Archean gravity-driven tectons on hot and flooded continents: Controls on long-lived mineralised hydrothermal systems away from continental margins

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**A B S T R A C T**

We present the results of two-dimensional numerical modelling experiments on the thermal evolution of Archean greenstones as they sink into a less dense, hot and weak felsic crust. We compare this thermal evolution to that obtained via the analysis of isotopic data and fluid inclusion microthermometry data obtained in the Paleoarchean to Mesoarchean Warrawoona Synform (Eastern Pilbara Craton, Western Australia). Our numerical experiments reveal a two-stage evolution. In the first stage, cooling affects zones of downwelling as greenstone belts are advected downward, whereas adjacent domes become warmer as deep and hot material is advected upward. We show that this is consistent with stable isotopes data from the Warrawoona Synform, which reveal an early episode of seafloor-like alteration (90–160°C) strongly focused along steeply dipping shear zones. In a second long-lived stage, lateral heat exchanges between domes and basins dominate the system as domes cool down while downwelling zones become increasingly warmer. In the Warrawoona greenstone belt, stable isotopes in gold-bearing quartz veins post-dating the sagduction-related vertical fabrics reveal that rock–fluid interaction occurred at much higher temperatures (234–372°C) than seafloor-like alteration. We propose that emplacement of thick and dense continental flood basalts, on flooded hot and weak continental plates, led to conditions particularly favourable to hydrothermal processes and the formation of mineral deposits. We further argue that sagduction was able to drive crustal-scale deformation in the interior of continents, away from plate margins. On largely flooded continents, sagduction-related shear zones acted as fluid pathways promoting gold mineralisation far away from active plate boundaries, continental rift zones or collisional mountain belts.

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1. Introduction

Several features made the Paleo- to Mesoarchean landscape (3.6–2.8 Gy old) fundamentally different than it is today, and particularly favourable to the formation and preservation of gold deposits. First, most Archean cratons are buried under up to 15 km thick blanket of autochthonous Continental Flood Basalts (CFBs, e.g. Maurice et al., 2009), which include high-magnesium basalts and komatiites that have only few equivalents on modern Earth. Second, despite their thicknesses most of these thick volcanic piles were emplaced below, and remained below, sea level for most of their Archean history (Arndt, 1999; Flament et al., 2008, 2011; Kump and Barley, 2007). This characteristic is of great importance as it implies that an infinite fluid reservoir was available to feed hydrothermal circulations in the CFBs. Third, radiogenic heat production in the continental crust was much higher than it is today which, in combination to the thermal insulation effect of CFBs, provided the heat engine, in the form of a strong thermal gradient, to power hydrothermal cells in the CFBs. Fourth, Paleoarchean to Mesoarchean cratons throughout the world are characterised by ovoid granitic domes, 40–100 km in diameter, encircled by greenstone belts. Greenstone belts form strongly foliated vertical sheets connected through vertical triple junctions where constrictional fabrics dominate (Bouhallier et al., 1995; Chardon et al., 1996; McGregor, 1951). This dome and basin pattern, which characterises many Archean cratons, is commonly interpreted in terms of gravitational sinking (sagduction) of dense greenstone belts into hot, and therefore weak, felsic crust (Chardon et al., 1996, 1998; Collins et al., 1998; Dixon and Summers, 1983; Mareschal and West, 1980; McGregor, 1951). This process is often wrongly described as “vertical tectonics”. Indeed, during sagduction horizontal and vertical displacements are perfectly coupled. Domes can rise and greenstone keels can sink because horizontal displacements provide the necessary space for vertical mass transfer (e.g. Mareschal and West, 1980). This process is also often wrongly opposed to plate boundary driven deformation. Indeed there is
2. The East Pilbara Craton: an example of sagduction-related “simple structural complexity”

