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doi:10.1016/j.gca.2005.02.035

Coupling of anatectic reactions and dissolution of accessory phases and the Sr and Nd isotope systematics of anatectic melts from a metasedimentary source

LINGSSEN ZENG,* PAUL D. ASIMOW, and JASON B. SALEEBY

Division of Geological and Planetary Sciences, California Institute of Technology, Pasadena, CA 91125, USA

(Received September 22, 2004; accepted in revised form February 25, 2005)

Abstract—Advances in field observations and experimental petrology on anatectic products have motivated us to investigate the geochemical consequences of accessory mineral dissolution and nonmodal partial melting processes. Incorporation of apatite and monazite dissolution into a muscovite dehydration melting model allows us to examine the coupling of the Rb-Sr and Sm-Nd isotope systems in anatectic melts from a muscovite-bearing metasedimentary source. Modeling results show that (1) the Sm/Nd ratios and Nd isotopic compositions of the melts depend on the amount of apatite and monazite dissolved into the melt, and (2) the relative proportion of micas (muscovite and biotite) and feldspars (plagioclase and K-feldspar) that enter the melt is a key parameter determining the Rb/Sr and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the melt. Furthermore, these two factors are not, in practice, independent. In general, nonmodal partial melting of a pelitic source results in melts following one of two paths in $\epsilon_{\text{Nd}}\text{-}^{87}\text{Sr}/^{86}\text{Sr}$ ratio space. A higher temperature, fluid-absent path (Path 1) represents those partial melting reactions in which muscovite/biotite dehydration and apatite but not monazite dissolution play a significant role; the melt will have elevated Rb/Sr, $^{87}\text{Sr}/^{86}\text{Sr}$, Sm/Nd, and ϵ_{Nd} values. In contrast, a lower temperature, fluid-fluxed path (Path 2) represents those partial melting reactions in which muscovite/biotite dehydration plays an insignificant role and apatite but not monazite stays in the residue; the melt will have lower Rb/Sr, $^{87}\text{Sr}/^{86}\text{Sr}$, Sm/Nd, and ϵ_{Nd} values than its source. The master variables controlling both accessory phase dissolution (and hence the Sm-Nd system), and melting reaction (and hence the Rb-Sr systematics) are temperature and water content. The complexity in Sr-Nd isotope systematics in metasediment-derived melts, as suggested in this study, will help us to better understand the petrogenesis for those granitic plutons that have a significant crustal source component. Copyright © 2005 Elsevier Ltd

1. INTRODUCTION

Crustal anatexis occurs in various tectonic environments in which either: (1) crustal temperature is elevated by intrusion of hot magmas from depth (Bergantz, 1989); (2) the solidus is lowered by addition of water (Harris et al., 1993); or (3) decompression during rapid exhumation of deep seated crustal rocks elevates the geotherm (Harris and Massey, 1994). An important mechanism for crustal anatexis is dehydration partial melting of hydrous-mineral-rich metasedimentary rocks at various crustal levels (Wyllie, 1977; Le Breton and Thompson, 1988; Vielzeuf and Holloway, 1988; Rushmer, 1991; Harris and Inger, 1992; Gardien et al., 1995; Patiño-Douce, 1996; Thompson, 1996; Knesel and Davidson, 2002).

Crustally derived melts have complicated trace element and isotopic geochemistry in part due to the disequilibrium nature of crustal anatexis at relatively low temperatures. Experimental and field observations have shown that the geochemical and isotopic characteristics of anatectic melts are determined by mineral phases entering the melt, rather than the bulk source rock (e.g., Hammouda et al., 1996; Tommasini and Davies, 1997; Knesel and Davidson, 1999, 2002). That is, as diffusion through the residual minerals becomes slow relative to melt extraction, a limit is approached whereby the phases contributing to the nonmodal melting reaction simply impart their

budget of trace elements and their isotopic signature directly to the melt, whereas material that does not melt quantitatively withholds its trace elements. Partial melting of mica-rich metasedimentary rocks is particularly interesting, because it has been demonstrated that a given source material can generate a wide spectrum of melts and crystalline residues with a wide range of trace element and radiogenic isotopic signatures (Patiño-Douce, 1996; Patiño-Douce and Harris, 1998; Knesel and Davidson, 2002). It follows that contamination of an intruding granitic magma body by partial melts of the crust, for example, may not admit a straightforward interpretation in terms of simple mixing of components.

Over the last two decades, experimental petrology and field observations have suggested that the behavior of trace elements hosted dominantly by rock-forming minerals may be quite different from those hosted mainly by accessory minerals such as apatite and monazite (Miller and Mittlefehldt, 1982; Mittlefehldt and Miller, 1983; Watson and Harrison, 1984; Hogan and Sinha, 1991; Watt and Harley, 1993; Nabelek and Glascock, 1995; Nabelek, 1999; Bea, 1996a; Bea and Montero, 1999; Zeng et al., 2005). In addition, the behavior of light rare earth elements (LREE), particularly Sm and Nd, in felsic magma is very different from that which has been inferred for mafic magmas (Miller and Mittlefehldt, 1982; Mittlefehldt and Miller, 1983). The participation of phases in which LREE are highly compatible has important implications for trace element and Nd isotopic systematics of the products of crustal anatexis. For instance, Watson and Harrison (1984) pointed out the possibility that unmelted accessory minerals may carry with them the radiogenic components (Pb, Nd and Hf) of the source

* Author to whom correspondence should be addressed, at Key Laboratory for Continental Dynamics of MLR, Institute of Geology, Chinese Academy of Geological Sciences, Beijing, 100037, China (lzeng@ccsd.org.cn).

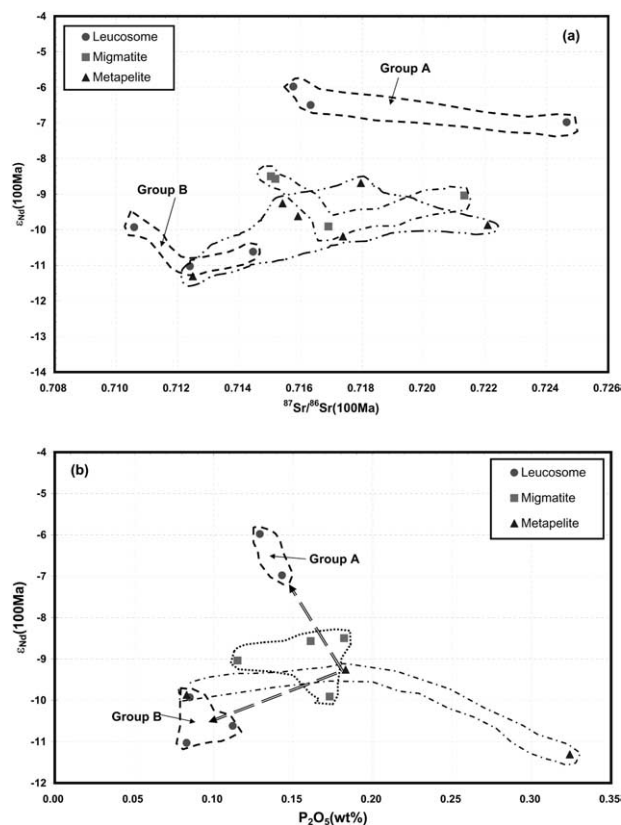


Fig. 1. (a) Nd and Sr isotopic covariation diagram showing the Sr-Nd systematics of the leucosomes, migmatites, and metapelites from the Goat Ranch migmatite complex, Southern Sierra Nevada. The leucosomes can be subdivided into two groups. Group A leucosomes have higher $\epsilon_{Nd}(100\text{ Ma})$ values than those of either metapelites or migmatites, suggesting a Nd disequilibrium partial melting. (b) Plot of $\epsilon_{Nd}(T)$ vs. P_2O_5 contents in the leucosomes, metapelites and migmatites from the Goat Ranch migmatite complex, Southern Sierra Nevada. Data are from Zeng (2003).

region. Hogan and Sinha (1991) presented a model and discussed the role of zircon in the redistribution of lead isotopes during crustal anatexis. If the differences in Sr or Nd isotopic compositions among phases in an aged source rock persist until the onset of melting, it is conceivable that melts from crustal partial melting will develop complicated but distinct isotopic compositions with respect to their source. Whether or not the Rb-Sr system is coupled with the Sm-Nd system during partial melting of typical metasedimentary rocks is still not a well-understood problem. Because most granitoids contain some crustal components derived from assimilation of melts from crustal rocks, understanding the joint behavior of the Sm-Nd and Rb-Sr isotope systems during anatexis of metasediments is critical to discuss the Sr-Nd systematics of granitoids and to formulate and test various petrogenesis models for their formation.

