^T + \rho_w^0 C_w q \cdot \nabla T - q_v^T = 0, \]  

(3.9)

where \( q^T \) is the thermal flux (W/m\(^2\)), \( q_v^T \) is the volumetric heat source intensity (W/m\(^3\)), \( \rho_w^0 \) is fluid reference density (kg/m\(^3\)), \( C_w \) is fluid specific heat (J/(kg·°C)), and \( C_T \) is the effective bulk specific heat, which is defined as

\[ C^T = [\rho_s C_v + n S \rho_0 C_w] / \rho, \]  

(3.10)

where \( C_v \) is the mean specific heat capacity (J/(kg·°C)) of the mineral grains, and \( S \) is degree of saturation, which has a value of 1 in this study.

6) The temperature dependence of fluid density,

\[ \rho_w = \rho_0[1 - \beta_w (T - T_0)], \]  

(3.11)
where \( T_0 \) is the reference temperature (°C).

7) FLAC2D (Version 5.0) contains eleven basic constitutive models, which describe various stress-strain relationships. All of these models can be coupled to fluid flow and heat transport. As an example, the coupled equation for an elastic material is

\[
\frac{\partial \sigma_{ij}}{\partial t} + \alpha \frac{\partial p}{\partial t} \delta_{ij} = 2G \left( \frac{\partial \varepsilon_{ij}}{\partial t} - \beta \frac{\partial T}{\partial t} \delta_{ij} \right) + \left( K_e - \frac{2}{3} G \right) \left( \frac{\partial \varepsilon_{kk}}{\partial t} - 3 \beta \frac{\partial T}{\partial t} \right) \delta_{ij},
\]

\[ \text{(3.12)} \]

where \( \sigma_{ij} \) is the total stress, \( \varepsilon_{ij} \) is the total strain, \( \delta_{ij} \) is the Kronecker delta, \( \beta \) is the bulk thermal expansion coefficient, and \( K_e \) and \( G \) are bulk and shear moduli of porous media (Pa) (for additional details refer to "FLAC," 2005).

By default, FLAC2D does not update the porosity table during the computing cycle in order to save calculation time ("FLAC," 2005). Volume variation, rather than porosity, is the main variable for equations in which deformation-induced pressure change is involved. In order to simulate the deformation process more accurately, porosity and permeability tables were updated according to the following scheme at every tenth step. The relationship between the porosity at any given step and the initial effective porosity is,

\[
n = 1 - \frac{V_0}{V} (1 - n_0),
\]

\[ \text{(3.13)} \]

where \( n_0 \) is the initial porosity, and \( V_0 \) and \( V \) are the initial and current volumes of an element (m³), respectively. According to the FLAC2D documentation ("FLAC," 2005), the volume strain may be approximated by
\[ \varepsilon = \frac{2(V-V_0)}{V+V_0}, \quad (3.14) \]

thus,

\[ n = 1 - \left( \frac{2-\varepsilon}{2+\varepsilon} \right) (1 - n_0). \quad (3.15) \]

An equation is consequently required to depict the change of permeability during deformation using the above known parameters. The well-known Kozeny-Carman model, which serves as a starting point for many of the subsequent models, is derived based on considering a porous medium as a bundle of tubes with different radii. For well-sorted, unconsolidated, and granular porous media, the Kozeny-Carman equation can be expressed as follows,

\[ k = \frac{n^3}{[5s_0^2(1-n)^2]}, \quad (3.16) \]

where \( k \) is the permeability (m\(^2\)), and \( s_0 \) is the solid surface exposed to the fluid per unit volume of solid material (Carman, 1956). Furthermore, \( s_0 \) can be related to the mean diameter \( d_m \) of granular particles by \( s_0 = 6/d_m \) (Bear, 1988), such that

\[ k = \frac{n^3}{(1-n)^2} \left( \frac{d_m^2}{180} \right). \quad (3.17) \]

This equation was employed by Ju and Yang (2010) in a numerical modeling study of a polymetallic deposit in southern China.

Since the advent of the Kozeny-Carman relationship, a variety of new models have been proposed for unconsolidated sediments and compacted rocks. These models...
generally fall into four categories, depending on their controlling parameters (Nelson, 1994): 1. models based on grain sizes and mineralogy; 2. models based on surface area and water saturation; 3. well log models; and 4. models based on pore dimension. For sedimentary rocks, most models show that porosity is linearly proportional to the log value of permeability. These models, however, all contain some unknown terms that require verification in specific cases. There are no available data to determine the parameters required for this study, and therefore, a simplified scheme was chosen. It is assumed that all of the deformation-induced porosity changes lead to permeability variations by,

\[
\frac{n - n_0}{n_0} = \frac{k - k_0}{k_0}.
\]  

(3.18)

Then,

\[
k = \frac{k_0}{n_0} \left(1 - \frac{2 - \varepsilon}{2 + \varepsilon} (1 - n_0) \right).
\]  

(3.19)

This is not consistent with the common linear log\(k\)-n relationship used for sedimentary rocks. The linear log\(k\)-n trend, however, is obtained based on the fact that some porosity does not contribute to the permeability of a rock due to mineral cementation and a lack of pore connection. Compared to the common log\(k\)-n relationship, this simplified model may subdue the permeability fluctuations caused by geo-mechanical deformation.
3.5 Results

Five models will be described in this section. Models 1 and 2 do not consider the influence of heat transport, whereas the remains are fully coupled thermal-mechanical-fluid flow models, although model 3 used a strain rate of zero.

3.5.1 Extensional deformation only (model 1)

Employing the parameters in Table 3.1, fluid migration and rock strain patterns are displayed in a series of snapshots (Figs. 3.5A, 3.5B and 3.5C) from a continuously run experiment for different degrees of strain at an extensional strain rate of $10^{-13}$ s$^{-1}$. The fluids show a general downward movement over the whole model. The fluid flow pattern is similar to this for strain of less than 1.5 percent. At strain values of more than 1.5 percent, the topography difference caused by deformation begins to affect the system, especially in the top part of the model (results for strain of greater than 1.0 percent are not shown here). The fault zone is the most important pathway along which the basinal brines in the sandstone sequence are focused into the basement. At the upstream end of the fault, fluids are sucked into the fault zone, then move along the fault, and then discharge into the basement from the downstream end. After 1.0 percent extensional strain (3,171 yrs) (Fig. 3.5C), the average fluid flow velocity around the unconformity interface is approximately $4.7 \times 10^{-3}$ m/yr, whereas the rate in the fault zone reaches a maximum of $3.2 \times 10^{-2}$ m/yr. Heterogeneity in rock properties leads to a non-homogeneous volumetric strain (Fig. 3.5D). The weak fault zone is subjected to the greatest positive volumetric strain of $1.0 \times 10^{-2}$.
Fig. 3.5. Fluid-flow patterns for different amounts of extensional strain at a strain rate of $10^{-13}$ s$^{-1}$. (A) 0.1 percent strain, (B) 0.5 percent strain, (C) 1.0 percent strain, and (D) the corresponding volumetric strain distribution and more detailed flow vectors around the fault for 1.0 percent strain. The vectors represent the Darcy flux and the positive volumetric strain values indicate mechanical expansion.

3.5.2 Compressive deformation only (model 2)

By contrast, during compressive deformation, for a strain rate of $10^{-13}$ s$^{-1}$ the fluids migrate up the fault due to the rapid increase of pore pressure in the low-permeability basement (Fig. 3.6). Figures 3.6A, 3.6C and 3.6D are also snapshots from a
continuously run experiment. The occurrence of abnormally high pressure in the basement is analogous to the development of overpressure in shale during a rapid burial of fluid-filled sediments, where pore fluids cannot easily escape. The fault zone, which transports fluids from the basement into the overlying sandstone units, still acts as the major conduits for fluid migration. As the strain increases, the elevated topography caused by deformation in the middle of the model causes a noteworthy local topography-driven flow (Figs. 3.6C and 3.6D). The average flow rate around the unconformity is $5.1 \times 10^{-3}$ m/yr for 1.0 percent strain, whereas the fluids in the fault migrate at up to $5.2 \times 10^{-2}$ m/yr. The fault zone still represents the most significant volumetric strain localization, with a maximum of $-1.0 \times 10^{-2}$ (Fig. 3.6).

3.5.3 Thermally-driven convection (model 3)

The modeling of Raffensperger and Garven (1995) and Cui et al. (2010) suggested that free convection may develop throughout the sandstone sequence in the absence of deformation. Consequently, an experiment was also carried out as part of this contribution to investigate thermally-driven flow and to serve as a starting point for further experiments that coupled heat transport to fluid flow and deformation. Using the properties in Table 3.1 and a geothermal gradient of 30 °C/km, the simulations suggest that it takes 6.6 m.y. to approach steady-state free convection, with an upwelling center in the middle of the model (Fig. 3.7). The temperature field is modified by the moving fluids, and two convection cells develop. The temperature is elevated significantly by advective heat transport around the upwelling center, and depressed close to the side.

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2 The top boundary for the flow system is assigned a constant pressure head. If topographic relieves occur during tectonic deformation, the elevation head along the top boundary becomes different, and so does the total hydraulic head. Thus, topographically-driven fluid flow occurs.
margins (Fig. 3.7). The magnitude of the temperature perturbations is directly related to the velocity of fluid

Fig. 3.6. Fluid-flow patterns for different amounts of compressive strain at a strain rate of $10^{-13}$ s$^{-1}$. (A) 1.0 percent strain, (B) the corresponding volumetric strain distribution and more detailed flow vectors for 1.0 percent strain around the fault, (C) 1.5 percent strain and (D) 2.0 percent strain. The vectors represent the Darcy flux and the negative volumetric strain values indicate mechanical contraction.
flow. In an individual convection unit, the fluid flow velocity decreases gradually towards the center of the cell (Fig. 3.7). The maximum flow rate is 0.9 m/yr along the unconformity. In addition, heat from the voluminous radiogenic granitoid intrusions in the basement, which is not taken into account in the presented model, may enhance such free convection.

![Fluid-flow patterns and thermal field for the thermally-driven only convection.](image)

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Fig. 3.7. Fluid-flow patterns and thermal field for the thermally-driven only convection. The vectors represent the Darcy flux. Only one upwelling center and two convection cells occur, given the parameters in Table 3.1. Temperatures are elevated in the upwelling center, and reduced at the margins due to downward flow. The initial geothermal gradient is 30 °C/km.

