Phase equilibria constraints on the melt fertility of crustal source rocks: The effect of sub-solidus water loss

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<th>Journal:</th>
<th>Journal of Metamorphic Geology</th>
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<td>Manuscript ID:</td>
<td>JMG-14-0034</td>
</tr>
<tr>
<td>Manuscript Type:</td>
<td>Original Article</td>
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<tr>
<td>Date Submitted by the Author:</td>
<td>01-May-2014</td>
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<tr>
<td>Complete List of Authors:</td>
<td>Webb, Gordon; University of Melbourne, School of Earth Sciences Powell, Roger; University of Melbourne, School of Earth Sciences McLaren, Sandra; University of Melbourne, School of Earth Sciences</td>
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<td>Keywords:</td>
<td>prograde metamorphism, melting, water, anatexis, THERMOCALC</td>
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Phase equilibria constraints on the melt fertility of crustal source rocks: The effect of sub-solidus water loss

G. Webb, R. Powell and S. McLaren

School Earth Sciences, The University of Melbourne, Vic 3010, Australia

*Corresponding author: gwebb@student.unimelb.edu.au

Short title: Prograde water loss and crustal melt fertility
ABSTRACT

During prograde metamorphism, free H$_2$O generated by ongoing dehydration reactions is likely to be continuously lost from a rock primarily driven by compaction. Classical melting experiments cannot easily simulate this phenomena because all run products are ideally conserved within the experimental charge even though significant equilibration and H$_2$O generation may occur during heating. Phase equilibria modelling is used to model the effect of subsolidus water loss on subsequent melting relationships of felsic lithologies (including metapelite, metagreywacke and metatonalite) in the suprasolidus. Subsolidus water loss drives the bulk composition towards the minimum saturation point within the wet melting field and results in significantly reduced subsequent melt generation when compared to melting experiments involving conservation of H$_2$O in an experimental charge from some temperature well below the solidus. This effect is most obvious at $P$-$T$ conditions just above the solidus. The reduction in melt fertility due to subsolidus water loss causes rheologically critical thresholds to be shifted to higher temperatures (e.g. melt connectivity threshold, melt escape threshold and the solid-to-liquid transition). Subsolidus water loss is also likely to impact on crystallising melts as they cross the solidus. As a result, the H$_2$O content of granites may not reflect the H$_2$O contents of the magma from which they were derived. Instead they record the H$_2$O contents in the vicinity of the minimum saturation point at the $P$-$T$ conditions at which they crossed the solidus.

keywords: prograde metamorphism, melting, water, anatexis, THERMOCALC
INTRODUCTION

Understanding crustal melt fertility is important for developing robust models for migmatite formation, granite petrogenesis and the differentiation and evolution of the continental crust (Brown & Fyfe, 1970; Huppert & Sparks, 1988; Clemens, 1990; Thompson, 1996; Sawyer, 1998; Annen et al., 2008). Hydrous felsic lithologies melt incongruously at mid- to lower-crustal depths (10–60 km) at temperatures attainable during high-grade metamorphism (650–1000°C; Thompson & Connolly, 1995; Thompson, 1999; White et al., 2001). The amount of melt produced and its composition is determined by the composition of the source rock, the pressure and temperature conditions and any open system behaviour including the loss or gain of fluid, melt or solid phases. Of these, H$_2$O content is one of the key compositional variables affecting melting behaviour in the crust (Tuttle & Bowen, 1958; Holtz et al., 1995; Johannes & Holtz, 1996).

Two dominant modes of melting have been identified to explain the observed characteristics of natural granites, migmatites and granulites. ‘Fluid-present melting’ or ‘wet melting’ describes the situation where a system is saturated with respect to a H$_2$O-dominated fluid phase as it crosses the wet solidus with increasing $T$ and/or $P$. In fluid-present melting the activity of H$_2$O is effectively unity ($a_{H_2O} \approx 1$), neglecting the proportions of other constituents in the fluid that are likely to be small, at least for most crustal conditions. The resulting melt phase is saturated with respect to H$_2$O. Fluid-present melting may occur in some migmatite terrains (especially at low $P$; e.g. Sawyer, 2010). However it is not generally considered to be the dominant mode of melting in the continental crust (Clemens, 1984). By contrast, ‘fluid-absent’, ‘vapour-absent’ or ‘dehydration’ melting proceeds via reactions involving the breakdown of hydrous minerals.
(mainly muscovite, biotite and hornblende) in the absence of a free fluid \( (a_{H_2O} < 1) \) at higher temperatures than the wet solidus. It has been argued that fluid-absent melting is the dominant mechanism for the formation of most granite magmas, which are generally thought to have formed at high \( T \) and are undersaturated with respect to \( H_2O \) (Clemens, 1984; Pickering & Johnston, 1998).

Estimates of the amount of melt generated during partial melting have largely come from melting experiments on a variety of starting compositions including metapelites (Vielzeuf & Holloway, 1988; Carrington & Harley, 1995; Gardien et al., 1995; Pickering & Johnston, 1998), metagreywackes (Patino-Douce & Beard, 1996; Vielzeuf & Montel, 1994; Montel & Vielzeuf, 1997), tonalites (Huang & Wyllie, 1986; Rutter & Wyllie, 1988; Skjerlie & Johnston, 1993; Patino-Douce, 1997; 2004) and amphibolites (Rushmer, 1991; Wolf & Wyllie, 1994). The majority of recent experimental studies have focussed on fluid-absent melting because of its perceived importance in magma genesis and crustal differentiation. The results of these fluid-absent melting studies indicate that different crustal lithologies display markedly different melting behaviour at similar \( P–T \) conditions. Furthermore, different melt fractions have been reported for compositionally-similar starting materials (e.g. Gardien et al., 1995).

Fluid-absent melting experiments are often described as having ‘no added water’ or being ‘vapour-absent’. This is generally achieved by pre-heating the starting material above \( \sim 150°C \) for a sustained period (ca. 24 hours), causing any adsorbed \( H_2O \) to escape, prior to sealing the experimental charge (e.g. Patino-Douce, 2004). It is then assumed that any \( H_2O \) retained within the starting material is structurally bound within hydrous minerals (e.g. muscovite, biotite and amphibole). However, hydrous minerals that are stable up to 150 °C at ambient pressures may begin to break down, via subsolidus dehydration reactions,
as the experimental charge is heated (and pressurised) to the desired conditions of the
experimental run. Because of the intentional ‘closed system’ nature of most high \( P \) melting
experiments, any \( \text{H}_2\text{O} \) generated via subsolidus dehydration will be effectively trapped
within the experimental charge. In this way a ‘vapour absent’ starting material can evolve,
becoming ‘fluid-present’ at the experimental \( P–T \) conditions. Additionally, London et al.
(2012) suggested that powdered samples loaded in air may absorb between 1 and 4 weight
percent \( \text{H}_2\text{O} \) before the experimental charge is sealed. It has also been suggested that
water is often lost from the experimental charge via diffusion of molecular \( \text{H}_2\text{O} \) through the
metal casing at temperatures above \( \sim 950^\circ \text{C} \) as suggested by decreasing melt modes and
increasing modal proportions of solid phases (e.g. plagioclase and garnet) with increasing \( T \)
during experimental runs (Patino-Douce & Beard, 1994). The combined effects of
absorption during handling and diffusive loss at high temperature work in opposite
directions, thereby increasing the uncertainty in the actual \( \text{H}_2\text{O} \) content inside the
experimental charge during high temperature melting experiments. Furthermore, if \( \text{H}^+ \) is
lost through the capsule rather than molecular \( \text{H}_2\text{O} \), as has been suggested by some
authors, this may affect the oxidation state during a melting experiment (Patino-Douce &
Beard, 1994; White et al., 2011).