In our studies on the migmatite zone of the Isabella pendant, southern Sierra Nevada, California, we found that the leucosomes, i.e., melts derived from partial melting of the pelites of the metamorphic framework rocks, preserve a record of large magnitude Nd but weak Sr isotopic disequilibrium with respect to their pelitic source (Fig. 1). The leucosomes have a wide

range of initial $\epsilon_{Nd}(i = 100\text{ Ma})$ from -6.0 to -11.0 , as compared to -8.7 to -11.3 for the source. These leucosomes can be subdivided into two groups based on their major elements and Sr and Nd isotope geochemistry. Group A leucosomes have higher P_2O_5 contents and $\epsilon_{Nd}(i)$ values, a tendency toward high $^{87}\text{Sr}/^{86}\text{Sr}$, and granitic composition. Group B leucosomes have lower P_2O_5 , lower $\epsilon_{Nd}(i)$, lower $^{87}\text{Sr}/^{86}\text{Sr}$, and trondhjemitic-like composition (Fig. 1). The $\epsilon_{Nd}(i)$ values of Group A leucosomes are significantly higher, by 2–4 epsilon units, than those of bulk migmatites and of correlative metapelites further from the Goat Ranch granodiorite pluton, suggesting substantial Nd isotopic disequilibrium with respect to their sources. We seek here to examine whether the coupling between accessory mineral dissolution and type of melting reaction can account for the correlation between the observed Sr isotope, Nd isotope, P_2O_5 , and trondhjemitic vs. granitic systematics.

In this study, we intend to take partial melting of muscovite-bearing metasedimentary rocks as an example to explore the effects of accessory mineral dissolution and nonmodal partial melting on the Rb-Sr and Sm-Nd isotopic systematics of meta-sediment-derived melts. We will limit ourselves to the well-constrained case of muscovite dehydration melting of a hypothetical source system consisting of the rock-forming minerals quartz, plagioclase, muscovite, biotite, and aluminosilicate with accessory minerals apatite and monazite. We will demonstrate how Rb, Sr, Sm, Nd, $^{87}\text{Sr}/^{86}\text{Sr}$, ϵ_{Nd} , P_2O_5 , and K/Na behave during melting of this system. The results presented in this study can be readily extended to systems with more accessory minerals and more complicated melting reactions. Although we address only Sr and Nd isotopes, our results have implications for understanding the behavior of other radiogenic isotope systems such as U-Pb and Lu-Hf during crustal anatexis.

2. MELTING REACTIONS AND PROCESSES

2.1. Melting Reactions

A number of experiments have determined the melting reactions of metapelites and metabasites. These results show that crustal anatexis in general can be classified as either fluid-present or fluid-absent partial melting, which differ in the importance of breakdown of hydrous minerals such as muscovite, biotite and hornblende. Patiño-Douce and Harris (1998) conducted partial melting experiments on metapelitic rocks from the High Himalayan Crystalline Sequence that are likely sources of the Miocene Himalaya leucogranites. In their experiments on a muscovite schist at pressures of 0.6, 0.8 and 1.0 GPa and temperatures of 700 to 900°C, both without added H_2O (dehydration-melting) and with 1–4 wt% added H_2O (H_2O -fluxed melting), they found melting reactions described by muscovite dehydration melting,

Muscovite dehydration melting



and H_2O -fluxed melting,



where x is the stoichiometric coefficient for H_2O added to the experiment and Ms, Pl, Qtz, Kfs, Sil and Bt represent muscovite, plagioclase, quartz, K-feldspar, sillimanite, and biotite,

respectively (mineral abbreviations are after Kretz, 1983). The stoichiometry of melting reaction MS1 was determined by the difference between the phase abundances at conditions that just exceed the muscovite-out boundary and those in the starting material (Patiño-Douce and Harris, 1998). In contrast, the stoichiometry of reaction MS2 was inferred from the difference in phase abundances between the experiments at 6 kbar, 750°C with 2 wt% added H₂O and the starting material. At 6 kbar, dehydration-melting begins at 750°C, and produces melts that are virtually identical in composition to the Himalayan leucogranites. Adding H₂O lowers the solidus by promoting plagioclase + quartz melting. Melts produced from these starting materials at T < 750°C by H₂O-fluxed melting are trondhjemitic, and different in composition from most Himalayan leucogranites.

In this study, we only evaluate the two reactions (MS1 and MS2) determined by Patiño-Douce and Harris (1998) for muscovite-bearing metasedimentary rocks. We did not examine the biotite dehydration partial melting reactions in that they share similar features to those of muscovite reactions, especially for the Rb-Sr isotope system. For a comprehensive review of experimentally determined partial melting reactions of metasedimentary rocks, please refer to Whittington and Treloar (2002) and references therein. We choose these particular partial melting reactions to evaluate the behavior of the Rb-Sr and Sm-Nd isotopic systems during melting because they provided detailed accounts of the residual mode and the melting reactions as functions of melt fraction. These data are essential for us to model the behavior of Rb, Sr, Sm, and Nd during crustal anatexis involving muscovite.

2.2. Dissolution of Accessory Minerals

Accessory minerals are the predominant hosts of U, Th and rare-earth elements (REE) in silicic rocks (Gromet and Silver, 1983; Bea, 1996b). In the majority of crustal rocks, four accessory minerals—apatite, zircon, monazite and xenotime—usually contain more than 80%–90% of total REE, Zr, Y, U, and Th. Therefore, understanding of the behavior of accessory phases during crustal anatexis is critical to evaluating the magnitude of Sm-Nd parent-daughter fractionation and Nd isotopic compositions in the metasediment-derived melts. The behavior of accessory minerals during high-grade metamorphism and crustal anatexis depends primarily on three factors: (1) solubility and dissolution kinetics into the melt; (2) solidus stability and involvement in metamorphic reactions; and (3) textural position and hence, availability for both the melt and other solid phases that react with them (Watson et al., 1989; Watt and Harley, 1993).

A series of experiments (Harrison and Watson, 1983, 1984; Watson and Harrison, 1983, 1984; Rapp and Watson, 1986; Pichavant et al., 1992; Montel, 1993; Wolf and London, 1994, 1995) have investigated dissolution of important accessory phases such as apatite, zircon, monazite and rutile. These results provide a basis from which the effects of accessory phases on the fractionation of LREEs, particularly Sm from Nd, can be understood. Unfortunately, almost all the experimentally determined partial melting reactions have focused on rock-forming minerals, and do not report the behavior of key accessory phases, even though a number of studies have indicated

the importance of accessory minerals in controlling the LREE, trace and some major elements (e.g., phosphorus and titanium) of crustally derived melts (Sawyer, 1991; Sevigny, 1993; Watt and Harley, 1993; Nabelek and Glascock, 1995; Wolf and London, 1995; Watt et al., 1996). Hence we must attempt to combine information from separate sets of experiments on major-mineral melting reactions and accessory mineral solubility and dissolution kinetics; if we can identify common variables that affect both systems, we can predict the coupling of these processes even though few experimental studies have examined them simultaneously. Before we proceed to discuss the effect of accessory phases on the Sm-Nd isotope systematics of the melts, we will briefly review factors that affect the dissolution of accessory phases during crustal anatexis.

Since metapelite has very low modal abundance of zircon and zircon has very low Sm and Nd concentrations (a few ppm at most) (Rubatto, 2002), the effect of zircon in Sm and Nd isotope compositions of the melts is minor. Therefore, we will not discuss its role in fractionation of Sm over Nd during anatexis of a pelitic source further in the following sections.