### 3.5.4 Coupling of thermally-driven convection and extensional deformation (model 4)

The introduction of extensional deformation at a strain rate of $10^{-13} \text{ s}^{-1}$ into model 3 suppresses the development of free convection cells (Fig. 3.8) and the whole system consists of downward flow. The fault still acts as a preferential path for fluid flow, with a maximum average linear velocity of 2.0 m/yr after 1.0 percent strain (3,171 yrs). This maximum average linear velocity is higher by almost two orders of magnitude than the
maximum in model 1, which only involved extensional deformation. Fluid flow in the sandstone away from the fault zone has a horizontal component, particularly near the basement (Fig. 3.8). This phenomenon is attributed to the interplay between the buoyancy caused by the thermal gradient and the hydraulic gradient induced by tectonic deformation. Thus, for a strain rate of $10^{-13}$ s$^{-1}$, the fluid flow pattern is dominantly influenced by tectonic deformation, but variable fluid-flow patterns will be expected depending on the configuration of heterogeneities of heat and deformation. Another interesting point is that there is an area that lacks fluid flow (zero flux) in the hanging wall of the fault (Fig. 3.8). Temperatures are reduced due to the downward flow, especially around the fault zone hosting fluids moving at high flow rates. Two models were also designed to investigate the influence of the initial fault permeability. Permeabilities of $1.0 \times 10^{-13}$ m$^2$ and $1.0 \times 10^{-14}$ m$^2$ were employed in the models shown in Fig. 3.8E and Fig. 3.8F, respectively. These show that fluids still prefer to flow through the fault zone, even at lower initial fault permeabilities.

3.5.5 Coupling of thermally-driven convection and compressive deformation (model 5)

A simulation that couples models 2 and 3 has also been carried out, the results of which are presented in Figure 3.9. As with model 4, mixed convection (the co-existence of thermally-driven free convection and deformation-driven fluid flow) is obtained. The highest average linear velocity of fluid flow in the fault zone is 7.0 m/yr (Fig. 3.9A), which is more than two orders of magnitude higher than that in model 2. The temperature is raised by about 20°C around the fault zone after 1.0 percent strain. The impact of the topography difference from deformation (Figs. 3.9C and 3.9D) is reduced significantly by
Fig. 3.8. Fluid flow patterns and thermal field for different amounts of extensional strain at a
strain rate of $10^{-13}$ s$^{-1}$. (A) 0.2 percent strain, (B) 0.6 percent strain, (C) 1.0 percent strain, (D) the expanded view of the fault zone for 1.0 percent strain, (E) 1.0 percent strain with a lower fault permeability ($1.0 \times 10^{-13}$ m$^2$), and (F) 1.0 percent strain with a much lower fault permeability ($1.0 \times 10^{-14}$ m$^2$). The vectors represent the Darcy flux. Note the zone of zero fluid flow on the left side of the fault.

the introduction of heat transport in comparison with the pure deformation model (Figs. 3.6C and 3.6D). The fluids are sucked into the fault from the surrounding basement rocks at the upstream end, migrate along the fault, and then discharge into the sandstone at the downstream end. As was the case in model 4, there is a barren area for fluid flow in the hanging wall of the fault. If a relatively low initial permeability is assigned to the fault zone, the fluid flux through the fault zone in the sandstone unit is reduced to some extent, depending on the permeability (Figs. 3.9E and 3.9F).

3.6 Discussion

3.6.1 Driving mechanisms of fluid flow

Fluids flow in response to the generation of a hydraulic gradient. The conceptual model and calculations described above show that extensional and compressive deformation leads to downward and upward flow, respectively, as a result of heterogeneous pressure variations. In order to better understand the driving mechanisms for such flow patterns, a number of experiments were run to determine the pressure evolution within different stratigraphic units during tectonic deformation at various strain rates (Fig. 3.10). It should be noted that these models did not consider the influence of heat transport. During extension, the basement displays a faster fluid pressure decrease than the overlying sandstone sequence because of the low permeability of the basement.
Fig. 3.9. Fluid flow patterns and thermal field for different amounts of compressive strain at a
strain rate of $10^{-13}$ s$^{-1}$. (A) 1.0 percent strain, (B) the expanded view of the fault zone for 1.0 percent strain, (C) 1.5 percent strain, (D) 2.0 percent strain, (E) 1.0 percent strain with a lower fault permeability ($1.0 \times 10^{-13}$ m$^2$), and (F) 1.0 percent strain with a much lower fault permeability ($1.0 \times 10^{-14}$ m$^2$). The vectors represent the Darcy flux. Note that there is also a zone of zero fluid flow on the left side of the fault in this model.

rocks (Fig. 3.10A). Thus, fluids migrate down into the underpressured basement, especially along the high-permeability fault zone. The underpressure becomes more significant as the strain rate increases (Fig. 3.10A). By contrast, compression results in overpressures in the whole system (Fig. 3.10B). The low-permeability basement, however, is subjected to a faster pressure increase than the clastic basin fill. Thus, a hydraulic head gradient develops between the basement and the sandstone. The pressure gradient within the two sequences varies according to the distribution of porosity and permeability as well as strain rates in the various units. The pressure gradient of the high-permeability clastic unit still remains hydrostatic, but it is different for the other low-permeability units (Fig. 3.10). The difference between the real pore pressure and hydrostatic pressure rises as the strain rate increases. As a result, during faulting, the fault zone may act as a fluid conduit that links the two reservoirs with different hydraulic heads. The heterogeneous permeability distribution also plays a key role in determining the fluid flow pattern during tectonic deformation.

The pore-pressure profiles also shift with the variation of initial pressure distribution. Although a hydrostatic pressure profile was assigned to the entire system as an initial condition in all of the models, overpressures are common in low-permeability rocks. If an initially overpressured basement is employed, the fluid flow velocity in the fault zone should be reduced for extensional deformation, and increased for compressive
deformation. Anisotropic permeability is another possible factor to impact the development of abnormal pressure. For sedimentary rocks, the ratio between the permeability along the bedding and the one perpendicular to the bedding is typically 2-10, depending on the rock types and fracture distribution. By contrast, the existence of steep fractures in the basement rocks may lead to a relatively high vertical permeability. Ignoring such anisotropy in the model may suppress the development of abnormal pressure.

Fig. 3.10. Fluid pore-pressure profiles through the centre of the conceptual model in Figure 3.4 with different strain rates for 1.0 percent strain. Note that heat transport was not considered in these models. (A) is for extensional deformation, and (B) is for compressive deformation. As the strain rate decreases, the profiles become closer to the hydrostatic profile. Below a threshold strain rate, the fluids will have enough time to flow freely to counterbalance the pressure variation resulting from deformation so that there is no abnormal pressure development.

In addition, it would be expected that the introduction of thermally-driven flow (a geothermal gradient) would cause a decrease of fluid flow rate during extensional deformation and an increase for compression because the upward buoyancy should
counteract the impact of the downward pressure gradient. The addition of free convection to both extensional and compressive deformation experiments, however, both resulted in a velocity increase (Figs. 3.8 and 3.9). This may be attributed to the nonlinear relationship between deformation-driven flow and free convection. Heat transport does not only affect fluid density and viscosity, but also the geo-mechanical deformation of porous media. Conversely, pressure variations resulting from mechanical deformation influences fluid flow, and consequently, heat transport.

3.6.2 Implications for uranium sources and deposit genesis

Disagreement still exists regarding the source of uranium in URU deposits. Archean granite, Paleoproterozoic sediments, leucogranites, pegmatoids, and calc-alkaline granitoids in the basement all represent possible and favorable uranium sources due to their high uranium content (Cuney, 2005). The question of how and why the oxidized basinal brines, however, may have penetrated the low-permeability basement and extracted uranium has not been satisfactorily answered. The relatively small volumes of alteration in the basement also constrain any model in which the uranium originated from the basement (Jefferson et al., 2007). Free convection cells will penetrate the basement where the permeability difference between the sedimentary cover and the metamorphic basement is less than two orders of magnitude (Oliver et al., 2006). Based on compilations of permeability data for rocks (Freeze and Cherry, 1979; Garven and Freeze, 1984b), the Archean to Paleoproterozoic crystalline basement below the basins in question is predicted to have an extremely low permeability that is more than two orders of magnitude lower than that of sandstone. The study by Mercadier et al. (2010),
however, showed that basinal brines may flow into the basement to depths of hundreds of meters below the unconformity through a dense network of microfractures. In contrast to the high uranium content of basement rocks, the current average uranium content of the sandstone units, away from the mineralized areas, is below 1 ppm (Cuney, 2005). However, as these siliciclastic sediments mainly originated from the surrounding basement (Palmer et al., 2004), their original uranium content would have been similar to that of the current basement below these basins. Assuming that the eroded basement that was the source for the siliciclastic sediments had a similar composition and uranium content to the present basement, the low uranium concentrations in the sandstone units suggests that these sedimentary rocks have been leached of their uranium either during the transport process at surface or after their sedimentation. One possible supposition that stems from this is that the missing uranium is now concentrated in the URU deposits (Jefferson et al., 2007).

This study, along with previous modeling (Raffensperger and Garven, 1995; Cui et al., 2010), shows that thermally-driven free convection may have occurred pervasively through the sandstone sequence, which provides a viable mechanism for the leaching of uranium. The modeling presented above suggests a mechanism by which uranium-bearing basinal brines could have interacted with fluids or reduced lithologies in the basement. Oxidized uraniferous basinal brines would have been sucked down into the basement during extensional deformation (Fig. 3.8), where they could have interacted with reduced lithologies and/or fluids, precipitating uranium. By contrast, during compression, reduced fluids would have been expelled from the basement upwards along faults into the sandstone sequence to mix with oxidized basinal brines to form deposits.
(Fig. 3.9). Away from the fault zone, fluid interaction between the sandstone sequence and the basement rocks also could have occurred along the unconformity, although the fluid fluxes are much lower than that in the fault zone (Figs. 3.8 and 3.9). Uranium could have been leached out from basement during this process if tectonic extension and compression happened alternately. In addition, it is worthwhile noting that although a very low, albeit reasonable, basement permeability was used in this study, it is possible that the shallow part of the basement had higher permeability due to the presence of microfractures and paleoweathering. More studies are required to test the actual permeability of the basement in order to perform more accurate simulations.