An important constraint on the amount of \( \text{H}_2\text{O} \) available for melting reactions during
regional metamorphism can be obtained by considering the evidence for the nature of the
fluid regime prior to the onset of melting. During prograde metamorphism of rocks with an
usually high \( \text{H}_2\text{O} \) content (e.g. metapelites and metagreywackes), free \( \text{H}_2\text{O} \) is likely to be
continuously generated via devolatilisation reactions up to the point at which the wet
solidus is crossed with increasing \( T \) and/or \( P \). Above this point, \( \text{H}_2\text{O} \) is consumed via wet
melting reactions and decreases until it is no longer stable as a free phase at progressively
higher T. Several authors have argued that any H\textsubscript{2}O generated during subsolidus dehydration has a strong tendency to escape the source rock due to the combined effects of compaction and increasing fluid pressure (Thompson & Connolly, 1990; Connolly, 1997; Yardley, 2009). Decreasing H\textsubscript{2}O content with increasing metamorphic grade is observed in most metamorphic terrains providing evidence that much of the H\textsubscript{2}O generated during subsolidus dehydration escapes the system prior to the onset of melting (Miyashiro, 1961). Indeed, the loss of fluid from metamorphic terrains may be a requirement for preserving metamorphic isograds during retrograde cooling and/or decompression (Thompson, 1983; Guiraud et al., 2001). Subsolidus H\textsubscript{2}O loss (SWL) may be especially significant at the pressure conditions of the middle to lower crust where rock porosity is thought to be small (<0.1–0.5 %), and positive excursions in hydrostatic pressure above the prevailing lithostatic pressure cannot be maintained for significant time periods (Norton & Knapp, 1977; Connolly, 1997; 2010).

Importantly, there are few existing datasets to constrain sub-solidus open system processes involving H\textsubscript{2}O, and experimentally determined phase relationships do not adequately constrain melting relationships, particularly with respect to H\textsubscript{2}O. Quantitative thermodynamic calculations using internally-consistent datasets can be readily employed in multicomponent systems that approach the compositional complexity of natural rocks (e.g. White & Powell, 2002; Powell et al., 2005; White et al., 2005; White & Powell, 2010). P-X or T-X pseudosections can be used to visualise the effect of changes in bulk composition, as a result of the loss of a fluid and/or solid component from the system. This kind of thermodynamic modelling has the potential to complement experimental studies to provide a clearer picture of open-system processes that are thought to occur in the continental crust. Here we use pseudosections to investigate the effect of SWL on melting relationships
for a variety of bulk compositions representing common crustal rock types (metapelites, metagreywackes, tonalites, granodiorites). The impact of SWL on melt fertility is modelled at temperatures appropriate to partial melting during regional metamorphism at mid- to lower-crustal pressures. The modelling results are compared to the results derived from melting experiments to provide improved constraints on the nature of melting during crustal anatexis.

CALCULATED PHASE RELATIONS

Mineral equilibria calculations were undertaken using THERMOCALC 3.37i (Powell & Holland, 1988; updated 2013) with an updated version of the internally consistent data set of Holland & Powell (1998; data set tcds55, file created 22 November 2003). The calculations were undertaken in the chemical system Na₂O-CaO-K₂O-FeO-MgO-Al₂O₃-SiO₂-H₂O-TiO₂-O (NCKFMASHTO) of White et al. (2007). The phases considered in the calculations, and references to the activity-composition models used are: garnet, biotite and silicate melt (White et al., 2007), cordierite and epidote (Holland & Powell, 1998), orthopyroxene and magnetite (White & Powell, 2002), muscovite (Coggon & Holland, 2002), plagioclase - K-feldspar (Holland & Powell, 2003), ilmenite (White et al., 2000), and hornblende (Diener et al., 2007).

Sillimanite, kyanite and quartz are pure end-member phases. Mineral abbreviations used in diagrams and in the text are: opx – orthopyroxene; g – garnet; cd – cordierite; hb – hornblende; bi – biotite; mu – muscovite; ky – kyanite; sill – sillimanite; ksp – K-feldspar; pl – plagioclase; ilm – ilmenite; mt – magnetite; q – quartz; liq – silicate liquid/melt.

Bulk compositions for phase equilibria modelling were derived from starting
compositions of published melting experiments for which melt fraction (Φ) estimates are available (Table 1). The four starting compositions were chosen to represent common felsic rock types and have compositions that are comparable to the average compositions of the middle and upper crust (Fig. 1). Two metapelitic bulk compositions are from Vielzeuf & Holloway (1988) and Gardien et al. (1995) and denoted VH88P and G95P, respectively. VH88P is modelled on a natural metapelite from the Cabo Ortegal metamorphic complex (Galicia, NW Spain) composed of quartz (39 wt%), plagioclase (19 wt%), kyanite (7 wt%), muscovite (9 wt%), biotite (21 wt%), garnet (2 wt%), staurolite (1 wt%), secondary chlorite (1 wt%) and accessory minerals (1 wt%). VH88P is strongly peraluminous and has comparatively high molar K$_2$O/Na$_2$O and high FeO contents. G95P is derived from a model two-mica metapelite generated from a mixture of quartz (39 wt%), plagioclase (30 wt%), muscovite (15 wt%) and biotite (15 wt%). It is weakly peraluminous and has comparatively low K$_2$O/Na$_2$O, high SiO$_2$ and low CaO. A metagreywacke bulk composition (denoted VM94G) is derived from melting experiments on a natural metagreywacke from the vallée de la Beaume (Ardèche, France) composed of quartz (40 wt%), plagioclase (32 wt%), and biotite (25 wt%), and accessory minerals (3 wt%; Vielzeuf & Montel, 1994; Montel & Vielzeuf, 1997). This composition is weakly peraluminous with high K$_2$O/Na$_2$O, SiO$_2$ and low FeO. A tonalitic bulk composition (denoted PD97T) was derived from melting experiments on a natural tonalite starting material from the Sierra Nevada Batholith (Lee Vining Canyon area; Patino-Douce, 1997). The sample was reported to contain of quartz (20 wt%), plagioclase (45 wt%), K-feldspar (7 wt%), hornblende (13 wt%), biotite (13 wt%) and accessory minerals (2 wt%). PD97T is metaluminous with low K$_2$O/Na$_2$O, SiO$_2$ and high CaO and Na$_2$O.