2.2.1. Apatite Solubility

Phosphorus is an essential structural component of apatite, thus the concentration of phosphorus in a granitic melt is buffered by apatite solubility. Apatite saturation levels are a function of temperature and composition of the melt, but are independent of water contents in the melt for a range of 0 to 10 wt% H₂O (Harrison and Watson, 1984; Pichavant et al., 1992; Wolf and London, 1995). Apatite solubility can be expressed as a partition coefficient for phosphorus between apatite (Ap) and melt; the lower this partition coefficient, the more apatite must dissolve to saturate the melt with phosphorus. Harrison and Watson (1984) gave the expression

$$\ln D_p^{Ap/Melt} = \frac{8400 + [10^4 \times 2.64(\text{SiO}_2 - 0.5)]}{T} - \{3.1 + [12.4 \times (\text{SiO}_2 - 0.5)]\}, \quad (1)$$

where SiO₂ is the weight fraction of silica in the melt, and T the absolute temperature. This fit appears to be valid between 45 and 75% SiO₂, 0 and 10% water, and for the range of pressures expected in the crust. For apatite solubility in peraluminous felsic melts, Pichavant et al. (1992) have extended the Harrison-Watson model and shown that:

$$P_2O_5^{PMR} = P_2O_5^{HW} + P_2O_5^{Per}, \quad (2)$$

where $P_2O_5^{HW}$ can be calculated from equation (1), and $P_2O_5^{Per}$ is given by:

$$P_2O_5^{Per} = \left(\frac{A}{CNK} - 1 \right) \exp\left(\frac{-5900}{T} - 3.22\text{SiO}_2 + 9.31 \right), \quad (3)$$

where SiO₂ is the weight fraction of silica in the melt, T is the absolute temperature and A/CNK is molar ratio of Al/(2Ca + K + Na), PMR, HW, and Per represent Pichavant-Montel-Richard, Harrison-Watson, and peraluminous, respectively. This model can account for high P₂O₅ contents observed in peraluminous granites.

2.2.2. Monazite Solubility

Both LREE and phosphorus are essential structural components of monazite. Experimental studies have demonstrated that: (1) monazite plays a significant role in buffering LREE concentration in melts, assuming sufficient phosphorus is available (Rapp and Watson, 1986; Montel, 1993), and (2) LREE saturation levels are a function of melt temperature, composition and water content. Field observations also have shown that monazite plays a dominant role in generation of LREE-depleted granitic melts (Mittlefehldt and Miller, 1983; Sawyer, 1991; Sevigny, 1993; Watt and Harley, 1993; Nabelek and Glascock, 1995) either as a residual phase in the source or a fractionating phase during differentiation. Solubility of monazite in Ca-poor felsic melts is described by (Montel, 1993):

$$\ln(REE_i) = 9.50 + 2.34D + 0.3879\sqrt{H_2O} - \frac{13318}{T}, \quad (4)$$

where

$$REE_i = \sum \frac{REE_i(\text{ppm})}{\text{at} \cdot \text{weight}(\text{g mol}^{-1})}$$

and

$$D = \frac{(Na + K + Li + 2Ca)}{Al(Al + Si)};$$

Na, K, Li, Ca, Al and Si are in atomic%; H_2O is in wt%, and T is the absolute temperature. The REE considered extend from La to Gd, excluding Eu. A higher concentration of LREE in monazite-saturated melts implies more monazite must dissolve to achieve saturation.

2.2.3. Apatite and Monazite Together

For a typical metapelite, apatite is commonly about an order of magnitude more abundant than monazite. Because phosphorus is an essential structural component common to both minerals (~43 wt% of P_2O_5 in apatite and ~29 wt% of P_2O_5 in monazite) and the concentration of phosphorus in the melt will be dominated by dissolution of apatite, the solubility of monazite in the melt will in turn be limited by apatite dissolution. In a source rock without monazite, apatite dissolution will release both LREE and phosphorus into the melt. If LREE in the melt becomes oversaturated, monazite crystallization will occur. In their experiments on the dissolution of apatite in peraluminous granitic melts, Wolf and London (1995) found that: (1) apatite has high solubility in strongly peraluminous melts, whereas monazite and xenotime solubility is lower by an order of magnitude; and (2) apatite dissolves incongruently and crystallizes REE-rich monazite on its surface, which results in LREE-depleted melts with elevated Sm/Nd ratios. Hence, the crystallization of monazite during incongruent apatite dissolution will enhance the apatite signature in the melt.

For both apatite and monazite, temperature is the most important parameter that determines their dissolution (Fig. 2). But for peraluminous or metaluminous crustal melts with D (see Eqn. 4 above) typically in the range of 0.9 to 1.1, monazite dissolution strongly depends on the water content (Rapp and

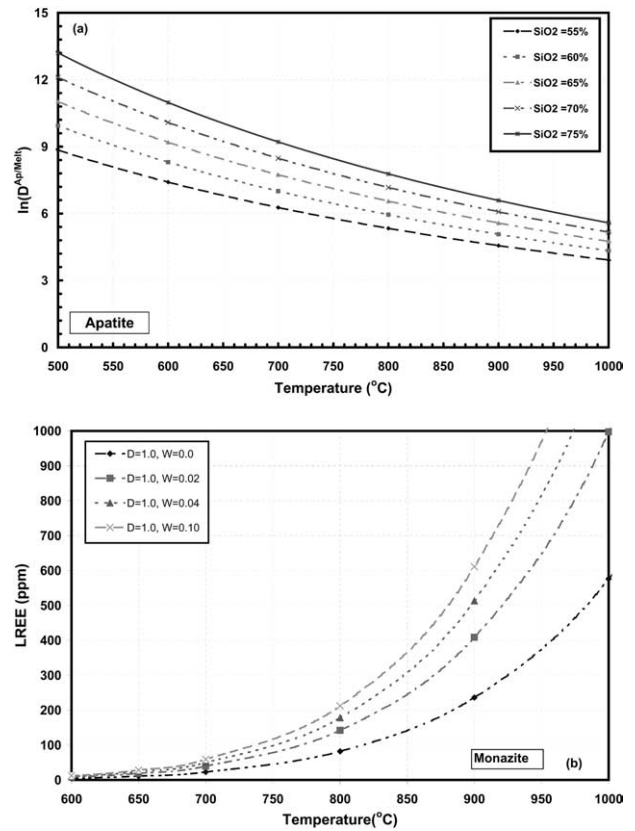


Fig. 2. (a) Solubility of apatite, expressed as a partition coefficient for P_2O_5 between apatite and melt, as a function of temperature and melt composition (silica content in wt%). A large value of $D^{Apt/Melt}$ implies a small amount of apatite needs to dissolve to reach equilibrium. (b) Solubility of monazite, expressed as LREE concentration at monazite saturation, as a function of temperature and water content (wt%) in a silicic melt with $D = 1.0$. D is defined in the text. W is the water content in the melt. A high value of LREE concentration at saturation implies a large amount of monazite needs to dissolve to reach equilibrium. At dry conditions, dissolution of monazite is inhibited.

Watson, 1986; Montel, 1993), whereas apatite dissolution does not (Fig. 2b). For common anatectic conditions, increasing water content will lower the solidus and drive partial melting of a pelitic source toward the MS2 reaction, a relatively low temperature melting reaction. This will favor dissolution of monazite relative to apatite, and less apatite will dissolve into the melt. At high temperature and low water content conditions, on the other hand, partial melting of a metasedimentary source proceeds following the MS1 reaction, and these conditions favor apatite dissolution, which in turn suppresses monazite dissolution (or encourages monazite crystallization) by raising P_2O_5 and LREE contents of the melts.

2.3. Rb-Sr and Sm-Nd Isotope Systems

In metasedimentary rocks, Sm and Nd, like the other LREE, reside predominantly in accessory phases such as apatite and monazite. In their study of Nd isotopic geochemistry of the Himalaya leucogranites, Ayres and Harris (1997) found that LREE concentrations in a melt would be controlled by mona-

zite dissolution, while apatite would contribute ~10% of the LREE, 50% of MREE, and >90% of HREE. Sm/Nd ratios are significantly elevated in apatite, whereas monazite has a ratio only slightly less than the bulk source (Bea et al., 1994; Ayres and Harris, 1997). Hence, as apatite dissolution becomes significant, the liquid will begin to reflect a higher Sm/Nd ratio than the bulk source. This result suggests that partial melts with strongly fractionated Sm/Nd ratios may have been produced under conditions favoring apatite dissolution. Furthermore, if the minerals have been evolving for some time since the last isotopic homogenization, then apatite will also carry a much more radiogenic Nd isotopic signature than the bulk rock, and apatite dissolution may lead to large Nd isotopic fractionations as well.

