Widely variable granite production from a restitic metasedimentary terrane by volatile redistribution

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ABSTRACT

The amount of melt extracted from a metasedimentary source is a major parameter in quantitative models of crustal processes; however, quantification of melt volume is thwarted by the heterogeneity of the crust and a lack of estimates concerning the amount of fluid present. We observe that highly systematic trends in metasedimentary rocks allow treatment of protoliths as two component mixtures, which, coupled with mass balance and minimization by a residuals function, permit development of a general algorithm for quantification of melt loss and protolith composition based on the composition of the restitic rocks. The utilization of anhydrous compositions results in melt loss estimates independent of volatile phase, and particularly suitable for demonstration of the role of fluid. The application of this method to restites of the Mount Stafford complex in Australia reveals melt loss with unprecedent detail, and shows that melt loss from rocks of the same

metamorphic grade was highly variable from negligible to 70 wt%. The melting mechanism in the complex changed from fluid saturated near the solidus, to fluid fluxed at higher temperatures, and fluid deficient in adjacent rocks. Heterogeneous melt productivity was caused by the redistribution of fluids generated by muscovite breakdown during heating. The overall melt productivity of mid-crustal melting was remarkably high and volume of extracted granitic melt perhaps was similar to the volume of restite.

INTRODUCTION

Partial melting transforms crustal rocks by drainage of granitic magma leaving behind dehydrated and melt depleted restites (Stepanov et al., 2024). The amount of melt generated during anatexis is a key parameter for tectonic and geochemical models (Rushmer, 2001; Yakymchuk and Brown, 2014) and numerous attempts have been made to quantify melt production based on composition of the residual crust. One approach, known as melt-reintegration, is based on the thermodynamic calculation of P-T pseudosections coupled with assumptions about water content and melt extraction history (White et al., 2004; Korhonen et al., 2013; Bartoli, 2017). The melt-reintegration methods require assumptions about fluid content, and commonly free fluid at the solidus is assumed. The caveat is that an increase in melt productivity with the addition of an external fluid, as in case of fluid-fluxed melting, and a decrease in melt productivity due to dehydration are difficult to demonstrate with this method.

Another approach is mass balance, which utilizes the law of conservation of matter and assumptions about protolith composition to quantify the mass transfer in metamorphic and metasomatic rocks (MacRae and Nesbitt, 1980; Olsen, 1985; Bea, 1989). For mass balance, the selection of the protolith is crucial and a major source of uncertainty.

This work presents a method that utilizes the systematic variations in the composition of the turbiditic metasedimentary rocks to obtain estimates of the mass of melt loss from the melt-depleted restitic rocks, with appropriate uncertainties. The application of the method to rocks of a metamorphic complex in Australia reveals the heterogeneity of melt generation in restites with unprecedented detail and offers new perspective on the fate and role of hydrous fluids during anatexis. The equations behind the calculations are general and similar approaches could be developed for other mass transfer problems in Earth sciences.

MODEL

Melting mass balance

The mass balance of melting is described by a general equation:

$$M_{protolith}=M_{melt} + M_{restite} \qquad (1)$$

Where $M_{protolith}$, M_{melt}, and $M_{restite}$ are the masses of original rock, extracted melt, and restite, respectively. For a concentration of the element C^{El}, equation (1) becomes:

$$C_{protolith}^{El} = \varphi * C_{melt}^{El} + (1-\varphi) * C_{restite}^{El} \quad (2)$$

Where φ is the mass fraction of the melt extracted from protolith.

A model for protolith composition

The Paleoproterozoic (1795-1805 Ma (Rubatto et al., 2006)) Mount Stafford metamorphic complex is a prime example of a low pressure, mid-crustal, anatectic terrane (Vernon et al., 1990; Collins and Vernon, 1991; Greenfield et al., 1998; White et al., 2003). The Mount Stafford is located in the Proterozoic Arunta Inlier Block, central Australia, and is part of the larger Anmatjira–Reynolds Range belt. The investigations of the geochronology, geochemistry and petrology of Mount Stafford has made it one of the most documented metamorphic complexes in the world (Rubatto et al., 2006; Palya et al., 2011; Bartoli, 2017; Williams et al., 2018).