The East Pilbara Granite Greenstone Terrane (Fig. 1a) is one of the best documented examples of dome–and-basin pattern interpreted in terms of gravitational instability (Collins, 1989; Delor et al., 1991; Hickman, 1983; Teysseyer et al., 1990; Thébaud, 2006; Van Kranendonk et al., 2004a), although alternative views exist (Blewett, 2002; Kloppenburg et al., 2001). The East Pilbara Granite Greenstone Terrane provides a Paleaoarchean to Mesaoarchean geologic backdrop to study tectono–thermal processes and fluid–rock interactions in a sagduction setting. It preserves a geologic history often largely overprinted in many Neoarchean cratons. The bulk of the domes consist of 3324–3300 Ma old syn- to post-kinematic suites of high-K granitic suites that are derived from, and intrusive in, older 3460–3430 Ma Tonalite–Trondjhemite–Granodiorite (TTG) gneisses and greenstones of the Warrawoona Group (Hickman, 1983; Hickman and Van Kranendonk, 2004; Smithies et al., 2003; Van Kranendonk et al., 2002, 2007). The domes are themselves intruded by younger, mainly 3300–3240 Ma old, granites. The older TTG gneisses formed the basement of greenstones, the emplacement of which is associated with two major volcanic cycles starting with the deposition of Warrawoona Group at ca. 3490 Ma and followed by the deposition of the Kelly Group at ca. 3335 Ma (Fig. 2) (Van Kranendonk et al., 2007). The thicknesses of the greenstone covers show significant lateral variation from 8 to 12 km for the Warrawoona Group, and 4–9 km for the Kelly Group (e.g. Hickman, 1983; Van Kranendonk et al., 2007). Considerations on both the present crustal thickness of the Pilbara (35–37 km) and the average erosion level (ca. 7 km) suggest that the greenstones accumulated on top of a 30–35 km thick continental crust. Ubiquitous pillow-lava, hydrothermal cherts, VMS mineralisation, epidote-chlorite-Ca–Na-plagioclase-Ca-amphibole secondary mineral assemblages and intense silification throughout the greenstone pile imply subaqueous emplacement (Barley, 1984; Barley and Pickard, 1999; Buick and Barnes, 1984; DiMarco and Lowe, 1989; Van Kranendonk, 2006).

The structural complexity of greenstone covers is a function of their position with respect to the domes. On the NE flank of the Mount Edgar dome, the Marble Bar greenstone belt lies directly on top of the Mount Edgar granitic complex and is structurally part of the dome (Fig. 1b). In the Marble Bar greenstone belt, the stratigraphy of the Warrawoona and Kelly Groups is relatively well preserved and structurally simple with a monotonous dip of 20–60° to the NE. In contrast, in the Warrawoona synform (white star in Fig. 1b), greenstones are steeply dipping and strongly deformed against the near-vertical southwest margin of the Mount Edgar dome (Fig. 3). At this location, the greenstone cover belongs to a basin pinched between the Mount Edgar dome to the north and the Corunna Down dome to the south. Compared to that of the Marble Bar greenstone belt, the structure in the Warrawoona synform is more complex showing multiple phases of folding, a prominent horizontal to vertical stretching lineation, and numerous shear zones and quartz vein arrays (e.g. Thébaud et al., 2008). Contrasting, yet predictable structural complexity, with complex structures above downwelling regions (e.g. Warrawoona syncline) and simpler structures above rising domes (e.g. Marble Bar belt) is a key attribute of sagduction settings (e.g. Bouhallier et al., 1995; Thébaud, 2006).

3. Numerical experiment setup

In order to interpret the thermal history derived from isotopic studies, and to understand fluid flow in a sagduction setting, we have performed a series of numerical experiments to document the thermal and mechanical history of sagducted greenstones. The process of sagduction has been tested through numerical and physical experiments (de Bremond d’Ars et al., 1999; Dixon and Summers, 1983; Mareschal and West, 1980; Robin and Bailey, 2009; West and Mareschal, 1979). This process is driven by the need for minimisation of internal gravitational potential energy of a system involving a density inversion (denser layer above a layer of lower density). It is resisted by the viscosity of the system, in particular by that of the stronger layer involved in the sagduction. Hence, the timing of sagduction is inversely proportional to the density contrast and proportional to the viscosity of the stronger layer. The emplacement of greenstones increases the geothermal gradient (e.g. Rey et al., 2003; Sandiford et al., 2004; West and Mareschal, 1979), which in turn reduces the viscosity and accelerates sagduction. To model this process, we use Ellipsis, a Lagrangian integration point finite element code capable of tracking time dependent variables in combination with an Eulerian mesh. This coupled Lagrangian/Eulerian approach allows for the accurate tracking of density interfaces during large deformation (Moresi et al., 2001, 2002). We use viscoplastic rheologies mimicking standard rheological profiles for the continental lithosphere (e.g. Brace and Kohlstedt, 1980). Our experiments include realistic geotherms with self-radiogenic heating and partial melting with feedback on viscosities and densities.

In our numerical experiments, the greenstone cover is 15 km thick in average, consistent with the thickness of many Archean greenstone covers, and in particular the average cumulative thickness of the Warrawoona and Kelly Groups in the EPGG (e.g. Van Kranendonk et al., 2007) to which we will confront our results.
Fig. 1. (a) Simplified Geological Map of the North Pilbara Terrain modified after Van Kranendonk et al. (2002). EPGGT, East Pilbara Granite-Greenstone Terrane; CPTZ, Central Pilbara Tectonic Zone; WPGGT, West Pilbara Granite-Greenstone Terrane; MB, Marble bar greenstone belt. Legend: (1) Greenstones, (2) Granitoid complexes, (3) Sedimentary basins, (4) Hamersley basin, (5) Phanerozoic cover. (b1) Simplified structural sketch of Mount Edgar and Corunna Down granitic domes complexes and surrounding regions. (b2) The first derivative of the topography shows that domes (thick dashed lines) are characterised by gentle topography and include both granitic rocks (crossed region on the right panel) and greenstone cover (grey region). In contrast a rough topography characterises the basins surrounding the domes. The mining districts of Klondike in the Warrawoona Synform (white star), and the mining district of Bamboo Creek (white hexagon) are both located above downwelling regions.