We emphasize the role of apatite, but monazite has complementary effects. Given a Sm/Nd ratio slightly less than its host rock (e.g., Bea et al., 1994; Ayres and Harris, 1997), monazite will evolve over time toward ϵ_{Nd} values lower than the whole rock. At the time of anatexis, if a significant amount of monazite stays in the residue, the melts will have more radiogenic Nd isotopic compositions, higher Sm/Nd ratios, and lower LREE contents than those in their source rocks. The role of monazite in fractionation of Sm from Nd in granitic magmas has been invoked to explain the Nd isotopic differences of the Sunapee Pluton from the Long Mountain Pluton in New Hampshire (Harrison et al., 1987) and signatures in the Harney Peak granite, Black Hills, South Dakota (Nabelek and Glascock, 1995). This residual monazite effect may be hard to distinguish from the apatite dissolution effect, in the absence of other geochemical information. However, if monazite stays in the residue, the melt will have a strongly LREE-depleted signature. On the contrary, the melts will have lower ϵ_{Nd} and LREE-enriched characters if a significant amount of monazite dissolves into the melt. As compared to monazite, apatite commonly has a flatter or concave-down LREE pattern. The combination of more apatite and less monazite dissolution will impart a LREE-depleted but elevated Sm/Nd ratio signature to an anatectic melt. In the case of the Lake Isabella leucosomes, both group A and group B are LREE-enriched, but less than the source rock; hence we consider apatite dissolution to be a more important control than residual monazite. For simplicity, we will only model the effect of apatite, one of the common accessory minerals in metapelites, in regulating the behavior of the Sm-Nd isotopic system during partial melting, although our discussion will consider the effects of both apatite and monazite.

In contrast to the Sm-Nd isotopic system, the Rb-Sr system is mainly controlled by rock-forming phases (e.g., plagioclase, K-feldspar, muscovite and biotite). Field and experimental observations have demonstrated the nonmodal nature of partial melting of metasedimentary sources (Patiño-Douce and Harris, 1998; Knesel and Davidson, 2002). The effect of nonmodal partial melting of a pelitic source on Rb-Sr-Ba elemental systematics was modeled by Harris and Inger (1992). Knesel and Davidson (2002) modeled the Sr isotopic compositions of anatectic melts for different melting reactions, based on their experimental results. They successfully demonstrated that incongruent melting of muscovite can account for the bimodal nature of Sr isotopic compositions of the Himalaya leucogranites. In the following, we will discuss the Rb-Sr and Sm-Nd

isotopic systematics in partial melting of a hypothetical pelitic source rock. The melting reactions we adopted for this modeling were determined experimentally by Patiño-Douce and Harris (1998). The same reactions were used by Knesel and Davidson (2002) in their evaluation of Sr isotopic disequilibrium in anatectic melting during collisional orogenesis.

2.4. Disequilibrium Melting and Crustal Anatexis

Disequilibrium melting refers to any melting event where the liquid phase is not in chemical equilibrium with the restite before melt extraction, particularly with respect to radiogenic isotopic compositions. Chemical equilibrium of trace elements will be controlled by the diffusion rates of elements between melt and residue. Chemical disequilibrium between melt and restite can occur if melt is extracted from its source too rapidly to allow complete equilibration between them. In disequilibrium melting, the concentration of an element in the melt depends on its concentration in the minerals of the initial solid and on the mineral proportions which dissolve into the melt (Allègre and Minster, 1978). Disequilibrium melting has been documented in laboratory experiments (e.g., Johannes, 1980; Hammouda et al., 1996; Knesel and Davidson, 1996, 1999) and inferred from a number of geochemical observations on anatectic complexes (Naslund, 1986; Kaczor, et al., 1988; Barbey et al., 1989; Sawyer, 1991; Watt and Harley, 1993; Barbero et al., 1995; Watt et al., 1996; Ayres and Harris, 1997). Rubie and Brearley (1990) performed experiments on incongruent melting of muscovite plus quartz and suggested that disequilibrium can result either from significant temperature oversteps or from fluid infiltration in rocks that are above the fluid-present melting solidus. Both mechanisms will result in rapid melting and achieve significant melt fractions in less than one year. Disequilibrium in formation of anatectic complexes has also been observed in many field studies (Naslund, 1986; Kaczor et al., 1988; Barbey et al., 1989; Sawyer, 1991; Watt and Harley, 1993; Barbero et al., 1995; Ayres and Harris, 1997; Watt et al., 1996). In particular, recent studies of migmatites in the Hercynian anatectic complex of Toledo, Spain (Barbero et al., 1995), and of pelitic gneiss and leucogranite in the Himalaya (George and Bartlett, 1996) show that Sr and Nd isotopic heterogeneity can be preserved on the grain scale before and during anatexis. Deniel et al. (1987) presented Sr and Nd isotopic composition data from the Manaslu granite, Himalaya, Nepal and demonstrated a wide range of initial Sr and Nd isotopic ratios showing that isotopic heterogeneity on a meter scale survived magmatic processes. These results suggest that isotopic heterogeneity among phases in a source rock may well be preserved until a major melting event, which is consistent with the observations of fine scale disequilibrium in high-grade metamorphic suites and extremely sluggish rates of metamorphic reactions (e.g., Baxter and DePaolo, 2000; Carlson, 2002). These examples are not intended to argue that isotopic equilibrium never occurs in crustal rocks, but to stress the widespread occurrence of disequilibrium and the geochemical consequences if such disequilibrium persists until and through a partial melting event.

Such field and experimental observations contradict with theoretical considerations suggesting that isotopic heterogeneity in the source may not survive prograde heating during

Table 1. Parameters for Rb, Sr, Sm, and Nd trace element modeling.^a

Mineral	wt%	Rb (ppm)	Sr (ppm)	Rb/Sr	Sm (ppm)	Nd (ppm)	Sm/Nd	⁸⁷ Str/ ⁸⁶ Sr(0)	$\epsilon_{Nd}(0)$
Quartz	43	0.0	0.0	—	0.00	0.00	—	—	—
Plagioclase	28	1.2	288.0	0.0042	2.25	19.53	0.1152	0.7376	-11.62
Muscovite	22	230.0	41.0	5.6098	0.05	0.29	0.1724	0.7722	-11.18
Biotite	2	580.0	30.0	19.3333	0.05	0.29	0.1724	0.8396	-11.18
Aluminosilicate	4.5	0.0	0.0	—	0.00	0.00	—	—	—
Apatite	0.5	0.0	0.0	—	175.30	432.00	0.4058	—	-9.36
Monazite	0.01	0.0	0.0	—	18000	104000	0.1731	—	-11.17
K-feldspar	0	35.0	135.0	0.2593	0.09	0.40	0.2250	—	-10.77
Whole-Rock	100	62.5	90.3	0.6928	2.6885	12.6296	0.2129	0.7417	-11.09

^a Rb, Sr, Sm and Nd concentration data are adapted from [Bea and Montero \(1999\)](#), [Bea et al. \(1994\)](#), [Ayres and Harris \(1997\)](#), and [Knesel and Davidson \(2002\)](#). Initial Sr and Nd isotopic compositions are set arbitrarily.

collisional orogenesis ([Harris and Ayres, 1998](#)). However, rapid heating induced by an external heat source such as shear heating on a major shear zone or input from magma underplating or intrusion may favor disequilibrium melting ([Harrison et al., 1999](#); [Petford and Gallagher, 2001](#)). Furthermore, there are several reasons to expect that the Sm-Nd system in particular may preserve heterogeneity. (1) Diffusion of Sm and Nd is much slower than that of Rb and Sr in a silicic melt. (2) Sm and Nd concentrations in the melt and the residue are dominantly controlled by accessory minerals such as apatite and monazite. (3) A large fraction of accessory minerals reside as inclusions in major minerals ([Bea et al., 1994](#)), which leads to high potential for them to be protected from homogenization during prograde metamorphism. Such a textural relationship has been a major problem in using the Sm-Nd isotope system to date the timing of metamorphic events. (4) The relationship between degree of melting and degree of dissolution of accessory minerals in crustal anatexis is poorly known. (5) Since phosphorous is an essential structural component in apatite, the behavior of associated LREE in apatite is coupled by the behavior of the highly charged and very slowly diffusing phosphorus cation. Hence, during crustal anatexis, we consider the Sm-Nd isotope system as a strong candidate for disequilibrium. For the Rb-Sr system, the relatively large grain size of micas and feldspars leads to the possibility that this system also will not be completely homogenized by volume diffusion or metamorphism. Therefore, it is of use to explore Sr and Nd isotopic systematics of anatectic melts by allowing each phase to begin the melting process with its own isotopic composition and applying disequilibrium melting equations.