Based on the modeling, an association should exist between basement-hosted deposits and extensional deformation, and between the deposits above or straddling the unconformity and compressive deformation. This prediction is supported by the deposit locations and tectonic history in the Kombolgie basin. After the Kombolgie basin formed, an extensional event occurred at ca. 1730 Ma (Lindsay, 2001), which may explain why most of the URU deposits in the Kombolgie basin are basement-hosted (Kyser and Cuney, 2009). By contrast, the situation for URU deposits in the Canadian basins appears to be more complicated. Northern Laurentia experienced a long period of crustal stability after ca. 1.8 Ga, which is represented by intracontinental and passive-margin sedimentation related to limited extension (Davidson, 2008). To the south, it is characterized by a range of late Paleoproterozoic orogenic events (Fig. 3.2). Neither the details of the tectonic evolution of the Athabasca and Thelon basins, nor the localized tectonic stress states around deposits at the time of mineralization, however, are well known. According to the compilation by Gandhi (2007), for 66 well-known deposits in
the Athabasca and Thelon basins, 32 of them are basement-hosted, and the other 34 are just above or straddling the unconformity. This distribution may be reflective of a complex tectonic deformation history in these basins. Further work is required to investigate the relationship between the deposit locations and the tectonic history, which may provide new insights to the genesis of URU deposits if enough data are obtained.

3.6.3 Limitations

The finite difference method is employed in FLAC2D to solve its governing equations ("FLAC," 2005). This reduces the massive memory requirements of other computational methods and allows modeling to be accomplished in a reasonable time period using desktop computers. A small time step, however, is required in order to reduce computational error, which increases the computing time significantly, especially for a fully coupled thermal-mechanical-fluid flow model. Computing time is also determined by the sizes of the grid. These limitations restrict the modeling time to thousands of years and a relatively small model domain. The FLAC2D package supports a large-strain mode, in which the material can yield and flow, and the grid can be updated with the coordinates to express large amounts of deformation. On the other hand, FLAC2D still uses continuous media to represent its simulated objects. As the strain increases to a critical value that varies with the problem of interest, a "bad geometry" error will appear. This is why the modeling was run to a maximum strain of 2% in most cases. Computing time is also a constraint of the modeling; in this study an experiment using 2% strain (6342 years for a strain rate of $10^{13}$ s$^{-1}$) required weeks to months to run. The modeling time scale is very short compared to the time scale of basin evolution and
mineralizing processes (tens to hundreds of millions of years). Tectonic deformation, however, is, in reality, a cyclical and multi-episodic process that includes strain accumulation, yield/failure, and cementation (although cementation is not considered in this study). The time scale for every single cycle is unknown, but this modeling, at the very least, provides new insights into the fluid flow patterns that develop during a deformation cycle.

3.7 Conclusions

Numerical modeling of fluid flow in and below several typical Proterozoic siliciclastic basins that host unconformity-related uranium deposits indicates that:

1. Thermally-driven free convection may develop throughout the sandstone sequence under a normal geothermal gradient (25–35 °C/km) during tectonically quiet periods. Such convection, however, is easily suppressed or disrupted by tectonic deformation. Thus, steady-state free convection may not have readily developed due to the need for a long period of tectonic quiescence (millions of years). Mixed convection and non-steady-state thermally-driven convection could have dominated the basins at most times as minor tectonic events usually accompany basin evolution.

2. During extensional deformation at a strain rate of $10^{-13}$ s$^{-1}$, the low-permeability basement experiences a faster pressure reduction than the sedimentary cover. Thus, basinal brines are sucked into the basement, mainly along fault zones. In contrast, a more rapid accumulation of pore pressure occurs in the basement during compressive
deformation. As a result, fluids are expelled from the basement into the overlying clastic rocks, also mainly through fault zones.

3. The specific fluid flow pattern that develops is determined by the relative roles of strain rate and a variety of hydraulic properties. For relatively high strain rates, deformation-driven flow dominates the system. Mixed convection, however, can occur at low strain rates, which means that deformation-driven flow dominates around the fault zone (high-strain localization), whereas thermally-driven flow controls the remainder of the model. The critical strain rate, which divides the deformation-dominated system and the heat transport control system, varies depending on the specific properties of fluids and rocks.

4. A tentative model is proposed in which oxidized basinal brines leach uranium from the clastic basin fill by free convection during periods of tectonic quiescence (Fig. 3.11A). When tectonic compression occurs, basement-derived reduced fluids may flow up along fault structures and mix with uranium-bearing basinal fluids to form sandstone-hosted deposits (Fig. 3.11B). Basement-hosted deposits will tend to develop within zones that have undergone localized extensional deformation, where oxidized uranium-bearing brines migrate down faults into the basement rocks and encounter reduced minerals or fluids (Fig. 3.11B). Thus, primary mineralization and remobilization of uranium are controlled by both free convection and tectonic deformation resulting from far-field tectonic events. This, however, does not preclude the possibility that uranium was leached out from the basement rocks.
Fig. 3.11. Schematic diagrams illustrating an idealized genetic model for unconformity-related uranium deposits. Arrows indicate fluid flow directions. (A) Free convection forms in the sandstone sequence during periods of tectonic quiescence. (B) The interplay between oxidized basinal brines and reduced fluids or minerals derived from the basement rocks during tectonic deformation.
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Chapter 4

Uranium transport across basement/cover interfaces by buoyancy-driven thermohaline convection: implications for the formation of unconformity-related uranium deposits

4.1 Introduction

Unconformity-related uranium (URU) deposits, which are mainly hosted by Paleoproterozoic sedimentary basins in Canada and Australia, show impressive U tonnages and grades (up to 200 kt U and up to 20% U on average) (Kyser and Cuney, 2009; Richard et al., 2011b). They are located within or around basal unconformities between Proterozoic basin fill and underlying Archean granitoid gneisses and Paleoproterozoic metamorphosed sedimentary rocks, where reductants and faults exist (Jefferson et al., 2007). Although intense research has been carried out on these deposits (e.g., Hoeve et al., 1980; Raffensperger and Garven, 1995b; Fayek and Kyser, 1997; Cuney, 2005; Richard et al., 2011a), the processes of uranium leaching, transport and precipitation are still not clearly understood. The diagenetic model (Hoeve et al., 1980; Hoeve and Quirt, 1984), in which ore formation and host rock alteration are directly linked to stages of basin diagenesis and evolution, is now widely accepted and has undergone considerable refinement (e.g., Fayek and Kyser, 1997; Cuney, 2005).

1 This chapter has been submitted to the "American Journal of Science" in a slightly modified form.
According to this model, the URU ores formed as a result of the interaction of oxidizing basinal brines and reductants derived from basement. Furthermore, different versions of the diagenetic models have proposed two possible uranium sources (Kyser and Cuney, 2009). One considers the overlying basinal sediments as the source of both uranium and oxidizing brines (e.g., Fayek and Kyser, 1997; Kyser et al., 2000), and the other uses the basement as the uranium source (e.g., Annesley and Madore, 1999; Hecht and Cuney, 2000; Cuney, 2005; Mercadier et al., 2010; Richard et al., 2010). Previous numerical modeling has shown that thermally-driven convection is likely to have occurred within the siliciclastic sequence overlying the basement (Raffensperger and Garven, 1995b; Cui et al., 2010), and this type of convection has been considered to be responsible for uranium uptake and transport in sedimentary basins (Kyser and Cuney, 2009). In these studies, convective fluid flow, however, was restricted to the high-permeability basinal siliciclastic strata, and therefore cannot explain models in which basinal brines penetrate the basement rocks, as proposed in some recent studies (Boiron et al., 2010; Mercadier et al., 2010; Richard et al., 2010). These studies showed that basinal brines may flow into the basement to depths of hundreds of meters below the unconformity, not only along major faults, but also through a dense network of microfractures. Cui et al. (2012) conducted coupled thermo-hydro-mechanical modeling that suggested that fluid interaction between basement and basinal fill occurs along major faults during tectonic deformation. Reactivation of preexisting basement structures and the generation of new faults suppressed free convection and led to deformation-dominated fluid flow or mixed convection, depending on strain rates. During compressive deformation, basement-derived brines are squeezed upward along faults. In contrast, basinal brines are able to be
sucked into basement rocks through major fault structures. To build on these recent studies (Boiron et al., 2010; Cui et al., 2010; Mercadier et al., 2010; Richard et al., 2010; Cui et al., 2012), a series of numerical modeling experiments have been conducted to address the following questions: 1. Is it possible that basinal brines penetrate basement by buoyancy-driven thermohaline convection during tectonically quiet periods? and, 2. If fluid mixing and/or water-rock interaction in the basement can be achieved by thermohaline convection, is this kind of convection capable of carrying enough uranium from the uranium-rich basement to a favorable environment to form large URU deposits?

4.2 A brief introduction of URU deposits and their geological setting

Unconformity-related uranium deposits are found near the unconformity between Paleoproterozoic basins and their Archean to Paleoproterozoic basement rocks. The most significant basins hosting URU deposits that are currently mined are the Athabasca and Kombolgie basins in Canada and Australia, respectively (Fig. 4.1A). In 2010, Canada accounted for 18% of world uranium production, followed by Australia at 11% ("World Uranium Mining," 2011). Most of this uranium is from URU deposits. In addition, the Thelon basin in Canada hosts two areas with known uranium mineralization: the Boomerang Lake prospect and the Kiggavik deposit, which lie at the eastern and western margins of the basin, respectively (Fig. 4.1A). The basin, however, is believed to hold substantial potential for uranium and other metal deposits due to its similarity to the Athabasca basin in terms of geological setting, paragenesis, and tectonic evolution (Jefferson et al., 2007). Unconformity U deposits are known from other settings, such as the Cuddapah basin, India (Banerjee, 2005) and the Pasha-Ladoga basin, Russia.
(Velichkin et al., 2005), however, the conceptual model used in this study is based on the characteristics of the Athabasca, Thelon and Kombolgie basins.

Fig. 4.1. (A) Regional geological map showing locations of the Athabasca and Thelon basins and major uranium deposits in them (modified from Jefferson et al., 2007). The rectangle in the index map indicates the location of the study area in Canada. (B) A simplified lithostratigraphic architecture of cross section B-B” in the Athabasca basin (modified from Hajnal et al., 1997). MF
= Manitou Falls Formation; LzL = Lazenby Lake Formation; FP = Fair Point Formation; WP = Wolverine Point Formation; LL = Locker Lake Formation; OF = Otherside Formation; TL = Tuma Lake Formation; DG = Douglas Formation; and CW = Carswell Formation.