Some of the bulk compositions contain low concentrations of F, P, Mn and CO$_2$. Small
amounts $P_2O_5$, $CO_2$ and $MnO$ are unlikely to have a measurable effect on the melting
behaviour of felsic rocks, whereas $F$ contents in the range 0.01–4 wt% may have the effect of
lowering the solidus of granitic melts (Manning, 1981) and stabilising biotite to higher
temperatures (Peterson et al., 1991). Fluorine is not currently incorporated into the
activity-composition models for biotite or melt. As a result, the models presented here may
differ slightly in terms of melting behaviour from the more complex starting compositions of
the corresponding experiments. Furthermore, it should be noted that the melt model used
here does not incorporate $TiO_2$ or ferric iron. $TiO_2$ is readily incorporated into biotite and
oxide minerals (e.g. ilmenite and magnetite) and is important for stabilising biotite at high
temperature (White et al., 2007). Hayden & Watson (2007) investigated the incorporation
of $TiO_2$ into haplogranitic melt. Their results indicated that even for rutile-saturated bulk
compositions the amount of $TiO_2$ incorporated into the melt phase was small and as such,
is unlikely to have a first order effect on melt fertility. The ferric iron content of all of the
bulk compositions in this study are fixed at low values such that the only stable iron oxide
minerals are ilmenite ± magnetite.

For each starting composition the $H_2O$ content was varied to produce a low-$H_2O$ and a
high-$H_2O$ end-member (Table 2). The high and low values were chosen to encompass the
transition from $H_2O$ undersaturated to $H_2O$saturated conditions at the wet solidus. Each
end-member pair were then used to construct $T–M_{H_2O}$ pseudosections (Guiraud et al.,
2001) for each model starting composition. Diagrams were constructed at 10 kbar for the
starting compositions VH88P, G95P and VM94G and at 8 kbar for composition PD97T to
facilitate comparison with the reported experimental results. Models were also generated at
5 kbar for VH88P, G95P and VM94G and at 4 kbar for PD97T to investigate melting
equilibria at low pressure.
Pseudosections

$T-M_{H_2O}$ pseudosections calculated for compositions VH88P, G95P, VM94G at 10 kbar, and PD97T at 8 kbar over the temperature range 450 to 1100°C are shown in Fig. 2. The pseudosections for all four compositions have a sub horizontal $H_2O$-saturated solidus (wet-solidus). VH88P has the highest $T$ wet solidus at $\sim$680°C. The wet solidi for G95P, VM94G and PD97T are all similar, occurring at $\sim$645°C. In each case the wet solidus extends beyond the range of the high $H_2O$ end-member but is limited at low $H_2O$ contents by the stability of free $H_2O$. The point at which the solidus is minimally saturated with respect to $H_2O$ is referred to here as the minimum saturation point (MSP; Fig. 2). The narrow fields immediately above the wet-solidus in all four $T-M_{H_2O}$ pseudosections are low variance fields related to the underlying NCKASH univariant reaction,

$$pl + q + H_2O \pm mu/ksp = liq.$$  

(1)

The $H_2O$-out boundaries of these ‘wet-melting’ fields are intersected within 10–40°C above the wet-solidus. The upper boundaries increase in $T$ only slightly as they trend from the minimum saturation point towards the high $H_2O$ end-member for each diagram.

At $H_2O$ contents less than the value of the minimum saturation point the temperature of the solidus increases steeply as the equilibrium composition becomes increasingly undersaturated with respect to $H_2O$. Inflections in the solidus curves at low $H_2O$ contents on each of the $T-M_{H_2O}$ pseudosections correspond to changes in the modal mineralogy of low variance fields emanating from ‘vapour absent’ univariant reactions in the underlying NCKASH and NCKFMASH systems including,
mu + pl + q = ksp + sill/ky + liq; \hspace{1cm} (2)

bi + sill/ky + pl + q = g + ksp + liq; \hspace{1cm} (3)

bi + pl + q = opx + ksp + liq. \hspace{1cm} (4)

For all compositions, the subsolidus H$_2$O-out boundary trends to lower bulk H$_2$O contents as it approaches the solidus (at the minimum saturation point) with increasing $T$ (Fig. 2). The H$_2$O-out boundary for VH88P rises steeply from $<450^\circ$C up to the plagioclase-in boundary ($\sim$615$^\circ$C) and then migrates towards a lower bulk H$_2$O value as it approaches the solidus. For compositions G95P and VM94G the H$_2$O-out boundary is steep and approximately linear from $<450^\circ$C up to the wet-solidus. The H$_2$O-out boundary for PD97T trends gradually towards lower bulk H$_2$O, with distinct inflections corresponding to muscovite-out, K-feldspar-in and hornblende-in boundaries, as it approaches the minimum saturation point at the wet solidus with increasing temperature.

The pseudosections for VH88P and G95P both contain sillimanite and kyanite. For VH88P, sillimanite is stable from the sill-ky phase transition ($\sim$800$^\circ$C) up to $>1000^\circ$C at high H$_2$O contents and is stable above 1100$^\circ$C at low H$_2$O contents (Fig. 2 a & b). Kyanite is stable from 800$^\circ$C down to below 450$^\circ$C over the entire range of modelled H$_2$O contents (0.11–14.02 mol%). Sillimanite is stable for G95P from the sill-ky phase transition to above 1100$^\circ$C near the low-H$_2$O end-member, but its upper limit of stability decreases to $\sim$900$^\circ$C at the high-H$_2$O end-member. Kyanite is stable from 800$^\circ$C down to 450$^\circ$C at H$_2$O contents below $\sim$3 mol%. Above this value the stability is limited by the kyanite-out boundary along the low $T$ sides of the quadrivariant bi-mu-ksp-pl-liq-ilm-q-ky and the quinivariant bi-mu-pl-liq-ilm-q-ky fields. Neither sillimanite nor kyanite are stable for
VM94G and PD97T over the range modelled (Fig. 2 c & d).

K-feldspar is unstable at temperatures above the solidus at high bulk H$_2$O contents for all four modelled compositions. At bulk H$_2$O contents close to the minimum saturation point, K-feldspar is stable over the ranges $\sim$790–1000°C, $\sim$700–980°C, $\sim$810–990°C and <450–990°C for compositions VH88P, G95P, VM94G and PD97T, respectively.

It is important to note that even at high $T$ (>1000°C) none of the bulk compositions reach their liquidus and form pure melts (including the metatonalite - PD97T; Fig. 2). Even for the high H$_2$O end-members the high $T$ equilibrium assemblages comprise liq-g-ilm, liq-opx-ilm, liq-opx-ilm-q and liq-hb-pl-ilm for VH88P, G95P, VM94G and PD97T, respectively. Lower variance high $T$ assemblages become increasingly stable at lower bulk H$_2$O contents in each pseudosection.