2.5. Melting Models

During nonmodal partial melting, the concentration of element *i* in the melt (C_L^i) is given by

$$C_L^i = (x_{m1}C_1^i + x_{m2}C_2^i + \dots + x_{mn}C_n^i) - (y_{c1}C_1^i + y_{c2}C_2^i + \dots + y_{cm}C_m^i), \quad (5)$$

where x_{mj} and y_{ck} are the weight proportions of mineral *j* entering and mineral *k* crystallizing from the melt, respectively, and C_j^i and c_k^i are the concentrations of element *i* in mineral *j* and mineral *k*, respectively ([Allègre and Minster, 1978](#); [Prinzhofer and Allègre, 1985](#); [Barbey et al., 1989](#); [Sawyer, 1991](#)). Eqn. 5 is written to account for incongruent melting processes;

for congruent partial melting reactions, y_{ck} , are zero. The initial concentration of element *i* in the source rock is given by

$$C_0^i = X_1C_1^i + X_2C_2^i + \dots + X_nC_n^i, \quad (6)$$

where X_j is the weight proportion of mineral *j* in the source rock. For the Sm-Nd isotope system

$$\left(\frac{Sm}{Nd}\right)_L = \frac{(x_{m1}C_1^{Sm} + x_{m2}C_2^{Sm} + \dots + x_{mn}C_n^{Sm}) - (y_{c1}C_1^{Sm} + y_{c2}C_2^{Sm} + \dots + y_{cm}C_m^{Sm})}{(x_{m1}C_1^{Nd} + x_{m2}C_2^{Nd} + \dots + x_{mn}C_n^{Nd}) - (y_{c1}C_1^{Nd} + y_{c2}C_2^{Nd} + \dots + y_{cm}C_m^{Nd})} \quad (7)$$

and

$$\left(\frac{Sm}{Nd}\right)_0 = \frac{X_1C_1^{Sm} + X_2C_2^{Sm} + \dots + X_nC_n^{Sm}}{X_1C_1^{Nd} + X_2C_2^{Nd} + \dots + X_nC_n^{Nd}}, \quad (8)$$

where $(Sm/Nd)_L$ and $(Sm/Nd)_0$ are the Sm/Nd ratios for the melt and initial source rock, respectively. If each phase entering the melt preserves its radiogenic isotopic compositions (e.g., Sr, Nd, or Pb), then the isotopic composition of the melt can be calculated. Nd isotopic composition of the melt is calculated by

$$(\epsilon_{Nd})_L = \frac{\sum x_{mi}C_{Nd}^i \epsilon_{Nd}^{mi}}{\sum x_{mi}C_{Nd}^i}, \quad (9)$$

where $(\epsilon_{Nd})_L$ is the Nd isotope composition of the melt and ϵ_{Nd}^i is Nd isotope composition in mineral *i*. Whole-rock $(\epsilon_{Nd})_{wr}$ of the source is given by

$$(\epsilon_{Nd})_{wr} = \frac{\sum X_i C_{Nd}^i \epsilon_{Nd}^{mi}}{\sum X_i C_{Nd}^i}. \quad (10)$$

For the Rb-Sr system, we adopt a precisely analogous treatment to obtain Rb/Sr and ⁸⁷Str/⁸⁶Sr ratios of the melts.

The parameters used in this study are listed in [Table 1](#). The mineral compositions of the hypothetical source rock are slightly different from those used by [Patiño-Douce and Harris \(1998\)](#). We replace garnet with aluminosilicate and lower its proportion by 0.5 wt% to accommodate apatite and monazite in our hypothetical source. Both garnet and aluminosilicate are not reactants in the melting reactions that we examined such that this modification will not affect the applicability of the MS1 and MS2 reactions to our trace element modeling. In all the models to follow, we assume that a partial melting event occurred at 100 m.y. after the last isotopic homogenization