The Athabasca and Thelon basins cover about 200,000 square kilometers in Northern Saskatchewan (Fig. 4.1A), Western Alberta, and the Northwest Territories of Canada, and have maximum formation ages of ca. 1730 Ma and 1753 Ma, respectively (Hiatt et al., 2003; Alexandre et al., 2009; Hiatt et al., 2009). The basement comprises Archean to Paleoproterozoic rocks, and the overlying sedimentary basin fill is represented by the Athabasca and Thelon groups (1-2 km preserved thickness). The basin fill mainly consists of quartz-rich sandstone and conglomerate deposited in alluvial, fluvial and supratidal environments. Four major sequences, which are bounded by basin-wide unconformities, have been recognized in the Athabasca group (Hiatt and Kyser, 2007). The basal sequence of the Athabasca Group (Manitou Falls and Fair Point Formations) consists of hematite-rich conglomerate and sandstone (Fig. 4.1B). Overlying the basal sequence is an arkosic formation of less-permeable marine sandstone, siltstone, and mudstone. In turn, this formation is capped by shale (Douglas Formation) and stromatolitic dolostones (Carswell Formation) (Jefferson et al., 2007). The sedimentary rocks in the Thelon basin include three sequences, which are, from oldest to youngest, the Thelon Formation, Kuungmi Formation, and Lookout Point Formation (Hiatt et al., 2003). In the Athabasca Basin, U-Pb dates of uraninite and Ar-Ar dates of syn-ore illite imply that the major uranium mineralization event began at 1600 Ma. Several subsequent remobilization events, which have been interpreted to reflect far-field tectonic events, have been identified at 1400 Ma, 1270 Ma, 1100 Ma, and 850 Ma (Alexandre et al.,
Studies in the Thelon basin indicate that peak diagenesis and primary mineralization began when fluid temperatures reached about 200 °C (Renac et al., 2002).

The Kombolgie basin, the northern part of the larger McArthur basin, is located in the Northern Territory of Australia and formed at ca. 1793 Ma. It is underlain by Archean and Paleoproterozoic gneisses and metasedimentary rocks. Overlying the steeply dipping basement, is the flat-lying Kombolgie Subgroup (1-2 km), which is mainly composed of sandstone and conglomerate with interlayered volcanic units. The uppermost sequence of the group is composed of distal fluvial and marine sediments that indicate a marine transgression. The U-Pb and $^{207}\text{Pb}/^{206}\text{Pb}$ ratios of uraninite in the Jabiluka uranium deposit imply that primary uraninite precipitated at 1680 Ma, followed by remobilization at 1300, 1190, and 800 Ma. Syn-ore illite and chlorite suggest a mineralization temperature of around 200 °C (Kyser and Cuney, 2009).

4.3 Methodology

4.3.1 Governing equations and code

Simulating variable-density fluid flow involves numerically solving a group of equations under specific boundary and initial conditions. These equations include Darcy's law, Fourier's law, and conservation of energy and mass. In this study, the computer code FEFLOW was used to solve these equations. Details of these equations and how to solve them numerically can be found in Garven and Freeze (1984a), Kolditz et al. (1998), Diersch and Kolditz (1998), and Diersch and Kolditz (2002). The built-in equations of state (EOS) in FEFLOW are only implemented for fluids with temperatures between 0 and 100 °C (Diersch, 2009). In order to accommodate the wide range of temperature,
pressure and salinity values used in this study, the original EOS was replaced in FEFLOW by

$$\rho^f(p, T, C) = 999.843633188666 - 100^2 q_2 - 100q_1 + (q_2 p^2 + q_1 p) + z(p, C) T^6 + w(p, C) T^5 + v(p, C) T^4 + u(p, C) T^3 + s(p, C) T^2 + r(p, C) T + (j_0 + j_1 p + j_2 p^2) C^2 + (h_0 + h_1 p + h_2 p^2) C,$$

(4.1)

$$z(p, C) = z_0 + z_1 p + z_2 p^2 + z_3 C + z_4 C^2,$$

(4.2)

$$w(p, C) = w_0 + w_1 p + w_2 p^2 + w_3 C + w_4 C^2,$$

(4.3)

$$v(p, C) = v_0 + v_1 p + v_2 p^2 + v_3 C + v_4 C^2,$$

(4.4)

$$u(p, C) = u_0 + u_1 p + u_2 p^2 + u_3 C + u_4 C^2,$$

(4.5)

$$s(p, C) = s_0 + s_1 p + s_2 p^2 + s_3 C + s_4 C^2,$$

(4.6)

$$r(p, C) = r_0 + r_1 p + r_2 p^2 + r_3 C + r_4 C^2,$$

(4.7)

where $h_0, h_1, j_2, j_0, j_1, j_2, q_1, q_2, z_0, z_1, z_2, z_3, z_4, w_0, w_1, w_2, w_3, w_4, v_0, v_1, v_2, v_3, v_4, u_0, u_1,$

$u_2, u_3, u_4, s_0, s_1, s_2, s_3, s_4, r_0, r_1, r_2, r_3$ and $r_4$ are constant coefficients (for details refer to Magri, 2010). This new EOS can describe the density changes for fluids within the ranges of $0.1 \leq p \leq 100$ MPa, $0 \leq T \leq 350$ °C and $0 \leq C \leq 350$ g/l. This extended EOS has been verified and used in a number of published studies (e.g., Magri et al., 2008; Magri et al., 2009a; Magri et al., 2009b).

### 4.3.2 Conceptual model

The Athabasca, Thelon and McArthur basins all have been interpreted as intracratonic basins (Plumb, 1979; Ramaekers, 1981; Hoeve and Quirt, 1984; Eaton and
Darbyshire, 2010), although their evolution and hydraulic architecture are still not completely understood. Intracratonic basins are generally considered to have formed as a result of some combination of lithospheric flexure, thermal contraction, isostatic balance and dynamic topography (Miall, 2000). These basins have been subjected to significant erosion since they formed, which negates the availability of complete present-day cross sections. Fortunately, the major characteristics of such basins can be gleaned from their Phanerozoic counterparts, such as the Michigan basin in North America (Leighton and Kolata, 1991), the Clarence-Moreton basin in Australia (Ingram et al., 1996) and the Ghadames basin in Libya (Dardour et al., 2004; Hassan, 2009). All of these well-known intracratonic basins are relatively symmetrical and have a very smooth basement relief, consistent with the preserved portion of the Athabasca basin (Fig. 4.1B). In addition, previous studies (e.g., Hoeve and Quirt, 1984; Raffensperger and Garven, 1995b; Cuney, 2005; Kyser and Cuney, 2009; Richard et al., 2010) have shown that: 1. The deposits are located within and around an unconformity between low-permeability metamorphosed basement rocks and high-permeability thick hematite-rich sandstone units with thickness of 1-2 km; 2. They are spatially associated with basement fault structures; 3. The overlying high-permeability sandstone sequences are covered by relatively low-permeability shallow marine sedimentary facies; 4. Primary mineralization generally occurred at about 200°C, requiring a sediment thickness of 6-7 km, assuming a typical intracontinental geothermal gradient of 30°C/km (Fridleifsson et al., 2008); 5. The ore-hosting sedimentary basins experienced protracted fluid histories (e.g., more than 600 m.y. for the Thelon basin); and 6. A series of remobilization events occurred after the initial primary mineralizing event.
A simplified sandwich-like model has been constructed to integrate these geological features (Fig. 4.2). This model also resembles the calculated profiles of typical intracratonic and rift basins formed through tectonic subsidence and lithospheric stretching by Watts et al., (1982). The Archean to Paleoproterozoic metasedimentary rocks and gneisses are represented by a low-permeability basement unit, and the overlying alluvial and fluvial sedimentary facies by a homogeneous sandstone sequence, which is covered by two relatively low-permeability layers that represent the shallow-marine sedimentary rocks. Hydraulic conductivity was assumed to be constant throughout a given unit. The currently preserved portions of the Athabasca, Thelon and Kombolgie basins all have a horizontal dimension of more than 300 km (Fig. 4.1). Given that portions of these basins have been eroded, the model basin is assumed to have been 450 km wide, and the horizontal dimension of the whole model is 600 km (Fig. 4.2).
Fluid flow, heat transport and solute transport have been modeled in this study. The side and bottom boundaries were assumed to be impermeable to fluid flow. The top boundary was assigned a constant hydraulic head of 0.0 m to reflect a flat water table. For heat transport, the top boundary was maintained at 20°C, whereas the bottom was assigned a temperature of 320°C based on a geothermal gradient of 30 °C/km. The two side boundaries were insulated to heat transport. Hydrostatic pressure and a thermal field with a gradient of 30 °C/km were the initial conditions assigned to the system. The influence of a range of geothermal gradients has been discussed by Cui et al. (2010).