Small thickness variations are introduced to allow for the initiation of sagduction. Gravity modelling in the East Pilbara Granite Greenstone Terrane (Blewett et al., 2004) constrains the bulk density of our CFBs (2840 kg m\(^{-3}\)). These CFBs are emplaced in three successive 2.5 km thick layers at \(t_0\), \(t_0 + 30\) myr (lower Warrawoona Group), \(t_0 + 60\) myr (upper Warrawoona Group), plus a 7.5 km thick layer (the Kelly Group) emplaced at \(t_0 + 140\) myr. Assuming \(t_0 = 3490\) Ma, this emplacement history approximates that of the Warrawoona and Kelly Groups (e.g. Bagas et al., 2002; Van Kranendonk et al., 2007). Our model CFBs are deposited on top of a 30 km thick basement to which we assign a depth-independent density of 2720 kg m\(^{-3}\). In comparison to Robin and Bailey (2009), our experiments use a smaller density contrast between greenstone and basement (120 kg m\(^{-3}\) vs 200 kg m\(^{-3}\)), a thicker greenstone cover (15 km vs 10 km), and the emplacement of the greenstone cover follows a stepwise history over 140 myr rather than an instantaneous emplacement.

Deformation in our models follows the mechanism that requires the least differential stress amongst: frictional faulting (Coulomb criteria), plastic faulting and viscous creep. The frictional faulting is
Fig. 2. Lithostratigraphy of the Pilbara Supergroup. Geochronology references: a (Buick et al., 1995), b (Thorpe et al., 1992), c (McNaughton et al., 1993) and d (Nelson, 1999, 2000, 2001).

Table 1
List of parameters used in the models.

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Value(s)</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>(a) Mechanical parameters</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$g$: Acceleration of gravity field</td>
<td>9.81</td>
<td>m s$^{-2}$</td>
</tr>
<tr>
<td>$\rho_{\text{air}}$: Air density</td>
<td>2.0</td>
<td>kg m$^{-3}$</td>
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<tr>
<td>$\rho_{\text{CFB}}$: CFB density</td>
<td>2840</td>
<td>kg m$^{-3}$</td>
</tr>
<tr>
<td>$\rho_{\text{cc}}$: Crustal density</td>
<td>3310</td>
<td>kg m$^{-3}$</td>
</tr>
<tr>
<td>$C_{\text{CFB}}$: CFB cohesion</td>
<td>10</td>
<td>MPa</td>
</tr>
<tr>
<td>$C_{\text{cc}}$: Crustal cohesion</td>
<td>40</td>
<td>MPa</td>
</tr>
<tr>
<td>$\epsilon_{\text{a}}$: Strain weakening factor</td>
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<td></td>
</tr>
<tr>
<td>$\epsilon_{\text{m}}$: Strain from which weakening is maximum</td>
<td>0.5</td>
<td></td>
</tr>
<tr>
<td>$\sigma_{\text{CFB}}$: CFBs maximum yield stress</td>
<td>100</td>
<td>MPa</td>
</tr>
<tr>
<td>$\sigma_{\text{cc}}$: Crustal maximum yield stress</td>
<td>250</td>
<td>MPa</td>
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<td>$\sigma_{\text{m}}$: Mantle maximum yield stress</td>
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<td>MPa</td>
</tr>
<tr>
<td>$\phi_{\text{CFB}}$: CFB internal angle of friction</td>
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<td></td>
</tr>
<tr>
<td>$\phi_{\text{cc}}$: Crustal internal angle of friction</td>
<td>15</td>
<td></td>
</tr>
<tr>
<td>$\phi_{\text{m}}$: Mantle internal angle of friction</td>
<td>25</td>
<td></td>
</tr>
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<td>$n_{\text{CFB}}$: CFBs stress exponent</td>
<td>3</td>
<td></td>
</tr>
<tr>
<td>$n_{\text{cc}}$: Crustal stress exponent</td>
<td>3</td>
<td></td>
</tr>
<tr>
<td>$n_{\text{m}}$: Mantle stress exponent</td>
<td>3</td>
<td></td>
</tr>
<tr>
<td>$Q_{\text{CFB}}$: CFBs Activation enthalpy</td>
<td>$1.9 \times 10^{9}$</td>
<td>J mol$^{-1}$</td>
</tr>
<tr>
<td>$Q_{\text{cc}}$: Crustal activation enthalpy</td>
<td>$1.9 \times 10^{9}$</td>
<td>J mol$^{-1}$</td>
</tr>
<tr>
<td>$Q_{\text{m}}$: Mantle activation enthalpy</td>
<td>$5.2 \times 10^{9}$</td>
<td>J mol$^{-1}$</td>
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<tr>
<td>$R$: Gas constant</td>
<td>8.3145</td>
<td>J mol$^{-1}$ K$^{-1}$</td>
</tr>
<tr>
<td>(b) Thermal parameters</td>
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<td></td>
</tr>
<tr>
<td>$a_{\text{cc}}$: Mantle coefficient of thermal expansion</td>
<td>$2.8 \times 10^{-5}$</td>
<td>K$^{-1}$</td>
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<tr>
<td>$c$: Thermal diffusivity</td>
<td>$0.9 \times 10^{-6}$</td>
<td>m$^{2}$ s$^{-1}$</td>
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<tr>
<td>$C_{\text{p}}$: Heat capacity</td>
<td>1000</td>
<td>J kg$^{-1}$ K$^{-1}$</td>
</tr>
<tr>
<td>$H_{\text{CFB}}$: CFBs heat production</td>
<td>$1.335 \times 10^{-7}$</td>
<td>W m$^{-3}$</td>
</tr>
<tr>
<td>$H_{\text{cc}}$: Crustal heat production</td>
<td>$1.335 \times 10^{-6}$</td>
<td>W m$^{-3}$</td>
</tr>
<tr>
<td>$H_{\text{m}}$: Mantle heat production</td>
<td>0</td>
<td>W m$^{-3}$</td>
</tr>
<tr>
<td>$q_{\text{u}}$: Basal mantle heat flux</td>
<td>0.025</td>
<td>W m$^{-2}$</td>
</tr>
<tr>
<td>(c) Partial melting parameters</td>
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<td></td>
</tr>
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<td>Latent heat</td>
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<td>J</td>
</tr>
<tr>
<td>CFBs: Solidus $(P) = 993-1.2 \times 10^{-7} P + 1.2 \times 10^{-10} P$</td>
<td></td>
<td>°C</td>
</tr>
<tr>
<td>Liquidus $(P) = 1493-1.2 \times 10^{-7} P + 1.2 \times 10^{-10} P$</td>
<td></td>
<td>°C</td>
</tr>
<tr>
<td>Basement: Solidus $(P) = 983-9.37 \times 10^{-8} P + 6.32 \times 10^{-17} P$</td>
<td></td>
<td>°C</td>
</tr>
<tr>
<td>Liquidus $(P) = 1393-9.37 \times 10^{-8} P + 6.32 \times 10^{-17} P$</td>
<td></td>
<td>°C</td>
</tr>
</tbody>
</table>