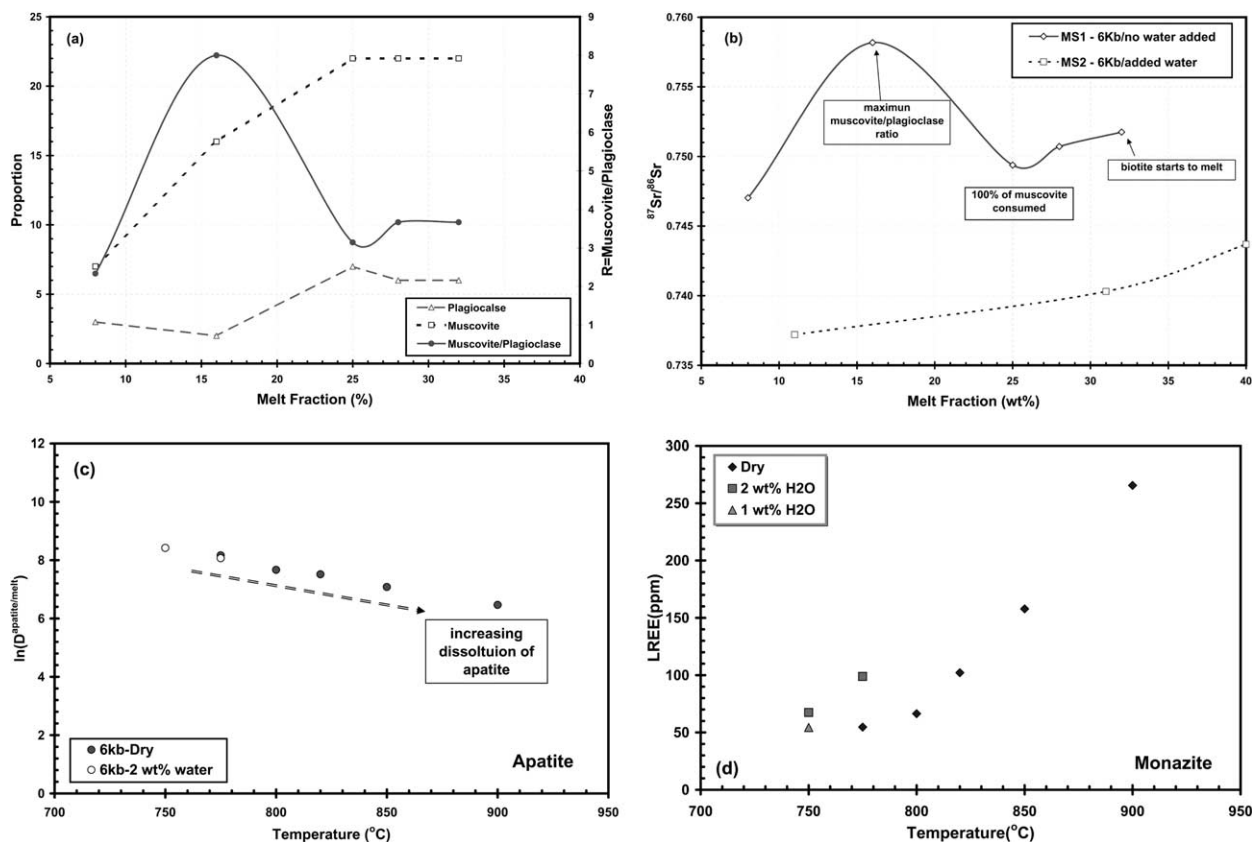


Fig. 3. (a) Plot of proportions of muscovite and plagioclase entering the melt (left-hand axis) and the ratio of these proportions (right-hand axis) against melt fraction for MS1 reaction. Data are from Patiño-Douce and Harris (1998). MS1 represents fluid-absent muscovite dehydration melting. (b) $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the melt as a function of degree of melting for MS1 and MS2 reactions at 6 kb. The proportions of reacting phases entering the melt are from table 2 of Patiño-Douce and Harris (1998). We assign a Sr isotopic composition of each phase involved in the MS1 and MS2 partial melting reactions. The data are listed in Table 1. For the melts derived from the MS1 reaction, their $^{87}\text{Sr}/^{86}\text{Sr}$ ratios increase as melting proceeds until the proportion of muscovite to plagioclase entering the melt reaches its maximum, and then decrease during continuous melting of muscovite + plagioclase + quartz until muscovite is completely consumed. $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the melts from the MS2 reaction increase as melting proceeds up to a higher degree (40%). Addition of water stabilizes muscovite and thus limits its participation in MS2 melting reaction, which results in lower $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the melts derived from such a melting reaction than those from MS1 reaction. (c) Plot of apatite solubility as a function of temperature for the melts from reaction MS1 at 6 kb with and without 2 wt% added water. (d) Plot of monazite solubility as a function of temperature for the melts from MS1 partial melting at 6 kb with 0, 1 and 2 wt% added water, respectively.

event (e.g., melting or high-grade metamorphism). All the phases began with arbitrarily chosen equal ε_{Nd} and $^{87}\text{Sr}/^{86}\text{Sr}$ values and evolved for 100 Ma with their respective Sm/Nd and Rb/Sr ratio, such that they have different ε_{Nd} and $^{87}\text{Sr}/^{86}\text{Sr}$ values at the time that the hypothetical partial melting occurred. In the incongruent MS1 reaction, the new phases crystallizing out of the melt are assumed to be in chemical and isotopic equilibrium with the melt, and the concentrations of Rb, Sr, Sm and Nd in each phase are equal to those in the source. Such assumptions will slightly affect the calculated Sm/Nd or Rb/Sr ratio but not Sr or Nd isotopic compositions in the melt.

3. RESULTS AND DISCUSSION

The profound geochemical consequences of accessory phase dissolution during partial melting of high-grade metapelites are illustrated in Figures 3 and 4. Given the details of how the incongruent melting of muscovite proceeds, we can examine

how the Sr isotopic composition evolves as a function of degree of melting. To calculate the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the melts from reactions MS1 and MS2, the Sr concentrations, $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, and proportions of each phase that contributes to the melt should be known. For each reactant, the Sr concentration and $^{87}\text{Sr}/^{86}\text{Sr}$ ratio are listed in Table 1. The proportions of each phase entering and leaving the melt at each degree of melting are from Table 2 in Patiño-Douce and Harris (1998). We only show the results for their experiments at 6 kbar both with and without H₂O (Fig. 3). Figure 3a shows, for the MS1 reaction, the amounts of muscovite and plagioclase entering the melt, and the ratio between these two, vs. degree of melting. Muscovite is always a larger contributor to the melt, but the ratio of muscovite/plagioclase has a distinct maximum at 16% melting, and muscovite is exhausted from the residue at 25% melting. Figure 3b shows $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the melts derived from both MS1 and MS2 reactions vs. degree of melting. There are

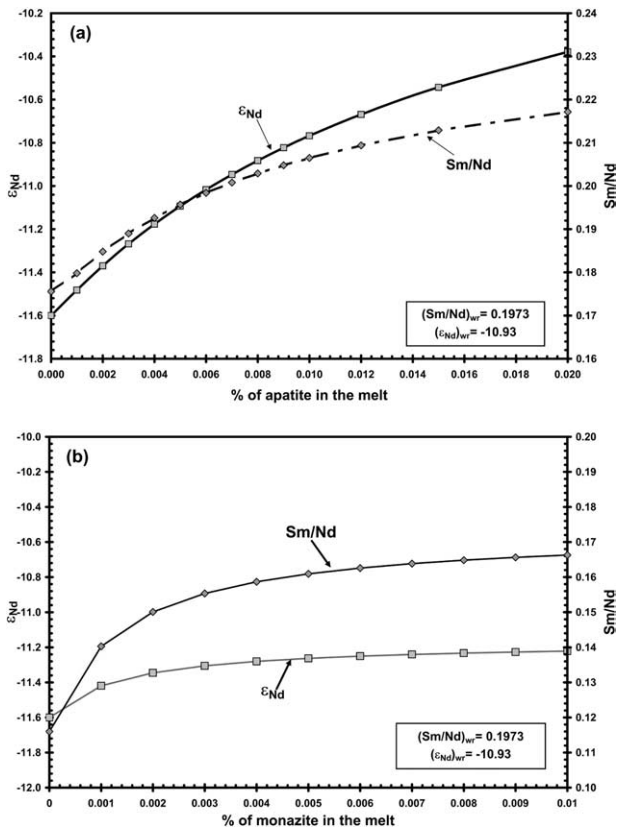


Fig. 4. Diagrams showing effect of apatite (a) and monazite (b) dissolution on Nd isotope systematics of the melts formed in a partial melting event occurring 100 m.y. after the last event that homogenized Nd isotope compositions among minerals. We assume all the minerals have the same initial ϵ_{Nd} value of -10.0 but different Sm/Nd ratios at a time 100 Ma before the melting event. At the time of melting, minerals involved in the melting will have different ϵ_{Nd} values and Sm/Nd ratios. (a) For the case of apatite, both ϵ_{Nd} value and Sm/Nd ratio of the melt are dominated by the amount of apatite dissolved into the melt. For this specific case, both ϵ_{Nd} value and Sm/Nd ratio of the melt approach the values in the source rock when the melt dissolves 0.008% apatite. With higher amount of apatite dissolved, the melt evolves toward higher ϵ_{Nd} value and Sm/Nd ratio than those in the source. (b) For the case of monazite, increasing dissolution will result in increases in both Sm/Nd ratio and ϵ_{Nd} value in the melt. But they will never exceed those in the source. Combination of increasing apatite dissolution with decreasing monazite dissolution will enhance the apatite signature, i.e., elevated Sm/Nd and ϵ_{Nd} in the melt.

several stages in the evolution of melts from reaction MS1: the $^{87}Sr/^{86}Sr$ ratio of the melt (1) increases until the ratio of amount of muscovite to the amount of plagioclase entering the melt reaches its maximum; (2) decreases until muscovite is exhausted; and (3) finally increases slightly due to a small amount of biotite entering the melt. All the melts from MS1 have higher $^{87}Sr/^{86}Sr$ ratios than their source (0.7417). On the other hand, the $^{87}Sr/^{86}Sr$ ratios of the melts from the MS2 reaction increase as melting proceeds up to a higher degree (40%), but they are all lower than the ratio in their source. In reaction MS2, water plays a dual role: it lowers the solidus of quartz + plagioclase and also stabilizes muscovite and thus limits its participation in this melting reaction. A reduced contribution from muscovite relative to plagioclase entering the melt results in lower Rb/Sr

and $^{87}Sr/^{86}Sr$ ratios of the melts derived from reaction MS2 than those from reaction MS1.

The apatite solubility in the melts derived either from MS1 or MS2 reaction is calculated using the Harrison-Watson model and the melt compositions reported by Patiño-Douce and Harris (1998). The result is shown in Figure 3c, which indicates that for both MS1 and MS2 reactions, the apatite solubility increases with melting temperature, but MS2 begins and produces significant melt at lower temperature than MS1, where apatite is relatively insoluble as compared to monazite. We used the melt compositional data from the experiments of Patiño-Douce and Harris (1998) at 6 kbar with 0, 1, and 2 wt% added H_2O , respectively, to evaluate the solubility of monazite under these conditions. As shown in Figure 3d, even though monazite solubility increases with the melting temperature, added water substantially increases its solubility. For example, at 775°C, monazite solubility at 2 wt% added H_2O is double as that at 1 wt% added H_2O (Fig. 3b). As discussed above, with more water added, melting temperature will decrease and lead to fluxed-melting reactions that are dominated by monazite dissolution. Hence, high- H_2O , low-temperature melting favors monazite dissolution, whereas low- H_2O , higher-temperature melting favors apatite dissolution and possible monazite precipitation.