Ilmenite is stable for all for starting compositions over the entire modelled ranges of temperature and H$_2$O contents. The pseudosection for composition G95P contains rutile at low temperature and magnetite is stable for intermediate H$_2$O contents between 800–880°C. Staurolite is stable from just below to just above the solidus for composition VM94G. While these phases are involved in the NCKFMASHTO equilibria, their modal abundances (1-oxide molar basis) are always low (<0.01) and are, therefore, unlikely to exert a significant influence on the melting behaviour of the system modelled.

Prograde metamorphism and sub-solidus water loss

The contours of melt and free H$_2$O corresponding to the pseudosections in Fig. 2 are shown in Figure 3. While each $T$–$M_{H_2O}$ pseudosection displays distinct phase relationships in detail, the relationships between the free H$_2$O and melt bearing regions in each case are
broadly similar. The proportion of free H$_2$O present along a particular prograde heating path in the subsolidus (vertical line) can be described in terms of df/dT (representing the change in free H$_2$O productivity with increasing T). Similarly, the amount of melt produced along a path in the suprasolidus can be described in terms of dm/dT (i.e. change in melt productivity with increasing T; Johnson et al., 2008).

At sub-solidus conditions df/dT is positive from <450°C up to the solidus at bulk H$_2$O values above the H$_2$O-out boundary for each of the four compositions (grey regions in Fig. 3). As a consequence, all four modelled compositions will produce free H$_2$O along a prograde path, with fixed H$_2$O contents, that traverses the ‘free H$_2$O-stable’ region (i.e. bulk H$_2$O contents greater than the value of the minimum saturation point). At supra-solidus conditions, H$_2$O & melt are both stable within the narrow wet-melting fields. Within this region, df/dT becomes negative and the amount of free H$_2$O decreases rapidly towards the H$_2$O-out boundary with increasing T. This region also corresponds to an initial increase in melt production with melt contours that are perpendicular to the T-axis (dm/dT > 10 mol%/10°C). Above the H$_2$O+melt region the melt contours have a slope influenced by the shape of the H$_2$O-undersaturated solidus such that increases in T and/or bulk H$_2$O correspond to increased melt production.

An idealised melting experiment involves conserving all of the starting components (including the initial H$_2$O contents) throughout the experimental run. This behaviour is represented by the conservation of subsolidus water (CSW) scenario, shown by the vertical grey arrows in Fig. 3. The water contents for the CSW scenario for VH88P, G95P and VM94G are derived from the reported water contents for the bulk starting material of the original experiments (2.15 w%, 1.9 w% and 1.44 w%, respectively; Table 2). The H$_2$O contents of PD97T was not reported. The CSW path for this composition was chosen
so as to intersect the observed melt fraction in the experiments (i.e. $\Phi = 0.30$ melting at 950°C; Patino-Douce, 1997).

If subsolidus water loss occurs, free H$_2$O will be incrementally lost along the prograde heating path (Guiraud et al., 2001), continuously driving the bulk composition towards the H$_2$O-out boundary. To reduce the bulk H$_2$O contents of the source rock below the value of the H$_2$O-out boundary would require the liberation of structurally bound H$_2$O from hydrous minerals. As there is no viable mechanism to achieve this, the subsolidus H$_2$O-out boundary effectively limits the minimum H$_2$O-contents of the source rock. As the rock crosses the solidus with increasing temperature any remaining free H$_2$O is rapidly consumed within the wet-melting fields and the newly generated melt becomes increasingly undersaturated with respect to H$_2$O. Because such melts are effectively ‘water-starved’ as soon as they cross the solidus, the H$_2$O contents of the system become fixed at the value of the minimum saturation point, if H$_2$O loss is efficient and the porosity is small, as generally suggested. As a result an isobaric prograde heating path for a rock composition affected by SWL will involve:

1. A sub-solidus path that traverses the free H$_2$O region at bulk H$_2$O contents just greater than the H$_2$O-out boundary as it approaches the minimum saturation point;

2. A supra-solidus path with the bulk H$_2$O fixed at the value of the minimum saturation point up to the maximum temperature of the system.

In natural systems, H$_2$O may additionally be lost via incorporation into escaping melt or as part of hydrous peritectic minerals that are themselves entrained into escaping melt (Clemens & Stevens, 2012). However, both of these mechanisms will lead to other changes in the bulk composition that cannot be determined by varying any single compositional
variable. The appropriate diagrams for modelling such phenomena are more complex and
involve varying the equilibrium bulk composition at each increment of melt- and/or
mineral-loss.

The mineral mode changes along the subsolidus sections of the SWL paths were
calculated along the H$_2$O-out boundary so that the bulk H$_2$O contents continuously changes
along this segment. The supra-solidus portion of the SWL path was modelled at a fixed
H$_2$O value (corresponding to the minimum saturation point; Table 2). Changes in the
modal abundances of the equilibrium assemblages along the modelled CSW and SWL paths
for the four starting compositions are shown in Fig. 4.

For composition VH88P the SWL heating path results in significantly less melt
generation when compared with the CSW path (especially at low T; Fig. 4a & 4b). The
reduction in melt fraction along the SWL path corresponds to increases in the modal
proportions of quartz, plagioclase and biotite. K-feldspar is stabilised at temperatures
above ∼790°C, whereas K-feldspar is completely absent in the CSW scenario. Very high
dm/dT melting (20 mol%/10°C) along the CSW path (Φ < 0.01 to ∼ 0.11) corresponds to
the H$_2$O-out boundary just above the solidus, at ∼680°C. This feature is not observed
along the SWL path. Moderate dm/dT melting (∼3 mol%/10°C) over the interval
790–880°C on the SWL path corresponds to breakdown of biotite and increased growth of
garnet. Melting at higher temperatures continues to increase in a broadly monotonic
fashion reaching melt fractions of ∼0.65 and ∼0.44 at 1000°C for the CSW and SWL paths,
respectively (dm/dT ∼1–2 mol%/10°C). Interestingly, muscovite begins to breakdown
below the solidus (at ∼615°C, corresponding to the first appearance of plagioclase) and
completely disappears just above the solidus, at ∼720°C, for both the SWL and CSW
paths; however, there is no significant increase in the melt fraction corresponding to the
disappearance of muscovite from the equilibrium assemblage along either path.