The protolith for the metamorphic rocks were the Lander Rock Beds Formation, which is composed of turbiditic metasedimentary rocks and orthogneisses. The complex has been divided (Greenfield et al., 1998) into five zones (Fig. 1A). Zone 1 is composed of muscovite-biotite schists. Zone 2 is defined by the breakdown of muscovite, appearance of andalusite, cordierite, and K-feldspar and localized melting (White et al., 2003; Palya et al., 2011). In zone 3 biotite content decrease coincides with intensive migmatization (Greenfield et al., 1998; White et al., 2003). Zone 4 is outlined by the appearance of orthopyroxene and zone 5 has been described as an "enigmatic hybrid diatexite" by Greenfield et al. (1998).

The temperatures ranged from ~500 °C in zone 1 to peak 800–810 °C in zone 4 (Vernon et al., 1990; Greenfield et al., 1998; White et al., 2003) and slightly lower temperatures in zone 5 (Greenfield et al., 1998). Pressure was 0.22–0.33 GPa during peak metamorphism and increased afterwards, indicating crustal thickening (Collins and Vernon, 1991). Metamorphism, deformation, and magmatism at Mount Stafford were synchronous, and migrated outward to the lower grade zones (Collins and Vernon, 1991). The granitic intrusions likely were heat source, with a possible contribution from mantle and pre-metamorphic mafic sills (Collins and Vernon, 1991).

The composition of subsolidus metasedimentary rock from zone 1 shows systematic variations, ranging from psammitic, with > 80 wt% SiO_2, to pelitic, with <55 wt% SiO_2, endmembers (Fig. 1B), as is typical for turbidite sediments (Forshaw and Pattison, 2022). There is a negative correlation between SiO_2 and (Na_2O+K_2O), and a positive correlation between

Al_2O_3 and K_2O (Fig. 1D). The range of pelite-psammite compositions can be described as a two-component mixture:

$$C^{El}_{protolith} = \pi * C^{El}_{pelite} + (1 - \pi) * C^{El}_{psammite} \quad (3)$$

Where π is the mass fraction of pelite in a mixture. For the model, the compositions of psammite and pelite endmembers were calculated from samples with the highest and lowest content of SiO_2 (Table 1) selected from a suite of representative whole rock analyses (Greenfield, 1997; Spicer et al., 2004; Palya et al., 2011; Williams et al., 2018).

Melt composition model

The rocks from zones 3 to 5 diverge from sedimentary trends by extending to lower SiO_2, Na_2O+K_2O and higher Al_2O_3 and FeO+MgO (Fig. 1C and E) and these differences provide the basis for quantification of melt loss. For the purposes of this study, the most suitable melt compositions are from Spicer et al. (2004), who studied the melting in subsolidus rocks from the Mount Stafford. They observed that with micas as only sources of water, up to 60% melting occurred between 750 and 800 °C at 0.3 GPa. With increasing temperature, the water content in the melt decreased and FeO, MgO, TiO_2 increased. However, the melt compositions recalculated on an anhydrous basis (Table 1) show limited variability for major melt components, and SiO_2, Al_2O_3, Na_2O, and K_2O compose 97.4 wt% of the anhydrous granite in Table 1 (with 1.5 % 1σ for SiO_2). The variability is higher for TiO_2, FeO, and MgO, however, the relatively low content of these components (FeO+MgO+TiO_2 <3.6 wt%) results in only a minor contribution to the model uncertainty.

Residual minimization function and uncertainty estimation

The combination of equations 2 and 3 yields:

$$\pi * C^{El}_{pelite} + (1 - \pi) * C^{El}_{psammite} = \varphi * C^{El}_{melt} + (1 - \varphi) * C^{El}_{restite} \quad (4)$$

With constrained content of an element in endmember pelite, psammite and extracted melt for a restite sample, equation (4) has two unknowns: the melt fraction, φ and the proportion of pelite/psammite in the protolith, π. For any two elements equation 4 produces a set of two equations with two unknowns, which could be solved for φ and π. However, a more robust approach is finding φ and π, which produce the closest match to all major elements in a restite. Such values could be found by minimization of a sum of square residuals:

$$M(\pi, \varphi) = \sqrt{\sum_N (C^{El}_{restite\ measured} - C^{El}_{restite\ calculated}(\pi, \varphi))^2} \qquad (5)$$