* Non listed values for the CFB are taken equal to crustal values.
described by a yield stress: $\sigma_{\text{yield}} = (C_0 + \tan(\phi)\sigma_n) f(\varepsilon)$, in which $C_0$ is the cohesion at atmospheric pressure; $\phi$ is the angle of internal friction, $\sigma_n$ is the normal stress. Strain localisation is achieved via strain weakening function $f(\varepsilon)$, which reduces the yield stress as the accumulated strain ($\varepsilon$) increases. The strain weakening function is given by:

$$f(\varepsilon) = \begin{cases} 1 - (1 - \varepsilon / \varepsilon_0)^{\varepsilon / \varepsilon_0} & \varepsilon < \varepsilon_0 \\ \varepsilon / \varepsilon_0 & \varepsilon \geq \varepsilon_0 \end{cases}$$

where $\varepsilon$ is the accumulated strain, taken as the second invariant of the deviatoric plastic strain tensor, and $\varepsilon_0$ is the “saturation” strain from which the yield stress is reduced by a proportion $\varepsilon_n$. $\varepsilon_n$ modulates the dependency between accumulated strain and strain weakening. The yield stress has an upper limiting value $\sigma_{\text{crit}}$, which describes semi-brittle deformation independent of pressure (Ord and Hobbs, 1989). For differential stresses that attain the yield stress, the material fails and deformation is modelled by an effective viscosity: $\eta_{\text{yield}} = \dot{\gamma}_{\text{yield}} / 2\dot{E}$ in which $\dot{E}$ is the second invariant of the strain rate tensor. In our model, viscous rheologies are based on both dislocation creep and diffusion creep (Bürgman and Dresen, 2008). This choice seems justified given that (i) diffusion creep is favoured by relatively small gravitational differential stress such as the one initiating sagduction, and (ii) diffusion creep is an important flow mechanism in feldspar-bearing assemblages prevalent in TTG. Therefore, viscous creep in our model crust is modelled using the fastest of dislocation creep and diffusion creep mechanisms (Turcotte and Schubert, 1982) following: $\eta_{\text{crit}} = (1/\dot{\gamma}_{\text{dis}} + 1/\dot{\gamma}_{\text{diff}})^{-1}$. Both creep regimes can be described by an Arrhenius formulation (Karato and Wu, 1993): $\dot{\gamma}_{\text{crit}} = (N_0 \exp(Q/(RT)))^{(1-n)/n}$ with $\dot{E}$ the second invariant of the strain rate tensor, $Q$ the activation energy...
of the creep mechanism, and \( n \) the power law stress exponent. For dislocation creep viscosity: \( N_{\text{dis}} = \frac{1}{(2^{\alpha_n}A_{\text{dis}})} \), with \( A_{\text{dis}} \) the pre-exponential factor in the power law (parameter values from \textit{Brace and Kohlstedt, 1980, Table 1}). In the crust, diffusion creep becomes important under low differential stress, high temperature, and when partial melt is present (\textit{Mecklenburgh and Rutter, 2003}). For many rocks, parameters \( Q \) and \( A \) for diffusion creep are unknown. However, they must lead to faster strain rate at differential stress lower than a given transition differential stress (here \( \sigma_{\text{tr}} = 30 \text{MPa} \)) at which both mechanisms result in the same strain rate. In our models, we assume that the activation energies for dislocation creep and diffusion creep are the same and we drop the \( N_0 \) factor for diffusion creep to a value for which the dislocation and diffusion viscosities are equal at the transition differential stress. For diffusion creep, viscosity \( n = 1 \) and \( N_0 \) is obtained by matching both diffusive and dislocation strain rates at the transition differential stress \( \sigma_{\text{tr}} \): \( N_{\text{dif}} = N_{\text{dis}} = 1/(2^{\alpha_n}A_{\text{dis}})\sigma_{\text{tr}}^{-1} \cdot \text{ndis} \). We disregard diffusion creep in the mantle because the transition differential stress between dislocation creep and diffusion creep is low (ca. 0.5 MPa, \textit{Turcotte and Schubert, 1982}). In all experiments, viscosities are clipped beyond a minimum of \( 5 \times 10^{18} \text{Pas} \) and a maximum of \( 5 \times 10^{22} \text{Pas} \). All parameters are reported in \textit{Table 1}. In most CFBs, magmatic mineral assemblages are deeply altered into micas and talc-bearing retrogressive assemblages. This alteration reduces both the density and strength of mafic rocks (\textit{de Bremond d'Arts et al., 1999}). Hence, we assume an effective viscosity one tenth of that of the basement. This is achieved by imposing that: \( N_0\text{bas} = 0.1N_{\text{Crust}} \) (i.e. \( A_{\text{bas}} = 10A_{\text{Crust}} \), \textit{Table 1}).