The SWL path for G95P corresponds to reduced melt fertility and slight increases in the modal proportions of quartz and plagioclase at temperatures just above the solidus (\(\sim 640^\circ C\)) and an increase in the proportion of K-feldspar at temperatures above 760\(^{\circ}\)C when compared to the CSW path (Fig. 4c & 4d). Two steep increases in melting occur along the CSW path at low melt fraction. The first corresponds to the H\(_2\)O-out boundary just above the solidus (\(dm/dT \sim 6-8\) mol%/10\(^{\circ}\)C). The second step corresponds to the abrupt disappearance of muscovite and the first appearance of kyanite and K-feldspar (at \(\sim 760^\circ\)C; \(dm/dT \sim 10-20\) mol%/10\(^{\circ}\)C). Due to the negligible amount of H\(_2\)O at the solidus in the SWL scenario the melt fraction remains low (\(\Phi < 0.02\)) until muscovite disappears at \(\sim 760^\circ\)C at which point it increases abruptly to (\(\Phi \sim 0.09\)). Melting at higher temperatures continues to increase steeply with melt fractions reaching \(\sim 0.69\) and \(\sim 0.51\) at 1000\(^{\circ}\)C for the CSW and SWL paths, respectively (\(dm/dT \sim 3-5\) mol%/10\(^{\circ}\)C).

For composition VM94G, the SWL scenario results in significantly reduced melt fertility, moderate increases in the modal proportions of quartz and plagioclase at temperatures just above the solidus (Fig. 4e & 4f). K-feldspar increases in modal abundance and becomes stabilised over a broader temperature range (\(\sim 805-1000^\circ\)C) when compared to the CSW path (\(\sim 905-930^\circ\)C). A sharp increase in melting along the CSW path corresponds to the H\(_2\)O-out boundary just above the solidus (at \(\sim 650^\circ\)C; \(dm/dT \sim 8-10\) mol%/10\(^{\circ}\)C). Again, this feature is absent from the SWL path. Immediately above the solidus the melt fraction remains low (\(\Phi < 0.01\)) along the SWL path with an initial, gradual increase in melt fraction, beginning at \(\sim 755^\circ\)C, corresponding to the first appearance of garnet and the initial breakdown of biotite (\(dm/dT\) increasing from <0.1 up to 2 mol%/10\(^{\circ}\)C). A steep increase in melting, from \(\Phi \approx 0.09\) to \(\Phi \approx 0.19\), corresponds to
the final disappearance of biotite and the first appearance of orthopyroxene across the temperature interval \( \sim 895–910^\circ C \) for the SWL path \((dm/dT \sim 7 \text{ mol}\%/10^\circ C)\). A similar step occurs along the CSW path over the same temperature interval with melting abruptly increasing from \( \Phi \approx 0.28 \) to \( \Phi \approx 0.36 \). Above the biotite-out boundary, melt fractions climb steeply to values of \( \sim 0.56 \) and \( \sim 0.34 \) at \(1000^\circ C\) for the CSW and SWL paths, respectively \((dm/dT \sim 3–4 \text{ mol}\%/10^\circ C)\).

The SWL scenario for composition PD97T results in decreased melt fertility and slight increases in the modal proportions of quartz, plagioclase and K-feldspar when compared to the CSW path (Fig. 4g & 4h). A small, moderate increase in melt fraction, from 0.0 to \( \sim 0.02 \), occurs just above the solidus, at \( \sim 640^\circ C\), along the CSW path \((dm/dT \sim 5 \text{ mol}\%/10^\circ C)\). Again, this initial step in melt fraction is absent from the SWL path. The melt fraction along the CSW path increases gently to 0.09 over the interval \( \sim 645–840^\circ C \) \((dm/dT \sim 0.1–0.2 \text{ mol}\%/10^\circ C)\) and then abruptly increases to 0.17 between \( \sim 840–850^\circ C \) \((dm/dT \sim 10 \text{ mol}\%/10^\circ C)\). A similar sharp increase in melt fraction occurs along the SWL path (from \( \Phi \sim 0.03 \) to 0.10) over the same temperature interval. For both paths this step corresponds to the first appearance of orthopyroxene, the breakdown and disappearance of biotite and an increase in the modal abundance of K-feldspar. Above the biotite-out boundary melt fractions increase moderately steeply to reach \( \sim 0.45 \) and \( \sim 0.36 \) at \( 1000^\circ C \) for the CSW and SWL paths, respectively \((dm/dT \sim 10 \text{ mol}\%/10^\circ C)\). Interestingly, hornblende modes do not noticeably decrease for either path over the calculated melting interval \((\sim 640–1100^\circ C)\).
Melting at low pressure

The minimum saturation point for each composition, and the corresponding SWL melting path, increases as a function of pressure. Therefore, it is not possible to create a single $P-T$ pseudosection for a bulk composition affected by SWL. Rather it is necessary to determine $T-M_{H_2O}$ phase relationships and $H_2O$ value of the minimum saturation point at each pressure of interest. As a result this method is only applicable to the consideration of isobaric melting paths. To examine the effect of pressure on the phase equilibria, temperature-mode diagrams were generated for the SWL scenario along prograde isobaric heating paths for VH88P, G95P and VM94G at 5 kbar and PD97T at 4 kbar (Fig. 5).

At 5 kbar, VH88P (SWL) produces very little melt ($\Phi < 0.03$) between the solidus, at $\sim 690^\circ C$, and $\sim 775^\circ C$ ($dm/dT \ll 0.1 \text{ mol}\% / 10^\circ C$). The melt fraction then rises sharply to $\Phi \approx 0.07$ at $\sim 780^\circ C$ ($dm/dT \sim 10-20 \text{ mol}\% / 10^\circ C$). This step corresponds to the disappearance of sillimanite from the equilibrium assemblage, a sharp decrease in the modal abundance of biotite, an increase in the modal abundance of cordierite and the first appearance of garnet. This feature is absent from the SWL scenario for composition VH88P at 10 kbar. Above $\sim 780^\circ C$ the melt fraction increases moderately steeply reaching $\Phi \approx 0.67$ at $\sim 1000^\circ C$ ($dm/dT \sim 2-3 \text{ mol}\% / 10^\circ C$; c.f. 0.44 for VH88P at 10 kbar).

The melt fraction for composition G95P (SWL) at 5 kbar remains low ($< 0.01$) between the solidus ($\sim 655^\circ C$) and $\sim 675^\circ C$ ($dm/dT \ll 0.1 \text{ mol}\% / 10^\circ C$). Between 675–680$^\circ C$ the melt fraction jumps abruptly to $\sim 0.07$ ($dm/dT \sim 10-20 \text{ mol}\% / 10^\circ C$). This step corresponds to the disappearance of muscovite and the first appearance of sillimanite in the equilibrium assemblage. A similar high $dm/dT$ step is observed at $\sim 760^\circ C$ in the SWL melting curve for G95P at 10 kbar. From 680–835$^\circ C$ the melt fraction increases gradually
to 0.19 ($dm/dT \sim 0.5\text{--}0.7 \text{ mol}\%/10^{\circ}\text{C}$). From 835$^\circ\text{C}$ the melting curve begins to increase steeply corresponding to the breakdown and disappearance of biotite and the first appearance of orthopyroxene in the equilibrium assemblage ($dm/dT \sim 8 \text{ mol}\%/10^{\circ}\text{C}$).

Melting then proceeds with a steep slope reaching 0.70 at $\sim 1000^{\circ}\text{C}$ ($dm/dT \sim 4\text{--}5 \text{ mol}\%/10^{\circ}\text{C}$; c.f. 0.51 for G95P at 10 kbar).