Salinities of basinal brines generally increase with depth at various rates depending on the geological setting. The rate of increase usually decreases gradually after a critical depth, and even becomes negative in some cases (Kharaka and Hanor, 2003). According to a number of fluid inclusion studies of URU deposits (e.g., Derome et al., 2005; Richard et al., 2010; Richard et al., 2011a), a typical salinity for NaCl-rich brines in the Athabasca basin is 300,000 mg/l, which are considered to be the predominant fluids in the sandstone unit, and 350,000 mg/l for the CaCl₂-rich fluids resulting from the interaction between NaCl-rich digenetic brines and Ca-rich rocks in the basement. Based on the studies mentioned above and typical salinity profiles in intracratonic basins, in the model, a salinity of 50,000 mg/l was used for the first 500 m, then the salinity was increased at a rate of 100 mg/l/m to a maximum of 300,000 mg/l (Fig. 4.3A). The basement fluids were assigned a salinity of 340,000 mg/l (the equation of state only supports a salinity range of 0-350,000 mg/l, but is reasonable given the fluid inclusion data available for the basement fluids). Salinities of 50,000 mg/l and 340,000 mg/l were specified at the top and bottom boundaries, respectively. The horizontal side boundaries
Fig. 4.3. Initial values of background salinity (A) and its evolution (B, C, and D) during modeling at 0.1, 1.0 and 5.0 m.y.. The white numbers on the isolines identify the localized salinity (g/l; uranium concentration is not included).
were assumed to be impermeable for solute transport. Various rock and fluid properties at reference conditions \((T_0 = 0^{\circ}C, P_0 = 100 \text{ kPa} \text{ and } C_0 = 0 \text{ mg/l})\) (Table 4.1) were determined based on data used in similar numerical modeling, published compilations and the previous studies in this thesis (e.g., Garven and Freeze, 1984b; Raffensperger and Garven, 1995b; McLellan et al., 2004; Yang et al., 2004; Oliver et al., 2006; Cui et al., 2010; Singhal and Gupta, 2010; Cui et al., 2012), and are appropriate for the geological setting of the Athabasca, Thelon and Kombolgie basins. The hydraulic conductivity of sandstones is in the range \(10^{-13} \text{ to } 10^{-5} \text{ m/s}\) (Freeze and Cherry, 1979; Garven and Freeze, 1984b). The sandstone layers in these Proterozoic basins mainly comprise conglomerate and sandstone deposited in alluvial and fluvial environments. Usually, it is expected to find highly permeable rocks in this kind of geological setting. Thus, a relatively high value of \(K = 3.0 \times 10^{-6} \text{ m/s}\) was used in this modeling. In addition, this value is consistent with the previous modeling in this thesis, so that the results can be compared among these studies. Values of longitudinal dispersivity are dependent on the scale at which they are measured. According to the data compilation by Gelhar (1986), the ratio of longitudinal

<table>
<thead>
<tr>
<th>Hydrogeological unit</th>
<th>(K_x \text{ (m/s)})</th>
<th>(K_x/K_z)</th>
<th>(\lambda \text{ (J/m/s/°C)})</th>
<th>(\Phi)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Carbonate</td>
<td>(3.0 \times 10^{-8})</td>
<td>10</td>
<td>3.0</td>
<td>0.15</td>
</tr>
<tr>
<td>Shale</td>
<td>(3.0 \times 10^{-8})</td>
<td>10</td>
<td>2.5</td>
<td>0.15</td>
</tr>
<tr>
<td>Sandstone</td>
<td>(3.0 \times 10^{-6})</td>
<td>2.5</td>
<td>3.5</td>
<td>0.20</td>
</tr>
<tr>
<td>Basement</td>
<td>(3.0 \times 10^{-11})</td>
<td>1</td>
<td>2.5</td>
<td>0.10</td>
</tr>
</tbody>
</table>

Notes: some of these parameters will be adjusted in different models, see text for details, \(K_x\) is the horizontal hydraulic conductivity, \(K_z\) is the vertical hydraulic conductivity, \(\lambda\) is the thermal conductivity, and \(\Phi\) is the effective porosity.
dispersivity to its measured scale is approximately 0.001 when the problem is of above 100 km in scale; the ratio is between 0.01 and 0.1 km for the measured scale ranging 0.1 to 10 km. Therefore, 600 m was used for the longitudinal dispersivity, and 150 m was used for transverse dispersivity since it is usually 4 to 5 times less than the longitudinal one.

Fig. 4.4. Fluid flow patterns (A, C) and temperature field (B) of the basic model using parameters in Table 4.1 at 5 m.y. modeling time. Darcy fluxes are indicated by "bullets" that are coloured depending on the localized Darcy flux in (A) and (C). The thicker ends of "bullets" point to fluid flow directions; some blue arrows are also drawn to help trace fluid flow.
Initial uranium conditions are specified based on recent fluid inclusion analyses (Richard et al., 2010; Richard et al., 2011a; Richard et al., 2011b). The average uranium concentration of seawater is 0.003 mg/l (Emsley, 2003). Uranium concentrations in basinal brines vary over several orders of magnitude (from 0.001 mg/l to 1000 mg/l) depending on redox conditions, temperature, solution composition, and pH (e.g., Weber Jr and Sackett, 1981; Kraemer and Kharaka, 1986; Hodge et al., 1996). Uranium concentrations of 1-1000 mg/l are expected in basinal brines with high chlorinity at 155 °C (Richard et al., 2010). In the Athabasca basin, based on fluid inclusion analyses, the uranium concentration (100-530 mg/l) in the CaCl₂-rich brines is one to two orders of magnitude higher than that of ~1 mg/l in the NaCl-rich brines, and it has been proposed that the CaCl₂-rich brines were responsible for leaching uranium from basement source rocks (Richard et al., 2010). Thus, a uranium background concentration of 1 mg/l was used in the modeling. The brines in the uranium source area have been assigned a uranium concentration of 500 mg/l. Chemical reactions were not considered during solute transport. All the boundaries are impermeable to uranium transport.

4.4 Results

4.4.1 Fluid flow and temperature distribution

Figure 4.4 shows the fluid flow patterns and temperature field at 5 m.y. based on application of the hydraulic properties in Table 4.1 to the conceptual model in Figure 4.2. Free convection cells not only develop within the highly permeable sandstone sequence as suggested by previous studies (Raffensperger and Garven, 1995a; Cui et al., 2010), but also penetrate into the overlying shale and the underlying basement to some depth. Fluid flow velocities vary over a range of nine orders of magnitude across the basin. The
traditional and most commonly used arrows, where arrow heads point out fluid flow directions and the length of arrows is proportional to magnitude of fluid flow, cannot depict the velocity distribution in the modeling. Thus, the colorful "bullets" (the equal-sized short line segments) were used to present fluid flow patterns (Figs 4.4A, 4.4C). The "bullets" are oriented and colored according to local fluid flow velocities. The colors of "bullets" change depending on the magnitude of fluid flow, and the dark ends of "bullets" point out fluid flow directions. The fluid flow "bullets" for the whole basin, however, were not presented here due to the resolution limitation of the figure sizes. The whole basin model contains hundreds of thousands of nodes. If all the "bullets" for all nodes are shown, the "bullets" will be pressed together. Figures 4.4A and 4.4C show fluid flow patterns for a central and marginal portion of the model domain, respectively. These expanded views show that basinal brines may penetrate into the basement rocks, and that basement fluids are also able to flow up into the basin by free convection. The typical fluid flow average linear velocity (3.5 m/yr) in the sandstone is several orders of magnitude higher than that in the basement (~2.5 × 10^-6 m/yr) mainly due to the large hydraulic conductivity difference. In this model, the moving fluids modify the temperature field (Fig. 4.4B). The central part of the sandstone unit is heated and has a relatively uniform temperature field. The portions close to the basin margins, in contrast, are cooled significantly. This may be attributed to the connection between the sandstone aquifer and the relatively cold water table (Fig. 4.2), i.e., if the sandstone sequence is relatively separated from the water table by the overlying low-permeability strata, the perturbation of the temperature field should be suppressed. The magnitude of the temperature perturbations is directly related to the velocity of fluid flow. In an individual
Fig. 4.5. Temperature field superimposed with the Darcy flux "bullets" of two sections for the same model described in Figure 4.4 at 0.1, 0.5, 1.0 and 5.0 m.y. modeling times. The insets show the location of expanded view in the conceptual model.
convection unit, the velocity decreases gradually towards the center of the cell. The horizontal dimensions of convection cells close to the side boundaries are larger than those in the middle because of the influence of increasing stratum slopes close to side margins. Note that, due to the image scale and resolution, the temperature changes caused by individual upwelling or downwelling centers cannot be seen in Figure 4.4B. Figure 4.5 provides a magnified view that shows both the fluid flow vectors and temperature contours. Figures 4.5A and 4.5B are temperature profiles of a section in the central part of the model domain at 0.1 and 0.5 m.y., respectively, and Figures 4.5C (4 m.y.) and 4.5D (5 m.y.) are for a section on the right-hand side of the model. The temperatures are elevated around the upwelling centers, and reduced close to the downwelling centers. Temperature perturbations become more significant towards the basin margin, relative to the center of the model. A sensitivity analysis of the effects of changing geothermal gradients (20-40 °C) and hydraulic heterogeneities (interbedded low-permeability unit in the sandstone sequence) was conducted by Cui et al. (2010). Changing these parameters modifies the nature of fluid flow in detail, such as fluid velocities, but not the general fluid flow patterns at the basin scale.

4.4.2 Solute transport

Figure 4.4 illustrates that fluids could flow across the basal unconformity during buoyancy-driven thermohaline convection. An important question that stems from this observation is: Does this fluid movement, with very low fluid flow velocities, contribute to uranium transport during uranium mineralization? In order to answer this question, uranium transport has been simulated, assuming that there is a basement uranium source
Fig. 4.6. Uranium concentration distribution at different modeling times when the uranium source is located in the center of the basement. The white numbers point out the localized uranium concentration (mg/l) along contours.
that is located in the center of the basin below the unconformity (Fig. 4.6A). The uranium source area is 30 km wide and 2 km thick. Leucogranites and pegmatoids from partial melting of the Paleoproterozoic metasediments, Paleoproterozoic high-K calc-alkaline granitoids and late Hudsonian vein type U-deposits in the basement of the Athabasca uranium province all represent favorable uranium sources (Cuney, 2005). The uranium source area in the model means a block enriched in U-bearing minerals. As chemical reaction is not considered in this study, it was assumed that the fluids that interact with these source rocks will have a constant uranium concentration of 500 mg/l in this model. This is based on recent uranium concentration measurement of the ore-forming brines in fluid inclusions (Richard et al., 2010; Richard et al., 2011a; Richard et al., 2011b). The model was run until 5 m.y.. The corresponding uranium migration for the model with a uranium source in the central basement is shown in a series of snapshots (Fig. 4.6) that represent four different times in the modeling duration. The uranium concentrations in the sandstone immediately above the uranium source reach about 11, 15 and 24 mg/l at 0.1, 1.0 and 5 m.y., respectively (Fig. 4.6B, 4.6C and 4.6D). After 5 m.y., uranium has spread across almost the entire basin (Fig. 4.6D). It should be noted that uranium minerals may precipitate at the unconformity where favorable physicochemical conditions exist, forming URU deposits. The background salinity distribution during the modeling is also shown in Figure 4.3. The basinal brines are diluted significantly after 5 m.y. modeling time. The dilution is caused by groundwater replenishment, as well as the neglection of fluid-rock reactions. Mass balance calculations were conducted to assess whether the modeled solute transport is able to supply enough uranium required to form giant URU deposits. To calculate this, it was assumed that the model is circular and
Fig. 4.7. Uranium concentration distribution at different modeling times when the uranium source is located in the basement close to the right basin margin. The white numbers point out the localized uranium concentration (mg/l) along contours.
symmetrical in the horizontal plane. Given that, the total volume of the sandstone above the uranium source is approximately 1800 km$^3$ ($30\times30\times2$). Given the effective porosity of the sandstone is 0.2, the volume of pore fluids is 360 km$^3$, and these fluids contain $5.4 \times 10^6$ and $8.6 \times 10^6$ t uranium at 1.0 and 5.0 m.y., respectively. The known uranium resource in the Athabasca basin is 587,063 t (Jefferson et al. (2007), therefore, the modeled amount of U transported out of the basement is more than enough to form the known deposits.