Values of φ and π bringing M to a minimum could be considered as optimal estimates for a particular restite sample. The minimization function typically forms a single valley where values of φ and π give a similarly good match to the restite (Fig. 2A), with the valley showing negative slope for π>0.2 and an inversion of the slope of the trough for lower values. Petrologically, a negative sloped "valley" means that restite composition can be described almost equally well by an increase in the proportion of pelite or a decreased in melt loss, and distinguishing between these two variables withing some range is impossible. The minimization function remains low only for a narrow range of φ and π, and outside of the minimal trough, the residuals increase sharply (Fig. 2B). The reverse of the minimization function has key properties of the probability distribution function: it reaches maximum when the difference between calculated and measured compositions is minimal, simulates a normal distribution (Fig. 2B), and could be used for the quantification of the uncertainties of the melt loss and protolith parameters. The uncertainty then can be described in terms of error correlation (Ludwig, 2003) by error ellipses (Fig. 3).

DISCUSSION

Model reproducibility and consistency

The method presented in this study estimates protolith composition and melt loss by minimizing the residual function of the mass balance of 9 major oxides. The test of quality of the model for subsolidus metasedimentary rocks yields total quadratic difference of 0.9–3.6 wt% with average 1.8 wt%. Considering the analytical uncertainty and the variability of sedimentary rocks, this is an excellent match. For the Mount Stafford restites from zones 3 to 5, after excluding 10 samples with elevated residuals (some of which could have a different protolith) out of 93, the cumulative quadratic residuals for 83 samplesare 0.9–5 wt% with an average of 2.0 wt%. The reverse calculation of the composition of restites reproduces the major oxides composition of the restitic rocks with high accuracy (Fig. S1). For sample 2000-MST45 from Palya et al. (2011), Bartoli (2017) estimated by the multi-step melt-reintegration 18–21 wt% of melt loss with 0.5 wt% H_2O at the solidus. The estimate by our method provides ~80% of pelite component and 25 wt% melt loss (Table 1), with a good agreement with Bartoli's estimate.

Entrapment of restitic and peritectic phases could significantly affect the composition of magmas generated by anatexis (Chappell et al., 1987; Clemens et al., 2011). While a complete numerical treatment is beyond the scope of the present work, it should be noted that entrapment of restitic minerals makes melt more similar to the sedimentary protolith, and mass balance then results in higher estimates of melt loss fraction.

Restites typically have some fraction of unextracted melt (Rudnick, 1992; Stepanov et al., 2014), and some could have external melt injected (Brown, 2007; Morfin et al., 2013). A welcome bonus of the model is that melt addition can be treated as well as melt loss: the addition of granite to a rock is resolved by the same set of equations as above but results in a negative φ value (Fig. 3). Additionally, the bulk composition of a heterogeneous restite sample with an originally heterogeneous protolith is mathematically indistinguishable from the intermediate value of pelite fraction. Therefore, the mass balance model for melt loss from heterogeneous protolith is capable of robustly solving mass redistribution in migmatites.

Identity of highly restitic rocks

Melt loss was highly variable in Mount Stafford zones 3–55 and reaches a maximum of a remarkable 70 wt% for rocks with 40–70% pelite component in protolith (Fig. 3), whereas rocks with higher or lower pelite proportions show lower melt loss. This dependence can be attributed to the decrease of mica content with the increase of the psammite, and low quartz in pelite (Fig. S2B), while the optimal granite productivity occurs at intermediate π values (Fig. 3D–F).

Rocks with φ>0.5 are characterized by Al_2O_3>21 wt% and FeO+MgO>10%. Based on THERMOCALC modeling Wang et al. (2019) concluded that rocks with FeO+MgO>10 wt% could not be explained by loss of minor (<20 mol%) melt fraction. Our results suggest that those compositions can be efficiently reproduced by loss of significant amount of melt (φ>0.25) from psammitic to psammo-pelitic (π 0.1–0.58) protoliths.