At 5 kbar, the melting curve for composition VM94G (SWL) increases slightly, from $\Phi = 0.0$ to 0.02, between the solidus ($675^{\circ}\text{C}$) and the first appearance of garnet in the equilibrium assemblage at 785$^\circ\text{C}$ ($dm/dT \ll 0.1 \text{ mol}\%/10^{\circ}\text{C}$). Between 785–840$^\circ\text{C}$ the melt fraction increases sharply to 0.22 corresponding to the first appearance of orthopyroxene and the breakdown and disappearance of biotite from the equilibrium assemblage ($dm/dT \sim 8\text{--}10 \text{ mol}\%/10^{\circ}\text{C}$). Above 840$^\circ\text{C}$ the melting curve rises steeply to reach 0.60 at $\sim 1000^{\circ}\text{C}$ ($dm/dT \sim 4 \text{ mol}\%/10^{\circ}\text{C}$; c.f. 0.34 for G95P at 10 kbar).

At 4 kbar PC97T (SWL) produces very little melt ($< 0.02$) between the solidus, at $\sim 655^{\circ}\text{C}$, and 795$^\circ\text{C}$ corresponding to the first appearance of orthopyroxene ($dm/dT \ll 0.1 \text{ mol}\%/10^{\circ}\text{C}$). Between 795–805$^\circ\text{C}$ the melt fraction increases sharply to 0.11 ($dm/dT \sim 10\text{--}15 \text{ mol}\%/10^{\circ}\text{C}$). This interval corresponds to the rapid breakdown and disappearance of biotite and an increase in the abundance of K-feldspar and orthopyroxene in the equilibrium assemblage. Above 805$^\circ\text{C}$, the melt fraction increases moderately steeply to reach 0.47 at $\sim 1000^{\circ}\text{C}$ (c.f. 0.36 for G95P at 10 kbar).
Comparisons with melting experiments

The experimentally derived melt fractions corresponding to VH88P, G95P, VM94G and PD97T are shown in Figs. 3 & 6. As previously noted, the starting H$_2$O contents for the melting experiments reported by Patino-Douce (1997) are not known, so it is not possible to make a comparison with the modelled CSW melting path for PD97T. Nevertheless, it is still useful to compare an inferred CSW path from matching the reported experimental melt fraction with the modelled SWL scenario.

Vielzeuf & Holloway (1988) infer a very steep increase in melt fraction, from $\Phi = 0.10$ to 0.60 over the temperature range 850–900°C corresponding to a biotite-out “reaction” (i.e. bi + sill/ky + pl + q $\rightarrow$ liq + g + ksp; Fig. 6a). Above 900°C, these authors describe a melting ‘plateau’ corresponding to the gradual breakdown of sillimanite and garnet. Neither of these features are observed for the CSW path for VH88P which displays broadly monotonic melting behaviour above 800°C. At temperatures higher than $\sim$950°C, they report lower melt fractions than the VH88P() melting curve. A similar misfit is observed between the the experimental results of Vielzeuf & Montel (1994) and the modelled CSW path for VM94 at 1000°C (Fig. 6c). Experiments above 950°C were not reported in Gardien et al. (1995) or Patino-Douce (1997), so any high $T$ effect cannot be assessed for studies. Nevertheless, the generation of lower than predicted melt fractions at high $T$ is consistent with observations from other experiments that melt modes often decrease at temperatures above 950°C (possibly due to diffusive H$_2$O-loss through the wall of the experimental capsule; Patino-Douce & Beard, 1994).

The experimental melt fraction estimates reported by Gardien et al. (1995) show a
general trend of increasing melting with increasing $T$ that broadly corresponds to the CSW melting curve for composition G95P (Fig. 6b). Interestingly, both the CSW and SWL melting curves for composition G95P record a steep increase in melt fraction at $\sim 760^\circ$C, corresponding to the mu-out boundary with increasing $T$, while the reported experimental melt fraction actually decreases by 5% between 750°C and 800°C before increasing at higher $T$. The melting estimates of Vielzeuf & Montel (1994) are slightly above the corresponding VM94G() melting curve at temperatures below $\sim 900^\circ$C (Fig. 6b).

Importantly, the experimentally-derived melt fractions for all starting compositions are higher than the corresponding SWL melting curves as a function of $T$. This general result is consistent with the assumption that SWL constrained melting represents a minimum melting end-member for a particular bulk composition.

The general advantages and limitations of melting experiments and phase equilibria modelling have been discussed in detail elsewhere (e.g. Johnson et al., 2008; White et al., 2011), so the following discussion focusses on the issues that are specific to estimating crustal melt fertility. The differences between the experimental estimates and the modelled CSW melting curves can be understood in terms of the different limitations inherent for each approach.

A key assumption of most melting experiments is that the analysed composition of the starting material is representative of the equilibrating bulk composition at experimental run conditions. As previously discussed above, the combined effects of $\text{H}_2\text{O}$ absorption into powders during handling as well as diffusive $\text{H}_2\text{O}$ loss through the experimental casing at high temperature may result in significant changes in the actual $\text{H}_2\text{O}$ content of a given melting experiment. The CSW scenarios reported here represent idealised melting
experiments that are unaffected by these undesirable experimental phenomena.

Detailed melting experiments are labour and time intensive with individual experiments
often taking several weeks to approach equilibration (e.g. Patino-Douce, 2004). As a result,
the total number of data points generated is usually low and performing repeat experiments
at the same $P$–$T$ conditions is uncommon making it difficult to assess the statistical
significance or the reproducibility of any single melting estimate. By contrast, phase
equilibria modelling allows the determination of melt modes over a range of $P$–$T$–$X$
conditions for which the constituent phase models are calibrated. This allows for the
calculation of continuous melting curves along a particular $P$–$T$ path of interest.

Melt fractions for each experimental charge are typically estimated either by mass
balance calculations or by graphical assessment of the proportion of melt in
two-dimensional back-scattered electron scans. Both methods have high analytical
uncertainties that may be difficult to quantify (e.g. Vielzeuf & Montel, 1994). Quantitative
determination of melt fractions from experimental data is particularly problematic at low
melt fractions due to the small size of melt/glass volumes that may be observed and
analysed within the matrix of the experimental charge. As a result, experimental studies
can often only approximate the degree of melting occurring at temperatures near the
solidus. This constraint does not apply to phase equilibria modelling where the degree of
melting can be calculated in detail from the solidus up to high temperatures.