The geometric characteristics of the hydrogeological units and the location of the uranium source are both primary factors that affect solute transport. The uranium plume is almost symmetrical due to the symmetrical nature of the conceptual model when the uranium source straddles the central line (Fig. 4.6). The symmetrical characteristics of the plume were lost when the uranium source was placed towards the basin margin (Fig. 4.7), where the increased dips of the sedimentary strata promote higher fluid velocities that cause high fluid flux and uranium transport across the basal unconformity. When the uranium source is placed to the right side of the basement (Fig. 4.7A), a very different pattern of uranium dispersion is obtained (Fig. 4.7). It was expected that the uranium plume would spread primarily toward to the center of the basin. The updip portion of the clastic sequence, however, gains more uranium than the basin center. This shows that buoyancy derived from heat and solute concentration differences tends to cause basinal brines to migrate up along the sloping basal unconformity, even though fluid salinity increases with depth. Typical uranium concentrations approach 390 (Fig. 4.7B), 470 (Fig. 4.7C) and 490 (Fig. 4.7D) mg/l in the sandstone above the uranium source at 0.1, 1.0 and 5.0 m.y., respectively. These are significantly higher than the concentrations generated
Fig. 4.8. Uranium concentration distribution at different modeling times when the uranium source is located in the center of the sandstone unit. The white numbers point out the localized uranium concentration (mg/l) along contours.
Fig. 4.9. Uranium concentration distribution at different modeling times when the uranium source is located in the sandstone unit close to the right basin margin. The white numbers point out the localized uranium concentration (mg/l) along contours.
when the uranium source was centrally located (Fig. 4.6). This result suggests that the basement uranium source close to the basin margins is more favourable for the formation of giant deposits when other conditions are the same. This may explain why most of the known deposits are located around the basin margins (Fig. 4.1), although this may also reflect the greater ease of exploration along basin margins.

Figures 4.6 and 4.7 demonstrate that basement-derived uranium can be transported into the overlying basinal sediments by buoyancy-driven thermohaline convection. The next step was to relocate the uranium source into the sandstone and investigate if uranium can spread into the basement in this hydrologic regime. Figure 4.8 shows the results of the scenario with a uranium source, where the fluids has the same uranium concentration as previous models, in the sandstone center. After 0.1 m.y., uranium has been transported into the basement below the uranium source (Fig. 4.8B). As time advances, the uranium plume expands gradually within the basin and the upper portion of the basement. By 5.0 m.y., it has spread throughout the basin, and has penetrated into the basement about 2000 m below the uranium source, and 500 m closer to the basin margins (Fig. 4.8D). In this model, the uranium concentration in the basement fluids varies depending on the distance from the uranium source area, and is about 350 mg/l at 500 m below the basal unconformity. Figure 4.9 displays the results for the model with the uranium source in the sandstones on the right side of the basin. Again, the uranium plume expands in all directions as the system evolves. At 5 m.y., the plume has not spread across the whole basin, but penetrates to a greater depth (about 2500 m) than when the source is centrally located (Fig. 4.8).
4.4.3 The effect of increasing hydraulic conductivity due to fractures

Based on laboratory measurements and aquifer tests, the hydraulic conductivity of crystalline basement rocks varies widely, over about 5 log-units, from $10^{-15}$ to $10^{-10}$ m/s, and increases to $10^{-11}$ to $10^{-4}$ m/s when the influence of fractures is taken into account (Davis, 1969; Freeze and Cherry, 1979; Garven and Freeze, 1984b; Stober and Bucher, 2007). In addition, it is worth noting that most of these data were obtained for basement rocks that are less than 5 km deep. The hydraulic conductivity ($3.0 \times 10^{-11}$ m/s) of basement rocks used in the model (Table 4.1) is in the upper range of values for unfractured crystalline rocks and in the lower range of values for fractured crystalline rocks. It is impossible to obtain the real hydraulic conductivity for the basement rocks during uranium mineralization. Faults and fracture networks, however, are known to exist widely in such Archean to Proterozoic basements (Card et al., 2007; Mercadier et al., 2010), and all the deposits are spatially associated with faults (Jefferson et al., 2007). These fractures could have increased the hydraulic conductivity of the basement significantly. Some empirical equations have been proposed to relate the permeability of continental crystalline basement rocks to burial depth over large temporal and spatial scales (Manning and Ingebritsen, 1999; Shmonov et al., 2003; Stober and Bucher, 2007). All these equations have the same form of:

$$\log(k) = -a \log(z) - b,$$  \hspace{1cm} (4.8)

where $k$ is the permeability in m$^2$, $z$ is the depth in km, and $a$ and $b$ are constant coefficients that vary slightly in different models depending on the database behind the equation. These equations are derived from measured laboratory and field data, as well as numerical modeling, and take account of the influence of fractures. In the recent model of
Stober and Bucher (2007), $a$ and $b$ are 1.38 and 16.4, respectively. The hydraulic conductivity is related to permeability by

$$K = \frac{k \rho f g}{\mu_f},$$  \hspace{1cm} (4.9)

where $K$ is the hydraulic conductivity, $g$ is the acceleration due to gravity, and $\mu_f$ is the dynamic fluid viscosity that evolves depending on temperature and pressure. In addition, based on the measured data in the literature and equations (4.8) and (4.9), a plot has been compiled by Stober and Bucher (2007) to show the hydraulic conductivity profile of crystalline basement with depth (Fig. 4.10). According to this plot, the crystalline

Fig. 4.10. Hydraulic conductivity of crystalline basement rocks versus depth (redrawn from Stober and Bucher, 2007). The dashed curve was compiled by Stober and Bucher (2007), and the solid curve was originally from Ingebritsen and Manning (1999). Data from well tests used by Stober and Bucher (2007) are also presented by rectangles for different rocks.
Fig. 4.11. Fluid flow patterns (A) and uranium concentration distribution (B, C and D) at different modeling times, when a hydraulic conductivity of $10^{-9}$ m/s is assigned to the uppermost 2 km of the basement. The white numbers point out the localized uranium concentration (mg/l) along contours.
basement rocks have a hydraulic conductivity of $10^{-9}$ m/s at 6 km depth. The next step in the modeling was to assign this higher hydraulic conductivity ($10^{-9}$ m/s) to the uppermost 2 km of the basement. In addition, the possible anisotropy in the hydraulic conductivity has to be considered, which will depend on the orientation of fractures. Inasmuch as most faults are steeply inclined in these basins, an anisotropy ratio of $5 (K_z/K_x)$ was assumed. Figure 4.11 shows the results of the model run with these modified conductivities applied to the basement. In general, the overall fluid flow pattern across the basin has not changed as a result of these new conductivity values. Fluid flow linear velocities in the relatively permeable portion of the basement, however, are significantly elevated, and lie between $1 \times 10^{-3}$ to $5 \times 10^{-3}$ m/yr. As a result, the uranium plume has expanded, and the uranium concentrations are higher compared to the model represented by Figure 4.6. A spectrum of other simulations using basement hydraulic conductivities of greater than $10^{-9}$ m/s, have also been carried out. As expected, increasing the hydraulic conductivity will increasingly facilitate the fluid interaction between the basement and sedimentary cover.

4.4.4 Fluid flow across the basement/cover interface

A number of simulations have been conducted to investigate the required conditions for the presence of convection cells straddling the basal unconformity. One may intuitively expect that convection will not occur in an extremely low-permeability basement, an expectation that has been borne out by modelling. When the hydraulic conductivity of the basement was reduced to $3 \times 10^{-15}$ m/s, which is in the lower range for unfractured crystalline rocks, fluid flow across the basal unconformity did not occur, even after 5 m.y. (Fig. 4.12A). If the hydraulic conductivity of the basement is increased in increments of one order of magnitude, fluid flow across the unconformity occurs once
$K$ reaches $3 \times 10^{-13}$ m/s, although fluid flow velocities are only $2 \times 10^{-7}$ m/yr (Fig. 4.12B). This implies that there is a critical value between $10^{-14}$ and $10^{-13}$ m/s for the onset of the flow across the unconformity, if the other parameters remain the same. Given that the typical hydraulic conductivity of fractured crystalline rocks ranges from $10^{-11}$ to $10^{-4}$ m/s (Davis, 1969; Freeze and Cherry, 1979; Garven and Freeze, 1984b; Stober and Bucher, 2007), such trans-unconformity flow is likely to have occurred in typical URU deposit setting.

Fig. 4.12. Fluid flow patterns of the models with relatively low basement hydraulic conductivities of $3 \times 10^{-15}$ m/s (A) and $3 \times 10^{-13}$ m/s (B), respectively. Streamlines are indicated by "bullets", and fluid flow velocities are indicated by the color scale. The thicker ends of "bullets" point to fluid flow directions; blue arrows are drawn to help trace fluid flow. The insets show the location of expanded views in the conceptual model.
Fig. 4.13. Fluid flow (A), temperature field (B) and uranium transport (C) of the model with a sandstone $K = 1 \times 10^{-8}$ m/s at 1 m.y. modeling time. Fluid flow directions are indicated by "bullets", and fluid flow velocities are indicated by the color scale. The thicker ends of "bullets" point to fluid flow directions.

The influence of the hydraulic conductivity of the sandstone sequence has also been investigated. The value of $K = 3.0 \times 10^{-6}$ m/s used in the modeling described above
represents a relatively high hydraulic conductivity for sandstones (Freeze and Cherry, 1979; Garven and Freeze, 1984b). The $K = 3.0 \times 10^{-11}$ m/s was assigned to the basement in the following models. When the hydraulic conductivity of sandstone decreased to $10^{-8}$ m/s, the free convection in the sandstone ceased (Fig. 4.13A). Although there are some random perturbations of fluid flow in the basement, they have no noticeable effects on heat and solute transport (Fig. 4.13).