In our experiments, the radiogenic heat production in the crust is calculated backward in time to 3.5 Ga from the averaged crustal heat production of modern Archean crusts (\textit{Taylor and McLennan, 1995}). The total crustal radiogenic production is partitioned into the pre-greenstone basement and we assume little heat production in the greenstone cover and none in the mantle (cf. \textit{Table 1}). Assuming a basal heat flux of \( 25 \times 10^{-3} \text{Wm}^{-2} \), the steady-state Moho temperature is at 710°C before the emplacement of the first member of the CFB. A free-slip boundary condition is applied to horizontal and vertical boundaries and no plate boundary force is applied. Deformation is achieved through gravity-driven processes initiated by small variations in thickness of the greenstone cover (Fig. 4).

Partial melting impacts significantly on the thermo-mechanical properties of the system (temperature, viscosities and densities). In our numerical experiments, viscosities decrease over three orders of magnitude as the melt fraction increasing from 0.2 to 0.3. In nature, the viscosity drop reaches many orders of magnitude (\textit{Clemens and Petford, 1999}), however only the first two or three orders are likely to have mechanical significance. As the temperature increases from the solidus to the liquidus, the density of the partially melted region decreases by 13%, which increases its buoyancy. Solidus and liquidus parameters are reported in \textit{Table 1}.

### 4. Results

**Crustal flow:** From \( t_0 = 140 \text{myr} \) to \( t_0 + 140 \text{myr} \), the Moho temperature increases from 704°C to 772°C leading to onset of partial melting, which is limited to only a few percent of melt at the very base of the crust (Fig. 4b1). This temperature increase is due to the thermal insulation of the radiogenic crust progressively buried under the greenstone cover (\textit{Mareschal and West, 1980; Rey et al., 2003; Sandiford et al., 2004; West and Mareschal, 1979}). There is no mechanical disturbance of the greenstone-basement interface (Fig. 4b1), which is consistent with the concordant contact between the Kelly Group and the Warrawoona Group (\textit{Van Kranendonk et al., 2004b}). At \( t_0 + 150 \text{myr} \), 10 myr after the deposition of the 7.5-km thick Kelly group, partial melting affects a growing portion of the lower crust facilitating the initiation of gravitational instabilities which strain the greenstone cover (Fig. 4b2). Convective motion affects the deep crust. From \( t_0 + 150 \text{myr} \) to \( t_0 + 158 \text{myr} \), a couple of gravitational instabilities develop, one at ca. 152 myr (Fig. 4b3), the other at ca. 158 myr (Fig. 4b4). Their length-scale (ca. 100 km) and time-scale (10 myr) are compatible with those of the dome and basin pattern of the East Pilbara Granite Greenstone Terrane (\textit{Collins et al., 1998}), and consistent with the results of \textit{Robin and Bailey, 2009}). In the upper crust, above downwelling regions, as much as 60% bulk horizontal shortening accommodates sagdution downward flow. Upward motion in domes and downward motion in basins are coupled through horizontal motions in the upper and lower part of the crust. Notably, enclaves of greenstones are caught into the upward flow in domes and exhumed from the base of the crust into the upper crust.