The NCKFMASH melt model used in this study is constrained by well calibrated
experimental data on unary, binary and ternary subsystems (Holland & Powell, 2001). It
can, therefore, expected to satisfactorily describe the topology of the melt bearing fields in
NCKFMASHTO, if not, for example, the precise $PT$ position of fields. As previously
discussed, the effects of TiO$_2$ and Fe$^{3+}$ on the modelled melt fertility are likely to be small and certainly less than the uncertainties inherent in the determination of experimental estimates on the corresponding natural rock compositions. The influence of other minor components (e.g. Mn, F, Cl, P, B) on melting behaviour cannot be assessed with the models used here and in any case it is unlikely that any of these components (with the possible exception of F) will significantly affect melting behaviour at the low concentrations reported for the experimental starting compositions (shown in Table 1). We, therefore, argue that the modelled melting curves presented here represent reliable estimates of the melt fertility of the starting compositions under conditions of CSW and SWL.

**Estimates of crustal melt fertility**

It is clear that the overall ‘shapes’ of the melting curves for the CSW and SWL scenarios for each starting composition are similar and reflect similar underlying melt-producing equilibria (Fig. 6). Inflections in the melting curves occur at similar $T$ indicating that they are largely independent of bulk H$_2$O content over the ranges modelled. The key difference between the CSW and SWL curves, in each case, is that the CSW melting curve contains an initial steep increase, just above the solidus, corresponding to the disappearance of free H$_2$O from the equilibrium assemblage. This initial melting step effectively increases the baseline for subsequent melting but does not significantly affect the shape of the curve at higher $T$. The most obvious observation that can be made from comparing these curves is that CSW paths that incorporate excess free H$_2$O will yield more melt than the corresponding SWL path.

The SWL melting curves for each starting composition at high and low pressure are
shown in Figure 7. With decreasing pressure the solidus migrates to higher $T$ (shown by
the grey and black arrows in Fig. 7), while at the same time the H$_2$O solubility of the melt
phase increases. At low pressure, SWL constrained heating paths intersect key melting
equilibria at lower $T$ compared to the corresponding high $P$ paths (cf. SWL versus CSW
paths at constant $P$; Fig. 6). This is because the underlying fluid-absent melting reactions
have a positive slope in $P$–$T$ space. The resulting melting curves show increased melt
generation in the lower pressure scenarios for all starting compositions. This is likely to be a
general result that is applicable to most felsic rock compositions.

As noted above there is no obvious geological mechanism that would cause the bulk
composition to contain less H$_2$O than the value corresponding to the minimum saturation
point. As a result the SWL path results in the minimum melt fertility scenario for a given
source rock. Increasing melt fertility above the minimum defined by efficient SWL may be
achieved via the retention of sub-solidus H$_2$O or the addition of H$_2$O to the system at
temperatures above the solidus (compare paths A, B & C in Fig. 8).

H$_2$O may be released from a source rock with a higher $T$ solidus into an adjacent rock
with a lower $T$ solidus (White et al., 2005). For example, at 10 kbar composition VH88P
intersects its wet-solidus at $\sim$680°C whereas the solidus for composition G95P occurs at
$\sim$640°C. If these two compositions represent hypothetical adjacent source regions subjected
to the same heating path, VH88P could provide up to 3% H$_2$O to G95P over the
temperature interval between their respective solidi. assuming the source rocks were in
equal proportions, this would result in a significant increase in the melt fertility of the G95P
source rock ($\sim$10% more melt at 800°C, $\sim$15% more melt at 1000°C). Water may also be
added to a particular source rock via dewatering of migrating melts. During cooling and
crystallisation water is released as the melt composition intersects the H$_2$O saturated fields
near the solidus (e.g. paths D & E in Fig. 8). H$_2$O saturation in melt is strongly dependent on pressure (Tuttle & Bowen, 1958; Johannes & Holtz, 1996). Therefore, the volume of water released by a crystallising melt will depend on the initial H$_2$O content of the melt and the pressure at which the melt intersects the H$_2$O-in boundary. These three mechanisms may result in greater melt fertility than is defined by a SWL-moderated minimum melting curve. However, under most geological conditions SWL paths and melt fertility are likely to apply.

The modelling presented here does not take into consideration the effect of melt loss on melt fertility of felsic rocks as this process involves modification of several compositional vectors the cannot be represented on a simple $T$–$M$$_{H_2O}$ diagram. The loss of melt will cause the bulk composition to become more residual with the largest compositional change likely to be a decrease in water content due to the hydrous nature of the melt derived from melting of felsic rocks (White & Powell, 2002; 2010).

**DISCUSSION and CONCLUSIONS**

**General conclusions**

The water content of the rock comprises the sum of free H$_2$O and structurally-bound H$_2$O residing in hydrous minerals (e.g. biotite, muscovite and hornblende). While free H$_2$O can be lost from a source rock (due to compaction, deformation, migrating porosity, etc.), structurally-bound H$_2$O is only made mobile when the hydrous host mineral becomes unstable due to changes in the $P$–$T$–$X$ conditions at equilibrium. This may occur at sub solidus and/or suprasolidus conditions. Above the solidus, any H$_2$O liberated by the breakdown of hydrous phases is incorporated into the melt phase (until it is H$_2$O
The $\text{H}_2\text{O}$ contents of the systems described here cannot evolve to be less than the value of the minimum saturation point without also losing melt and/or solid phases (Fig. 8). In this way subsolidus water loss can be viewed as defining the minimum $\text{H}_2\text{O}$ condition for a particular bulk composition that has undergone prograde heating. A corollary of this is that the subsolidus water loss path defines the minimum melt fertility of a rock, at least prior to melt loss.

Most melting experiments are likely to conserve any $\text{H}_2\text{O}$ generated by devolatilisation of hydrous minerals as the experimental charge is heated and pressurised from ambient temperatures up to the conditions of the experimental run. Given the likelihood that subsolidus water loss is a significant phenomena affecting prograde metamorphic rocks, most experimentally derived melting curves will tend to overestimate the melt fertility of the starting material at the onset of melting within the continental crust. The modelled curves presented here allow subsolidus water loss to be taken into account and, therefore, allow determination of the minimum melting relationships for compositions of interest.

Implications for melt connectivity, melt loss and the rheology of the crust during anatexis

An important consequence of the reduced melt fertility implied by the subsolidus water loss model is that critical melt thresholds will be intersected at significantly higher temperatures (Fig. 6). At low-melt fractions melt will initially form in isolated domains along grain edges and corners between reactant phases (Melnert et al., 1973). As the melt fraction increases these isolated melt volumes begin to coalesce eventually forming an interconnected network through the rock volume (Vigneresse et al., 1996). Rosenberg & Handy (2005) identified a
melt connectivity threshold which occurs at $\Phi \approx 0.07$ and corresponds to $\sim 80\%$ of grain boundaries containing melt. Experimental data suggests that rocks experience significant loss of strength between the onset of melting and the melt connectivity threshold (from $\sim 800$ MPa down to $\sim 200$ MPa; Rosenberg & Handy, 2005). It follows that rocks having experienced subsolidus water loss will remain comparatively stronger at higher temperatures than rocks that have not. For the bulk compositions modelled here the differences in the temperature of the melt connectivity threshold range from $<10^\circ C$ to $>150^\circ C$. Such a relationship between degree of melting, temperature and strength is likely to significantly influence the rheological behaviour of crust undergoing regional metamorphism and anatexis.