### 4.5 Discussion

#### 4.5.1 Relative contributions of advection, ion diffusion and mechanical dispersion on solute transport

The transport of solutes in porous media can occur through advection, ionic (molecular) diffusion, and mechanical dispersion. Ion diffusion results from the existence of chemical potential or concentration gradients. Advection is solute migration along with the flowing groundwater. In the current study, due to the relatively low fluid-flow velocities in the basement, one may wonder which mechanism plays a dominant role in the transport of uranium. In order to isolate the effect of the various mechanisms, some parameters have been modified from the basic model. Figure 4.14A depicts results for a model in which the influence of thermal-convection and a non-uniform NaCl concentration have been removed. This was achieved by using a uniform temperature, thermal conductivity and NaCl concentration; the longitudinal and transverse dispersivities were also set to zero. The heterogeneity of initial uranium concentrations still leads to a density difference, but this may be ignored due to the relatively low uranium concentration. Thus, Figure 4.14A approximately shows the result of ion diffusion at 1 m.y.. Figure 4.14B is the uranium distribution when the coefficient of ionic
diffusion was set to zero. Figure 4.6C is the corresponding model that coupled advection, ion diffusion and mechanical dispersion. Mechanical dispersion is a phenomenon that always accompanies groundwater advection (Ingebritsen et al., 2006), thus, there is no way to isolate its influence from advection and ion diffusion in the modeling. Based on these results, it can be concluded that the uranium transport is dominantly caused by groundwater advection and mechanical dispersion, rather than ion diffusion.

Fig. 4.14. Uranium concentration at 1 m.y. modeling time. (A) The influence of thermal-convection and heterogeneous NaCl concentration has been removed; longitudinal and transverse dispersivities are zero. (B) The result when the diffusion coefficient is set to zero. The white numbers point out the localized uranium concentration (mg/l) along contours.

### 4.5.2 Uranium initial and boundary conditions

The uranium source in this study is represented by fluids with a uranium concentration of 500 mg/l, a value that does not change during the modeling. In the model, as new fluids pass through the source area, they attain a uranium concentration
Fig. 4.15. Uranium concentration distribution at different modeling times when the uranium source is located in the center of the basement. The white numbers point out the localized uranium concentration (mg/l) along contours. Uranium concentrations of fluids in the source area are free to change in this model.
of 500 mg/l. In reality, uranium may be gradually depleted during fluid flow and fluid-rock interaction. How the chemistry of the source fluids will evolve over time will be controlled by the identity of the U-bearing minerals, fluid chemistry, pressure, temperature, and fluid flux (i.e., mineral stability and alteration and the nature of fluid-rock interaction). Given the short modeling time of 5 m.y., compared to the possible mineralization time span of at least 100 m.y. (Kyser et al., 2000), the initial assumption of a constant U concentration in the source fluids is reasonable for the preliminary modeling. Nonetheless, a model in which the uranium concentration of fluids in the basement source area is free to gradually decrease as groundwater replenishes the source area, also has been run. As expected, less uranium is transported into the overlying sedimentary sequence from the basement (Fig. 4.15) compared to the same model with a constant uranium concentration in the source area (Fig. 4.6). The uranium concentration in the sandstone immediately above the uranium source decreased to 7.4 mg/l at 5 m.y. from 24 mg/l in the model depicted by Figure 4.6. The same approach in section 4.2 was used to conduct mass balance calculation, and the sandstone above the uranium source hosts $2.7 \times 10^6$ t uranium after 5 m.y. modeling time, which is still sufficient to form giant uranium deposits given favorable conditions. The significant drop from 24 mg/l to 7.4 mg/l, however, indicates the primary control role of uranium content of source rocks in the formation of giant deposits.

4.5.3 Implications for URU deposit genetic models

A disagreement exists regarding the source of uranium in URU deposits. The most commonly accepted diagenetic model involves interaction between oxidizing basinal brines and basement-derived reducing fluids and/or minerals around the basal
unconformity. Uranium has to be leached by oxidizing fluids from accessory minerals and/or previously formed uranium oxides either in the basin (e.g., Fayek and Kyser, 1997; Kyser et al., 2000) or in the basement (e.g., Annesley and Madore, 1999; Hecht and Cuney, 2000; Cuney, 2005; Mercadier et al., 2010; Richard et al., 2010). The thick and highly permeable sandstone formations in these Proterozoic basins are conducive to pervasive thermally driven convection (Raffensperger and Garven, 1995b; Cui et al., 2010). In addition, the lack of organic matter or Fe$^{2+}$ in these formations maintains the high oxidation state of the basinal brines. It has been proposed that the free convection of oxidizing basinal brines was responsible for uranium uptake from accessory minerals, such as zircon and monazite (Raffensperger and Garven, 1995b; Kyser and Cuney, 2009). On the other hand, Archean granite, Paleoproterozoic sediments, leucogranites, pegmatoids, and calc-alkaline granitoids in the basement all represent possible and favorable uranium sources due to their high uranium content (Cuney, 2005). For the latter model, however, how the oxidizing basinal brines may have penetrated into the low-permeability basement and extracted uranium is a question that has not been satisfactorily answered. Jefferson et al. (2007) stated that the relatively small volume of alteration in the basement constrains any model in which the uranium originated from the basement. Oliver et al. (2006) argued that free convection cells are able to penetrate basement only where the permeability difference between the sedimentary cover and the metamorphic basement is less than two orders of magnitude. Cui et al. (2012) showed, using coupled thermo-hydro-mechanical modeling, that fluid interaction may occur along major faults during tectonic deformation. The study by Mercadier et al. (2010), however, indicated that basinal brines may flow hundreds of meters into the basement below the
unconformity through a dense network of pervasive microfractures. Richard et al. (2010) believed that the NaCl-rich and CaCl₂-rich brines found in fluid inclusions in basement rocks all originated from the basinal sequences, and that the CaCl₂-rich brines are possibly the product of interaction between the NaCl-rich brine and the basement. The results presented here show that the pervasive interaction of basinal brines with basement rocks can occur as a result of buoyancy-driven thermohaline convection around the basal unconformity. Although fluid-rock interaction in the basement happens very slowly, first-order mass balance calculations show that such fluids are capable of transporting sufficient uranium from a source region to the unconformity to form giant URU deposits. The results cannot refute either of the models, i.e. whether the uranium source is in the basement rocks or the sedimentary cover, but they do demonstrate that it is hydrodynamically possible to move uranium from the basement and deposit it around the unconformity, assuming depositional requirements are satisfied, even though the basement has a low permeability.

Based on the new results in this paper and in previous studies (e.g., Raffensperger and Garven, 1995b; Kyser and Cuney, 2009; Cui et al., 2010, 2012), a tentative genetic model is proposed in which both thermohaline convection and tectonic deformation contribute to form giant URU deposits. Free thermohaline convection develops during periods of tectonic quiescence in Proterozoic basins. Oxidized basinal brines are able to flow downward into the basement at very low velocities through this type of convection even though the basement has an extremely low permeability. Uranium leaching may occur along the migration of the oxidized brines. When significant tectonic events occur, high-permeability fluid flow conduits develop and free convection is suppressed. Fluid
flow then will be controlled dominantly by tectonic deformation. Uraniferous fluids will flow mainly along fractured zones. When favorable conditions (e.g., redox interfaces around the basal unconformity) are met, giant uranium deposits will develop.

4.6 Conclusions

It has been widely recognized that basinal brines play an essential role in the formation of stratabound ore deposits, and even for deposits in basement rocks that were at one time overlain by sediments. With the help of new "bullets" plots of fluid flow in FEFLOW, fluid flow velocities that span several orders of magnitude can be observed and analyzed in the same graph. The modeling shows that buoyancy-driven thermohaline convection is able to pervasively penetrate into the low-permeability basement for up to 1-2 km below the unconformity, albeit at very low velocities, where typical hydrological parameters for these Proterozoic hydrogeological units are used. Interaction among basinal brines, basement fluids, clastic sedimentary rocks and basement rocks can occur through thermohaline convection. Hydraulic conductivities, heterogeneity of hydraulic conductivity, and the geometric characteristics of strata can significantly affect such interactions. Dissolved chemical mass can be transported and exchanged between basin and basement during this interaction. Based on mass balance calculations, given a geologic time of 1 to 5 m.y., it is possible for basinal brines to penetrate the basement, and presuming they can leach and transport sufficient uranium to form giant unconformity-related uranium deposits. The results presented above regarding possible fluid-fluid and fluid-rock interactions among relative low-permeability and high-permeability rocks may also prove useful for genetic models for other deposit types, such as the Irish-type and SEDEX Pb-Zn deposits.
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Chapter 5

Summary of original contribution and recommendations for future work

5.1 Summary of original contribution

5.1.1 Driving mechanisms of fluid flow

Driving mechanisms for subsurface fluid flow mainly include sediment compaction, topographic relief, tectonic deformation, and buoyancy due to variations in temperature and salinity. Compaction-driven fluid flow typically has a much lower velocity than the fluid flow resulting from other mechanisms (Bethke, 1985; Bjørlykke, 1993). In addition, URU deposits are believed to form after the main sediment compaction (Jefferson et al., 2007). Thus, the role of compaction-driven flow in the formation of URU deposits is not investigated in this study. Topography-driven fluid flow usually leads to a gradual evolution of fluid properties, such as temperature, pressure and composition, along flow directions. For example, temperatures are reduced at the recharge zone, and elevated in the discharge area as in Figure 2.6. As the thermal regime is an important factor for the formation of deposits, if topography-driven fluid flow dominates, deposits should be asymmetrically distributed, or at least have significantly evolved characteristics along fluid flow directions. URU deposits, however, have been found on all sides close to basinal margins, which is not consistent with the distribution expected from topography-driven fluid flow. Therefore, gravity from
topographic relieves must not be the main driving force for the formation of URU deposits.

In the environments where URU deposits are found, a thick and highly permeable basal sandstone unit overlies the basal unconformity. Due to the existence of this unit, thermal convection has been believed to have been the most likely fluid flow driving mechanism, as first suggested by Hoeve and Sibbald (1978). Tectonic deformation has been considered to be a fundamental driving force for the migration of mineralizing fluids in various geological settings (e.g., Cox et al., 2001; Sibson, 2001). Therefore, this study has focused on thermohaline convection and deformation-driven fluid flow. The best way to investigate their relative roles on controlling fluid flow is to simulate these processes together in one single model. There is, however, no such code available that is powerful enough to deal with heat transport, solute transport, fluid flow and mechanical deformation together for a basin-scale model that contains up to hundreds of thousands nodes. Thus, the code FLAC was selected in Chapter 3 to address fluid flow driven by tectonic deformation; FEFLOW was used to study the influence of thermohaline convection in Chapter 2 and Chapter 4.