**Thermal evolution:** From the onset of sagdution at \( t_0 + 148 \text{myr} \), downward flow advects cool greenstone rocks in downwelling regions where the temperature at 10 km depth decreases from ca. 210 to ca. 90°C (Fig. 5a). In contrast, in rising domes the temperature at 10 km depth increases from ca. 210 to ca. 660°C (Fig. 5a). In about 12 myr, partial convective overturn leads to the build-up of a long-wavelength (ca. 100 km) profound lateral thermal anomalies >500°C through fast advective cooling and heating of downwelling and upwelling regions respectively (Fig. 5a). The maximum horizontal temperature gradient reaches 25°C/km. From \( t_0 + 158 \text{myr} \) onward, following the main sagdution stage, temperature anomalies, around fully developed domes, decrease. This second post-deformation stage, led to rapid conductive heating of greenstone basins and conductive cooling of domes (Fig. 5b). The temperature at 10 km depth in the basins increases from 90 to 200°C in ca. 8 myr, whereas the temperature at the same depth in the domes decreases from 660 to 310°C (Fig. 5b). Our numerical experiments also revealed that, because of the sheer size of the thermal anomalies, and in part because domes are diachronous, deformation and thermal relaxation last over 150–200 myr after the bulk of sagdution. This is compatible with the folding of the Gorge Creek Group (<3240 Ma) as well as the hornblende Ar–Ar ages in the EPGT that spread from 3350 to 3100 Ma (\textit{Davids et al., 1997; Kloopenburg et al., 2001; Wijbrans and McDougall, 1987}).

### 5. Comparison to temperature records derived from petrological and isotopic data

To verify the 2-stage thermal history documented in our numerical experiments we compare it to that derived from geochemical and isotopic data from the Warrawoona Synform (\textit{Huston et al., 2001; Thébaud et al., 2008}). In the Warrawoona Synform, hydrothermal alteration has been described as strongly partitioned into ductile shear zones that accommodated the sagdution of the greenstones cover (Fig. 3) (\textit{Thébaud et al., 2006}). The Fielding’s Find Shear Zone, parallel to the northern border of the felsic volcanic Wynman Formation (Fig. 3), has been recently the focus of a detailed fluid–rock interaction study. A section across this shear zone shows that the increase in strain correlates with enhanced hydrothermal alteration, including strong silicification, coupled to a pronounced zoning of major elements, trace elements and oxygen isotopes values (\textit{Thébaud et al., 2008}). The bulk rock \( ^{81}O \) values increase towards the shear zone from 19 to 25‰, indicating a fluid-buffered system (Fig. 6). This zonation is consistent with the shear zones having acted as fluids pathways. Similar extreme (up to 18‰) \(^{18}O\)-enrichments have been documented in metabasalts of the Superior Province (\textit{Kerrich et al., 1981}), and in association with Paleoarchean to Mesoarchean hydrothermal vents in Barberton (\textit{de Wit et al., 1982; Knauth and Lowe, 2003; Lowe and Byerly, 1986}).
Fig. 4. Initial settings (top box) and snapshots of the numerical experiment from $t_0 + 140$ to $t_0 + 158$ myr following the emplacement of the Kelly Group at $t_0 + 140$ myr. The history from $t_0$ is $t_0 + 140$ myr is not shown. This period corresponds to the progressive emplacement of the greenstone cover, and to a warming up of the crust. During this time there is with little to no deformation. Thickness variations of the lower Warrawoona greenstone speedup initiation of sagduction. Two circular red markers record the horizontal versus vertical motions as well as the magnitude of shortening in the greenstone cover. From $t_0 + 140$ to $t_0 + 158$ myr, gravity-driven shortening above the downwelling region is $>60\%$. Blue shading shows post yielding plastic strain. Arrows pointing at passive vertical markers in the basement document the deformation pattern.
$^{18}$O-enrichment up to ca $14\%$ have been also described in ophiolites and modern-day seafloor alteration zones and is interpreted as the product of hydrothermal convection cells involving seawater at low temperature (e.g. Putlitz et al., 2001). In Archean sagduction settings, seafloor-like alteration would have been strongly enhanced by deformation and the development of faults and shear zones above and within downwelling regions. In the Warrawoona Synform, syn-deformation alteration process was driven by interaction with large volumes of low temperatures ($\sim$90–160 °C) seawater-derived fluids during the early stage of sagduction process (Thébaud et al., 2008). This is consistent with our numerical modelling showing that downward advection of greenstones maintains a low temperature environment ($<100$ °C) in the top 10 km of the downwelling greenstones.