Using the data listed in Table 1 and the apatite partition coefficients shown in Figure 3c, we can calculate the Nd isotopic composition of the melts. Figure 4a is a plot of ϵ_{Nd} values and Sm/Nd ratios of melts derived from the hypothetical source against the fraction of apatite dissolved into the anatectic melts. The ϵ_{Nd} value and Sm/Nd ratio of the melt strongly depend on the amount of apatite dissolved. With $\sim 0.008\%$ apatite dissolved into the melt, the melt has an ϵ_{Nd} value and Sm/Nd ratio about equal to those of the source rock. As more apatite dissolves into the melt, ϵ_{Nd} and Sm/Nd ratio increase in the melt. Note that the source rock contains 0.5% apatite, and therefore a modal melting model would be far off-scale to the right. We also evaluate the effects of monazite dissolution on Sm/Nd ratio and Nd isotopic composition in the melt. In the calculations, we assume only monazite is present in the source. The result is shown in Figure 4b, which suggests that: (1) monazite dominates the Sm and Nd budget; (2) if all the monazite stays in the residue, the melt will have substantially lower Sm/Nd and ϵ_{Nd} values than the source; and (3) as the amount of monazite dissolved into the melt increases, both Sm/Nd and ϵ_{Nd} values will approach but never exceed those in the source. The features shown in Figure 4b are very different from those in Figure 4a, which can be used to put some constraints on the processes that are responsible for the observed Sr-Nd systematics in anatectic melts.

The features shown in this modeling may account for observations of Nd isotope disequilibrium and Sm/Nd parent-daughter fractionation in the Himalaya leucogranites (Deniel et al., 1987; Ayres and Harris, 1997); the migmatites in the Hercynian anatectic complex of Toledo, Spain (Barbero et al., 1995); the Proterozoic Harney Peak Granite from Black Hills, North Dakota, USA (Walker et al., 1986; Nabelek and Glascock, 1995); and the leucogranitic dikes and leucosomes of the Isabella pendant, southern Sierra Nevada (Zeng, 2003; Zeng et al., 2005).

Given the implicit complexity of crustal anatexis for the

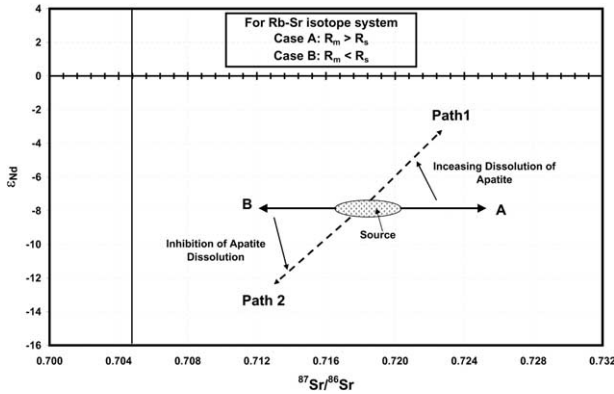


Fig. 5. Schematic diagram showing the effects of nonmodal partial melting and dissolution of accessory minerals on the Sr-Nd systematics of metasediment-derived melts. Path A represents melts dominated by muscovite/biotite dehydration melting characterized by high R_m/R_s ratios (>1), while Path B with low R_m/R_s ratios (<1) represents those reactions with minor involvement of either muscovite or biotite. Both A and B have no involvement of accessory phases. Path 1 represents partial melting reactions in which the T and X_{H_2O} environment associated with Type A melting reactions favor apatite but not monazite dissolution and in which the melt will have elevated Rb/Sr, $^{87}\text{Sr}/^{86}\text{Sr}$, Sm/Nd, and ϵ_{Nd} . Path 2 represents those partial melting reactions in which the T and X_{H_2O} environment associated with Type B melting reactions favor monazite over apatite dissolution, in which case the melt will have lower Rb/Sr, $^{87}\text{Sr}/^{86}\text{Sr}$, Sm/Nd, and ϵ_{Nd} than its source. R_m is the ratio of amount muscovite plus biotite over amount of plagioclase plus K-feldspar entering the melt. R_s is the ratio of amount of muscovite plus biotite over amount of plagioclase plus K-feldspar in the source. See text for discussion.

Rb-Sr and Sm-Nd isotopic systematics, it is instructive to examine likely trajectories for melts derived from partial melting of a pelitic source in $\epsilon_{\text{Nd}}-^{87}\text{Sr}/^{86}\text{Sr}$ space. To generalize our discussion beyond muscovite dehydration melting, we will define a ratio, R, describing the relative contribution from Rb-rich over Sr-rich phases to a source and a melt, which is critical for Rb and Sr geochemistry of anatectic melts. For the melt, R_m is given by

$$R_m = \frac{x_{Ms} + x_{Bt}}{x_{Pl} + x_{Kfs}} \quad (11)$$

and for the solid source rock, R_s is given by

$$R_s = \frac{X_{Ms} + X_{Bt}}{X_{Pl} + X_{Kfs}} \quad (12)$$

In Figure 5, case A represents melts dominated by muscovite/biotite dehydration melting and hence high R_m/R_s ratios (>1), including, for example, reaction MS1 given above. Case B, with low R_m/R_s ratios (<1) represents those reactions with minor or no involvement of either muscovite or biotite, e.g., reaction MS2. Under disequilibrium conditions, in case A, the melts will have more radiogenic Sr isotopic compositions than their source and in case B the opposite is true. Considering only relative contributions of the major rock-forming phases and not the behavior of accessory minerals, case A and B at this stage make no prediction of significant variations of ϵ_{Nd} . However, as we have emphasized, experiments on accessory phase dissolution provide a likely basis for coupling between A- and

B-type Sr isotopic behavior and characteristic Nd isotopic behavior. Experiments on partial melting of metapelites have shown that type B partial melting usually occurs at a lower temperature but higher water content than does type A partial melting. As shown by Eqns. 1, 3, and 4, at the higher temperatures and dry conditions where muscovite or biotite dehydration melting play a dominant role and case A is expected, dissolution of apatite will increase, but monazite dissolution will be relatively inhibited by low water activity (Fig. 2b). Melts derived from such partial melting reactions will have more radiogenic Nd isotopic compositions than their source (Path 1 in Fig. 5). The lower temperatures and high water contents that favor case B will inhibit apatite dissolution relative to monazite dissolution (Fig. 2), which results in more negative ϵ_{Nd} values in the melt than its source (Path 2 in Fig. 5). Therefore, for melting timescales where kinetics are relevant and the various P-T and water content conditions under which crustal anatexis occurs, a correlation between Sr and Nd isotopes will emerge: the most likely paths for a melt to follow in $\epsilon_{\text{Nd}}-^{87}\text{Sr}/^{86}\text{Sr}$ space will be either Path 1 or Path 2. Path 1 represents those partial melting reactions in which muscovite/biotite dehydration and apatite but not monazite dissolution play a significant role; the melt will have elevated Rb/Sr, $^{87}\text{Sr}/^{86}\text{Sr}$, Sm/Nd, and ϵ_{Nd} relative to its source. By contrast, Path 2 represents those partial melting reactions in which muscovite/biotite dehydration plays an insignificant role, and apatite but not monazite stays as a residue; the melt will have lower Rb/Sr, $^{87}\text{Sr}/^{86}\text{Sr}$, Sm/Nd, and ϵ_{Nd} than its source.