The modeling presented in Chapter 2 and Chapter 4 confirmed the statement of Raffensperger and Garven (1995) about thermally-driven convection in Paleoproterozoic basins. They examined the possibility of thermally-driven convection as the driving mechanism of ore-forming fluid flow for the URU deposits by numerical modeling, and concluded that thermally-driven free convection of basinal brines was the dominant mechanism producing regional groundwater flow within thick sandstone aquifers in the Proterozoic basins of Australia and Canada. Hiatt et al. (2003), however, argued that the
Thelon Formation in the Thelon basin had been partitioned into a series of aquifers and aquitards before uranium mineralization, and this partitioning may have suppressed the development of thermally-driven free convection. Thus, several models were designed to examine the influence of such heterogeneity. The results indicate that convection cells still develop, although they are partially confined by the introduction of aquitards (Fig. 2.5). According to the free convection theory of groundwater (Nield and Bejan, 2006), the geothermal gradient is another important factor affecting free convection in porous media. A gradient range of 20-35 °C/km has been used to investigate its influence on free convection in these Proterozoic basins; the results show that free convection occurs for any gradient in this range, although fluid velocities change depending on temperature gradients. Salinity also plays a role in affecting free convection. The salinity of basinal brines generally increases with depth at various rates depending on the geological setting (Kharaka and Hanor, 2003), and may counteract the effect of increasing temperature on brine density. The salinity profile, however, did not have a significant influence on free convection according to this study.

The modeling in Chapter 3 demonstrates the potential importance of tectonic deformation on mineralizing fluid flow. The accumulation and dissipation rates of pore pressure in various geological units are not uniform due to their different hydrological properties, such as permeability, porosity, bulk modulus and shear modulus (Fig. 3.10). As a result of these differences, hydraulic head gradients will develop between adjacent units, thus resulting in fluid interaction. This type of interaction, however, exclusively occurs in main faults. Furthermore, the occurrence of tectonic events, such as orogeny and crustal stretching, restrains free convection depending on their strain rates. On the
basis of these results, it can be concluded that buoyancy (mainly from temperature gradients) and pressure differences from mechanical deformation are the most profound factors controlling fluid flow for the formation of URU deposits.

5.1.2 Uranium sources

"Where does the massive amount of uranium come from?" This is an unresolved question that has been a subject of intensive debate. Although the diagenetic model has been widely accepted, two end-member models, which invoke different uranium sources, remain. One considers the sedimentary fill as the uranium source, while for the other the uranium comes from the basement rocks. The current average uranium content of the sandstone units, away from mineralized areas, is below 1 ppm (Cuney, 2005). Their original uranium content, however, would have been similarly high with that of the current basement below these basins, as these siliciclastic sediments mainly originated from the surrounding basement (Palmer et al., 2004). Thus, the low preserved uranium concentrations in the siliciclastic rocks may suggest that these sedimentary rocks have been leached off their uranium either during the transport process at surface or after their sedimentation. It is possible that the leached uranium is now concentrated in the URU deposits (Jefferson et al., 2007), although how and when uranium was removed from the elastic rocks is still not resolved. Monazite is a common accessory mineral that can contain significant amounts of uranium and thorium (Hecht and Cuney, 2000). Detrital monazite alteration has been proposed as a possible mechanism to release uranium to the oxidizing brines due to the anomalously high background thorium and REEs content currently in the Athabasca Group (Madore et al., 2000; Mathieu et al., 2000; Mwenifumbo and Bernius, 2007). The great differences between uranium and thorium
content within the clastic rocks have been attributed to selective geochemical extraction of uranium (Jefferson et al., 2007). In contrast to the low uranium content in the sedimentary fill, the basement contains a number of uranium-rich minerals, such as Archean granite, leucogranites, pegmatoids, and calc-alkaline granitoids (Cuney, 2005). However, the question of how and why the oxidized basinal brines may have penetrated the low-permeability basement and extracted uranium has not been satisfactorily answered. Jefferson et al. (2007) argued that the relatively low permeability in the basement constrains any model in which the uranium originated from the basement. On the other hand, the study by Mercadier et al. (2010) showed that basinal brines may pervasively flow into the basement to hundreds of meters below the unconformity.

Richard et al. (2010) also stated that a Na-Ca-Mg-K-Sr-Ba brine and a Ca-Mg-Na-K-Sr-Ba brine had mixed within the main URU deposit area. Both brines are derived from evaporated seawater, and they believed that the Ca-rich fluids resulted from the interaction between basin-originated Na-rich brines and basement rocks.

This study provides some new insights into this debate from the point of view of hydrodynamics. Under the conditions of low to moderate temperature, only the highly oxidizing fluids are able to leach and transport uranium significantly in rocks (Skirrow et al., 2009). Thus, if there are not sufficient oxidizing basinal brines that flow into the basement, it is impossible to leach so much uranium required for the formation of URU deposits from the basement, i.e., the sedimentary rocks are the only possible uranium source. On the other hand, if the oxidizing basinal brines are able to flow through both the basin fill and the basement, the uranium source could be either of them or both of them from the point of view of fluid flow. This study along with previous modeling
(Raffensperger and Garven, 1995) all demonstrate that free convection occurs thorough the sandstone sequence. This convection provides a viable mechanism to leach uranium from basin fill. The modeling presented in Chapter 4 suggests that uranium-bearing basinal brines could have interacted widely with fluids or reduced lithologies in the basement close to the basal unconformity through thermohaline convection. Although this interaction occurs very slowly, mass balance calculations indicate that it is capable of leaching enough uranium to form giant URU deposits. Thus, the results still cannot differentiate between the basement rocks and the sedimentary cover as the uranium source. These results nevertheless show that it is possible to leach uranium from the basement and deposit it at the intersection of unconformities and faults when other required depositional conditions are satisfied based on the point of view of hydrodynamics, even though the basement is relatively impermeable.

5.1.3 The role of basal unconformities and faults

A basal unconformity is the most favorable exploration target for URU deposits, although the role of the unconformity in the formation of these deposits remains uncertain. The thickness of the unconformity zone can range from a few centimeters up to 220 m (Jefferson et al., 2007). In Chapter 2, in order to assess the effect of the basal unconformity, an unconformity zone with a thickness of 250 m was added into the conceptual model below the Thelon Formation. Three horizontal hydraulic conductivities (200 m/yr, 400 m/yr and 1000 m/yr) were used to examine the effect of different permeabilities in this unconformity zone. Variation of this parameter has almost no effect on the temperature field in the vertical direction. In the horizontal direction, the isotherms have a tendency to become flatter with increasing hydraulic conductivity in the
unconformity zone. The convection cells still develop in the sandstone sequence, in a similar manner to the models without the consideration of a distinct unconformity unit (Fig. 2.3). The horizontal dimensions of the cells grow, however, when the hydraulic conductivity of the unconformity was increased (Fig. 2.4). In the models described in Chapter 4, although there is no separated basal unconformity between the basinal fill and the basement, the interface between the basin fill and basement still represents the highest velocity zone (Fig. 4.3). Additionally, as would be expected from free convection theory applied to strongly heterogeneous porous media, the unconformity zone represents a favored place for localized convections due to the sharp change of hydrological properties (Simmons et al., 2010). All these findings make the basal unconformity the most favorable area for mineralization area from the point of view of fluid focusing and flux.

Most URU deposits are located at the intersection of reactivated faults with offsets of the unconformity. The close association of URU deposits with faults indicates that they have played a significant role in the formation of URU deposits. They show variable effects on fluid flow depending on whether or not the influence of deformation is considered. In Chapter 2, faults are represented by high-permeability zones without considering the pore pressure changes resulting from mechanical deformation. The two most important factors controlling the influence of faults is 1) whether or not different faults are connected by aquifers and 2) whether or not the faults reach the surface. When several isolated blind faults were added into the conceptual model, no significant changes were observed in the temperature field. Free convection retains a similar pattern at the basin scale. When these faults are allowed to penetrate the overlying cover to the surface,
convection cells connecting different faults develop (Fig. 2.7). The temperature is significantly reduced in the recharge zone. In Chapter 3, the Mohr-Coulomb model was used to describe elastic-plastic deformation of porous rocks. According to this model, rocks under loading deform elastically up to a yield point, after which they deform in a plastic manner. Through this model, the interaction between fluid flow and mechanical deformation was modeled. During compressive deformation, brines in the basement flow out along faults into the overlying sedimentary fill. In contrast, during extension deformation, basinal brines migrate into faulted structures to interact with the basement. Additionally, reactivation of preexisting basement structures and the generation of new faults suppressed buoyancy-driven free convection, and led to deformation-dominated fluid flow or mixed convection, depending on strain rates.

5.2 Suggestion for future work

Although this study draws some important conclusions regarding the formation of URU deposits based on a number of 2D numerical experiments that integrated most published data relevant to the URU deposits, some aspects are still worthy of further investigation in more details. This study will serve a starting point to additional research on the following topics.

1. More accurate geological data: Lack of data is always a serious issue for geological modeling. Some important parameters in this thesis, such as permeabilities, thermal conductivities and porosities, were determined based on published data compilations. Although use of these data is reasonable for this preliminary study, directly measured data for the study area would improve the modeling and provide more reliable results.
2. Reactive transport modeling: Although solute transport has been coupled into this study, fluid-fluid and fluid-rock reactions were not considered. This is mainly due to the limitation of the codes and the computing speed of the available computers. Reactive transport modeling may demonstrate more realistic solute transport patterns.

3. Three-dimensional modeling: 2D modeling is still the most popular way to do subsurface fluid flow modeling because it provides more reliable information than 1D modeling and requires much less data for the model construction than 3D modeling. 3D modeling, however, would provide more realistic results. If the subsidence history of the Athabasca or Thelon or Kombolgie basin can be reconstructed, a 3D conceptual model can be built.

4. Fully coupled modeling: The present work simulated a number of processes using two different codes. Further effort is required to develop a code that is able to efficiently deal with fluid flow, heat transport, solute transport, chemical reactions and mechanical deformation together.
References


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