Fig. 5. Thermal evolution at 10 km depth. Upper graph shows the evolution from $t_0 + 140$ to $t_0 + 158$ myr following the emplacement of the Kelly Group. During this stage, the temperature in the basin decreases from 210 to 90 °C, whereas the temperature in the dome rises from 210 °C to 660 °C. This leads to a horizontal temperature difference of ca. 570 °C over ca. 65 km distance, with a horizontal gradients up to 26 °C per km. From $t_0 + 158$ to $t_0 + 166$ myr (lower graph), the temperature evolution changes with cooling of the granitic domes from 660 to 310 °C and heating in the greenstone basin from 90 to 200 °C.

In the Warrawoona Synform, ductile shear zones also host gold-bearing quartz ± calcite ± sulphide ± ankerite veins (Huston et al., 2001). These veins do not exist in the nearby less deformed Marble Bar Greenstone Belt, but they do in other downwelling regions such as the Bamboo Creek mining district on the NE margin of Mount Edgar (white hexagon in Fig. 1b1). These gold-bearing quartz veins develop largely in association with shear zones in downwelling regions and have a syn- to post-shearing origin (Thébaud et al., 2006). Across the Warrawoona Synform, quartz veins display a restricted range of $\delta^{18}$O values with a mean value of $+13.2 \pm 2\%$ (Fig. 6). This suggests that quartz precipitated from a homogeneous fluid under near-isothermal conditions. These veins are therefore in isotopic disequilibrium with their silicified host, which implies an emplacement postdating the silicification (Thébaud et al., 2008). Quartz-hosted CO$_2$–NaCl–H$_2$O–CH$_4$ fluid inclusions yielded homogenisation temperatures between 234 and 372 °C (Huston et al., 2001; Thébaud et al., 2006). Using Zheng (1993) quartz/water fractionation equation, Thébaud et al. (2008) calculated that the $\delta^{18}$O of the water from which these quartz veins precipitated is in the range of $+2.4$ to $+12.1\%$. This oxygen isotopic composition, together with the significant enrichment in potassium and base metals revealed by Synchrotron radiation X-ray fluorescence fluid inclusion analyses, support a metamorphic and/or magmatic source (Thébaud et al., 2006). These hotter fluids were mobilised at a later stage of the sagduction process, possibly during the devolatilisation of the greenstones keels and/or during the large production of potassic melt (Thébaud et al., 2008).

In summary, field observation, isotopic and fluid inclusions microthermometry data, and our numerical investigation presented here, document a long-lived two-stage hydrothermal history involving: (1) low-temperature (90–160 °C) syn-deformation seafloor-like fluid–rock interactions strongly focused along steeply dipping ductile shear zones during the early stage of sagduction, and (2) moderate-temperature (234–372 °C) syn- to post-deformation fluid–rock interactions, strongly focused along shear zones, leading to the formation of gold-bearing quartz veins...
post-dating the vertical fabrics. The early alteration assemblages can be linked to the early stage of greenstone sagduction and the co-development of (i) crustal-scale thermal anomalies, during which cooling affected the downwelling greenstone, and (ii) vertical shear zones which acted as efficient fluid pathways for supracrustal porous flow (Fig. 7a). In the early stage of sagduction, large volumes of fluid, trapped in hydrated minerals during seafloor alteration, would have been transported deep into the crust. The second hydrothermal stage can be linked to the release of metamorphic and/or magmatic fluid in association to the devolatisation of sagducted greenstone covers, and crystallisation of felsic magmas (Fig. 7b). These fluids would have been released at a time when sagduction-related lateral thermal anomalies entered the relaxation phase during which the temperature in the greenstone keels rapidly increased. In the strongly foliated steeply dipping greenstone sheets, these hot crustal fluids would have been efficiently channelled toward the surface through the well-established network of shear zones formed at an early stage. This is consistent with field observations showing that quartz veins are restricted to downwelling regions. The slow conductive relaxation of the thermal anomaly may have helped to maintain this plumbing system for several 10 s of myr.