As discussed in the introduction section, we found a strong correlation between $\epsilon_{\text{Nd}}(i)$ values and trondhjemitic vs. granitic character in the leucosomes from the migmatite zone of the Lake Isabella pendant. In $\epsilon_{\text{Nd}}-^{87}\text{Sr}/^{86}\text{Sr}$ space, the granitic group A leucosomes are distributed along Path 1 (high ϵ_{Nd} and $^{87}\text{Sr}/^{86}\text{Sr}$), and the trondhjemitic group B leucosomes are distributed along Path 2 (low ϵ_{Nd} and $^{87}\text{Sr}/^{86}\text{Sr}$), but their Sr isotopic variation is smaller as compared to variations in their Nd isotopic compositions (Zeng et al., 2005). Those leucosomes with high ϵ_{Nd} values (Group A) also have higher P_2O_5 contents than those with low ϵ_{Nd} values (Group B) (Fig. 1). At first glance, these correlations seem to be fortuitous, but they can be explained as systematic consequences of water activity variations, considering the melting reactions and the equations for apatite and monazite dissolution above. Dehydration melting without excess water (by a reaction similar to MS1) occurs at a fairly high temperature, involves major contributions of muscovite and biotite to the melt, and favors substantial dissolution of apatite (both absolutely and relative to monazite). The contribution of micas to the melt leads to potassium-rich, i.e., granitic, character and to high Rb/Sr and radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$, and the favorable apatite dissolution leads to a melt with high P_2O_5 , high Sm/Nd, and radiogenic ϵ_{Nd} . These are all the characters seen together in group A leucosomes, which we therefore associate with Path 1 in Figure 5. Conversely, fluxed-melting with excess water leads to melting by a reaction similar to MS2 at fairly low temperature, involving dominantly melting of plagioclase and favoring monazite dissolution over apatite, though in an absolute sense both accessory minerals dissolve more slowly. The contribution of plagioclase leads to melts that are sodium-rich, i.e., trondhjemitic, and to low Rb/Sr and nonradiogenic $^{87}\text{Sr}/^{86}\text{Sr}$. The slower dissolution of phosphates

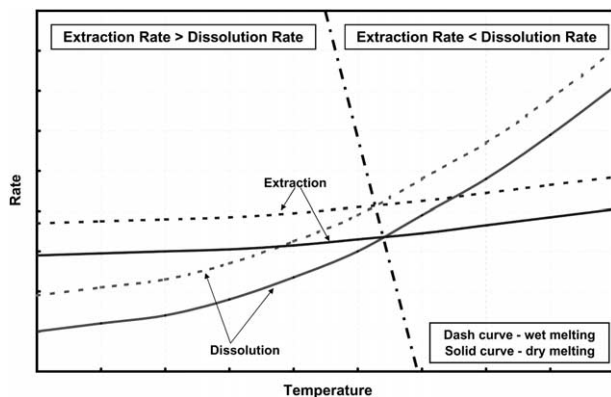


Fig. 6. Schematic plots showing the competing effects of melt extraction and accessory mineral dissolution on the isotopic geochemistry of the melts derived from partial melting of a metasedimentary protolith (modified after Sawyer, 1991; Watt and Harley, 1993). The steep curve represents the temperature and water-dependent accessory dissolution rate, and the gentle curve is the hypothetical temperature and water-dependent melt extraction rate. Melts generated within the area bounded by the dissolution and extraction curves to the left of the dot-dashed line will yield disequilibrium compositions. At lower water content (solid lines), the field of disequilibrium melting can extend to higher temperature. With increasing water contents (dashed lines), the field of disequilibrium melting shrinks to lower temperatures, as the rate of accessory mineral dissolution surpasses that of melt extraction.

overall leads to low P_2O_5 , but the larger role for monazite gives melts with low Sm/Nd and nonradiogenic ϵ_{Nd} . These are all the characteristics of group B leucosomes, which we therefore associate with Path 2 in Figure 5. Finally, since all these leucosomes have LREE patterns subparallel to the metapelite, and slightly lower LREE than their source (Zeng, 2003), we emphasize the variations in apatite dissolution rather than a large role of residual monazite in their formation.

Although Sr and Nd isotopic disequilibrium may be a common feature of crustal anatexis in dynamic orogenic environments, there are a number of factors that will complicate or obscure field observations. The primary factors affecting the fate of anatectic melts are their chemistry and physical properties, and the thermal and mechanical state of the crust around and above the zone of melting (Sawyer, 1991; Watt and Harley, 1993; Watt et al., 1996). For those elements concentrated in accessory phases, the melt extraction rate and accessory mineral dissolution rate are critical factors that will determine their partitioning into the melt. If melt extraction rate is greater than accessory phase dissolution rate, the melt will have a disequilibrium signature with respect to its source (Fig. 6). But at higher temperatures, accessory phases dissolve rapidly enough to exceed the rate of melt extraction (Sawyer, 1991; Watt and Harley, 1993). Therefore, assisted by rapid diffusion, the melts may be in equilibrium with their residue for those elements with high diffusivities. For instance, the leucogranite dikes which formed at relatively higher temperatures than leucosomes in the Isabella pendant have acquired equilibrium Sr isotopic compositions with respect to their protoliths (Zeng, 2003). For anhydrous melting such as reaction MS1, high dilatational strain generated by positive volume change of muscovite dehydration melting may encourage rapid melt segregation and extraction, and thus may assist in promoting both Sr or Nd isotopic disequilibrium. In

contrast, hydrous melting (e.g., MS2) may hinder complete extraction of melts from their source due to negative volume change during melting, and thus melts produced by hydrous reactions may have Sr or Nd isotopic compositions closer to equilibrium with their sources. Indeed, in the Lake Isabella suite, the magnitude of disequilibrium relative to bulk migmatites or source metapelites is larger for the group A leucosomes that we argue formed by a reaction like MS1.

4. SUMMARY AND CONCLUSIONS

We have incorporated accessory mineral dissolution into nonmodal partial melting modeling to examine the geochemical consequences of muscovite dehydration melting of pelitic sources. In particular, assuming a hypothetical source consisting of quartz, plagioclase, muscovite, biotite, aluminosilicate, apatite and monazite, and following the partial melting reactions MS1 and MS2 determined by Patiño-Douce and Harris (1998), dissolution of apatite or monazite into the melt combined with muscovite melting has profound effects on the Rb-Sr and Sm-Nd isotopic systematics of anatectic melts. Reasonable melting conditions and rates should lead to disequilibrium in the Sm-Nd isotopic system, where Nd isotopic compositions of the melts strongly depend on the amount of apatite dissolved in the melts. The Rb-Sr isotopic system is primarily controlled by rock-forming minerals, especially Rb-rich minerals (e.g., muscovite and biotite) vs. Sr-rich minerals (e.g., plagioclase and K-feldspar), which at the same time control the abundances of Na and K in the melt and the granitic vs. trondhjemitic character. Although these systems respond to different aspects of the melting process, we find that disequilibrium partial melting of pelitic source results in melts following one of two paths in ϵ_{Nd} - $^{87}Sr/^{86}Sr$ space. Path 1 represents those partial melting reactions that favor muscovite/biotite dehydration and apatite but not monazite dissolution, leading to melts with elevated K/Na, P_2O_5 , Rb/Sr, $^{87}Sr/^{86}Sr$, Sm/Nd, and ϵ_{Nd} values. This path is well represented by the group A leucosomes from the Lake Isabella pendant. In contrast, Path 2 represents those partial melting reactions in which muscovite/biotite dehydration plays an insignificant role, and favors monazite relative to apatite dissolution although both accessory minerals dissolve more slowly. These melts will have lower K/Na, P_2O_5 , Rb/Sr, $^{87}Sr/^{86}Sr$, Sm/Nd, and ϵ_{Nd} values than their sources. This path is well represented by the group B leucosomes from Lake Isabella. The master variables controlling both accessory phase dissolution (and hence the Sm-Nd system and P_2O_5 contents), and melting reaction (and hence the Rb-Sr systematics and K/Na ratio) are temperature and water content. These results can account for, but do not require, Sr or Nd isotopic disequilibrium as observed in a number of migmatite complexes and leucogranite plutons. Inefficient melt extraction and rapid diffusion can still lead to equilibrium between melts and their residues, as observed in other field areas.

Acknowledgments—Support for this research was provided by NSF grants EAR-9815024 and EAR-0087347. We thank D.W. Mittlefehldt, E.B. Watson, K. Knesel, M. Wolf and an anonymous reviewer for critical comments and suggestions. We greatly appreciate their comments that helped us clarify and improve the quality of our manuscript.

Associate editor: D. Mittlefehldt

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