Higher degrees of melting are associated with critical thresholds governing melt segregation and the transition from a solid to a liquid rheology. The melt escape threshold refers to the melt fraction at which melt is able to segregate from the source (Vigneresse et al., 1996). The melt escape threshold is likely to depend on a number of factors including viscosity and strain rate but is likely to occur at $\Phi \sim 0.2–0.25$ (Sawyer, 1994; Vigneresse et al., 1996). The ‘solid-to-liquid transition’ is the melt content at the transition from a solid-supported to a liquid-supported structure (Rosenberg & Handy, 2005). The solid-to-liquid transition occurs at $\Phi \sim 0.4–0.6$. Both the melt escape threshold and the solid-to-liquid transition will be shifted to higher temperatures for a rock that has experienced subsolidus water loss.
Subsolidus water loss during magma crystallisation

The discussion so far has concerned the consequences of subsolidus water loss during prograde heating, however crystallising magmas may also experience subsolidus water loss. Large amounts of H$_2$O may be generated as a magma approaches and then crosses the solidus if a magma had an initially high H$_2$O content. The magma is rapidly converted to solids + H$_2$O across the narrow melt + H$_2$O fields just above the wet solidus (Paths D & E in Fig. 8). Even at upper crustal conditions the inter-grain porosity is likely to be small and the tendency for free H$_2$O to escape the now solid rock volume is likely to be high (Connolly, 1997). The escape of free H$_2$O will cause the bulk H$_2$O contents to decrease towards the minimum saturation point. Further cooling will take the rock across the H$_2$O-out boundary and the H$_2$O contents will become fixed. An important consequence of this is that the final H$_2$O contents of the crystallised solid (e.g. a tonalite) will not reflect its H$_2$O content when it was a magma but rather reflects the water contents in the vicinity of the minimum saturation point. This would mean that analysing the H$_2$O contents of granites collected in the field may tell us nothing about the H$_2$O contents of the original magma.

ACKNOWLEDGEMENTS

RP acknowledges support from Australian Research Council DP0987731. SM acknowledges support from the Australian Research Council DP0987765.
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Fe2O3 on metapelitic assemblages at greenschist and amphibolite facies conditions:
mineral equilibria calculations in the system


Table 1. Starting compositions used in this study in wt.% and equivalent NCKFMASHTO mol.% values. The compositions are and the corresponding experimental studies are as follows: VH88P - high-Al metapelite from Vielzeuf & Holloway (1988), G95P - low-Al metapelite from Gardien et al. (1995), VM94G - metagreywacke from Vielzeuf & Montel (1994), and PD97T - metatonalite from Patino-Douce (1997).

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Table 2. Compositions used to construct pseudosections (in mol.%).

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FIGURE CAPTIONS

Fig. 1 Harker plots of mol.% oxides, molar K$_2$O / Na$_2$O and ASI (molar Al$_2$O$_3$/ (CaO+Na$_2$O+K$_2$O)) for the starting compositions: VH88P - metapelite, G95P - metapelite, VM94G - metagreywacke and PD97T - metatonalite. L, M and U represent the estimated average compositions of the lower-, middle- and upper-crust, respectively, of Rudnick & Gao (2003). All values are plotted on an anhydrous basis.

Fig. 2 Calculated NCKFMASHTO $T$–$M_{H_2O}$ pseudosections for the compositions: (a) VH88P - metapelite, (b) G95P - metapelite and (c) VM94G - metagreywacke at 10 kbar and (d) PD97T - metatonalite at 8 kbar (see Table 2). The dashed line indicates the solidus for each diagram. The white dot indicates the locus of the minimum saturation point for each diagram.

Fig. 3 Calculated contours for the modal abundances of melt (solid lines) and free H$_2$O (dashed lines) for the compositions: (a) VH88P - metapelite, (b) G95P - metapelite, (c) VM94G - metagreywacke and (d) PD97T - metatonalite (see Table 2). The dark grey areas represent the regions over which free-water is stable. The grey arrows represent the path followed for heating in a closed system with the H$_2$O contents fixed at the value reported for the experimental starting composition. The H$_2$O contents for the tonalitic bulk composition, (d), is unknown so the H$_2$O contents were determined so that 30% melt is generated at 950 °C as reported in Patino-Douce (1997). As with Fig. 2 the white dot indicates the locus of the minimum saturation point for each diagram. The black arrows represent heating combined with subsolidus water loss. The stars represent melt fraction estimates with increasing temperature derived from the corresponding melting experiments (assuming
uncertainty in the H₂O contents only; see text for discussion).

Fig 4 Calculated molecular proportions of phases \((X)\) with increasing temperature for the four compositions: VH88P, G95P and VM94G at 10 kbar and PD97T at 8 kbar under conditions of conservation of water (C) and sub-solidus water loss (SWL). The dashed vertical line indicates the solidus for each diagram.

Fig 5 Calculated molecular proportions of phases \((X)\) with increasing temperature for the four compositions: VH88P, G95P and VM94G at 5 kbar and PD97T at 4 kbar under conditions of sub-solidus water loss (SWL). The solidus is at the left edge for each diagram.

Fig 6 Comparison of the CSW and SWL melt fraction curves (shown in figure 5) with the available experimental estimates for the four compositions VH88P, G95P and VM94G at 10 kbar and PD97T at 8 kbar. The horizontal dashed lines indicate the melt fractions corresponding to the solid-liquid transition (SLT) and melt connectivity threshold (MCT; both after Rosenberg & Handy, 2005) and the melt escape threshold (MET; after Vigneresse et al., 1996). The arrows indicate the difference in the temperature of the MCT between the CSW and SWL scenarios (see text for discussion).

Fig 7 Comparison of the melt fraction curves with increasing temperature at low \(P\) (light grey curve) and high \(P\) (dark grey curve) for the four compositions VH88P, G95P and VM94G (5 kbar and 10 kbar) and PD97T at (4 kbar and 8 kbar). The respective solidi at each pressure are indicated by the corresponding grey arrows. The position of the MCT is shown by the horizontal dashed line.
Fig 8 Schematic diagram showing the relationships between various melting paths (dark grey dotted lines) and crystallisation paths (light grey dotted lines) as a function of water contents. Thin grey lines represent melt contours (increasing to the top-right of the diagram). The striped area is inaccessible because H\textsubscript{2}O contents less than this value are unobtainable without invoking changes in other compositional vectors (e.g. via melt and or solid segregation; see text for discussion).
Figure 1.
Figure 2.
Figure 3.
Figure 4.
Figure 5.
Figure 6.
Figure 7.
Figure 8.
Author/s:
Webb, G; Powell, R; McLaren, S

Title:
Phase equilibria constraints on the melt fertility of crustal rocks: the effect of subsolidus water loss

Date:
2015-02-01

Citation:

Persistent Link:
http://hdl.handle.net/11343/112394