6. Discussion and implication for Paleoarchean to Mesoarchean gold mineralisation

To form a gold deposit, the average crustal gold content of 1.3 ppb must be concentrated by one thousand to one hundred thousand times (e.g. Walshe and Cleverley, 2009). Temperature gradients above shallow magmatic intrusions typically drive
Fig. 7. Conceptual tectonograms illustrating Archean intracontinental mineral systems associated to gravity driven tectonics and fluid flow during sagduction of subaqueous greenstone covers (green envelopes) into their hot and weak basement (red envelopes). During the early stage of sagduction (upper panel), early deformation enhances low-temperature seafloor-like alteration along downwelling regions. Sagduction transfers cooler supracrustal rocks and large volumes of hydrated minerals deep into the crust. As shown in our numerical experiments, dome regions become a few hundred degrees hotter than the intervening downwelling regions. These lateral thermal gradients contribute to focus fluid circulations into downwelling regions. In a second stage (lower panel), the relaxation of this lateral temperature anomaly led to the rapid heating and devolatisation of hydrated steeply dipping greenstone sheets. Released metamorphic and magmatic fluids focus into pre-existing shear zones, which act as crustal-scale fluid pathways to the surface where mineral deposits form. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of the article.)

Efficient hydrothermal systems in which fluid–rock interaction is the main concentration mechanism (Eldersi et al., 2009). It is therefore not surprising that shallow magmatic activities in back-arc settings, as well as sub-seafloor volcanism associated with continental rift systems, are the main settings of gold deposits on the modern Earth (Goldfarb et al., 2001). In other words, mineralised hydrothermal systems are strongly partitioned onto ocean floors and active continental margins. However, it is likely that most Archean ocean floors and Archean continental plate margins did not survive the past 3 Gyr. This suggests that the bulk of the preserved Archean gold deposits may have formed in the interior of continents, away from plate margins. On the modern Earth, this setting is the least favourable for hydrothermal systems, but perhaps not so in the Archean.

In the Archean, basaltic piles up to 15 km thick emplaced on older felsic continental crusts were common (de Wit and Ashwal, 1997), and most of them emplaced onto flooded continents (Arndt, 1999; Flament et al., 2008; Kump and Barley, 2007). Higher sea level in the Archean would have maintained an infinite reservoir of water above the continental basalts, whereas higher surface heat flow, due to enhanced radiogenic heat production in the crust and due to CFBs thermal insulation, would have powered shallow hydrothermal cells at the interface between continent and seawater, accounting for widespread low temperature alteration and silicification observed in greenstone belts. Numerical and physical experiments have shown that density inversions related to the emplacement of thick continental basalts onto less dense and weak basement are large enough to drive the foundering of basaltic piles into the continental crust (Chardon et al., 1998; de Bremond d’Ars et al., 1999; Dixon and Summers, 1983; Robin and Bailey, 2009; West and Mareschal, 1979; this study). Our numerical experiments show that sagduction is capable of promoting significant horizontal temperature gradients up to 26 °C/km. These horizontal gradients may have been large enough to force fluids, released during devolatisation of greenstones and crystallisation of domes, back into the cooler greenstone keels, where steeply dipping shear zones would have channelised these hotter fluids toward the surface. Importantly, sagduction is a mechanism that explains crustal-scale deformation within the inner part of continental plates, away from their margins. Therefore in the Archean, the association of volcanic rocks above a felsic crust but below sea level, in tectonically active regions with high heat flow – critical to formation of gold deposits – was not restricted to continental rifting or active continental margins. Sagduction on flooded continents would have promoted the formation intracontinental hydrothermal systems expanding considerably the prospectivity of Archean cratons, compared to that of modern continents whose prospectivity is limited to active plate margins and continental rifts.
7. Conclusions

Using realistic rheologic and thermal parameters, our coupled thermo-mechanical numerical experiments confirm that, in the Archean, sagduction was a natural response to density inversions and thermal weakening due to the emplacement of thick continental flood basalts onto radiogenic crusts. Our experiments show that, during sagduction, the advection of cold rocks into downwelling basins and advection of hot rocks into rising domes lead to crustal-scale, horizontal, thermal anomalies in excess to 500 °C over an horizontal distance of ca. 30 km. Large temperature gradients up to 26 °C/km and the availability of large volumes of fluids (i.e. (i) seawater above flooded continents, (ii) metamorphic fluids released during dehydration of sagducted supracrustal greenstones, and (iii) magmatic fluids released during crystallisation of magma in domes) would have powered crustal-scale hydrothermal systems. The formation around granitic domes of networks of sagduction-related steeply dipping greenstone sheets and vertical shear zones connected through vertical triple junctions, would have facilitated crustal-scale transfer of fluids, matter and heat. Fluid–rock interactions in these tectonically active plumbing systems would have overprinted low-temperature fluid–rock interactions that occurred during the emplacement of CFBs onto flooded continents and during the early stage of sagduction. This is consistent with structural, petrological and isotopic data. Based on our numerical experiments, field observations and existing petrological and isotopic data, we propose that in the Archean, the sagduction of thick greenstone piles with a large and near-permanent seawater reservoir above, and a hot and weak basement below, allowed for the development of efficient and long-lived plumbing systems. This plumbing systems favoured crustal-scale fluid–rock interactions and the formation of craton-wide gold deposits in the interior of continents, far away from their margins, a context very different from the palaeoproterozoic and Phanerozoic, where orogenic and magmatic orebodies are closely associated with active margin settings.

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