Implications for the fluid regime during Mount Stafford anatexis

Our results are puzzling because they suggest that generally fertile psammo-pelitic rocks within the same metamorphic zones have undergone widely different melt losses. Moreover, 0.6–0.7 melt fraction for system closed for fluid loss above 500 °C require temperatures 850–900 °C (Fig. 3D) that is significantly higher than peak temperature estimates of 800–810 °C (Greenfield et al., 1998; White et al., 2003), and unattainable for rocks depleted in fluid after muscovite breakdown (Fig. 3E). This could be related to the availability of aqueous fluids, resulting from the metamorphic history of the rocks. The role of fluids in the generation of granitic melts during anatexis has been hotly debated. The principal endmember models are fluid-fluxed melting, where external sources of fluids or fluid retention cause extensive crustal melting (Buick et al., 2004; White et al., 2005; Weinberg and Hasalova, 2015; Collins et al., 2020), and “dehydration melting”, where melting is controlled by the breakdown of hydrous minerals (Vielzeuf et al., 1990; Clemens et al., 2011). Previous studies suggested that muscovite broke down at subsolidus conditions before the beginning of melting (White et al., 2003), and this is consistent with the decrease of LOI, a proxy for water content, from zone 1 to 2 (Stepanov et al., 2024). The gap between muscovite dehydration and the solidus (Fig. S2A) means that some rocks lost fluids before melting began, which reduced melt generation potential in these rocks, while simultaneously causing fluid enrichment in others, and resulted in high melt productivity (Fig. 3F). Therefore, the precise quantification of melt loss revealed a picture different from both endmember models: in the same complex, some rocks experienced fluid-fluxed melting, others experienced dehydration melting, while others were predehydrated and perhaps have not lost melt at all.

In this study, we obtained precise and robust estimates of the melt productivity of a restitic terrain. The 27–31 wt% average melt loss obtained for zones 3–5 (Fig. 3A–C) does not include sampling of the leucosomes remaining in the terrane, so the melt loss may be an underestimate. Nevertheless, considering that restite have density (calculated in Perple_X) 20% higher than melt, the volume of granite generated at Mount Stafford could have been close to the volume of the restites. These estimates suggest that intracrustal anatexis can result in efficient and voluminous transport of material from the middle to upper crust.

ACKNOWLEDGMENTS

Liu Penglei is thanked for valuable discussions, Renee Tamblyn for help with Perple_X calculations, and Michael Brown for valuable comments.

REFERENCES CITED

Bartoli, O., 2017, Phase equilibria modelling of residual migmatites and granulites: An evaluation of the melt-reintegration approach: Journal of Metamorphic Geology, v. 35, p. 919–942, doi:10.1111/jmg.12261.

Bea, F., 1989, A method for modelling mass balance in partial melting and anatectic leucosome segregation: Journal of Metamorphic Geology, v. 7, p. 619–628, doi:10.1111/j.1525-1314.1989.tb00622.x.

Brown, M., 2007, Crustal melting and melt extraction, ascent and emplacement in orogens: mechanisms and consequences: Journal of the Geological Society, v. 164, p. 709–730, doi:10.1144/0016-76492006-171.

Buick, I.S., Stevens, G., and Gibson, R.L., 2004, The Role of Water Retention in the Anatexis of Metapelites in the Bushveld Complex Aureole, South Africa: an Experimental Study: Journal of Petrology, v. 45, p. 1777–1797, doi:10.1093/petrology/egh033.

Chappell, B.W., White, A.J.R., and Wyborn, D., 1987, The Importance of Residual Source Material (Restite) in Granite Petrogenesis: Journal of Petrology, v. 28, p. 1111–1138, doi:10.1093/petrology/28.6.1111.

Clemens, J.D., Stevens, G., and Farina, F., 2011, The enigmatic sources of I-type granites: The peritectic connexion: Lithos, v. 126, p. 174–181, doi:10.1016/j.lithos.2011.07.004.

Collins, W.J., Murphy, J.B., Johnson, T.E., and Huang, H.-Q., 2020, Critical role of water in the formation of continental crust: Nature Geoscience, v. 13, p. 331–338, doi:10.1038/s41561-020-0573-6.

Collins, W.J., and Vernon, R.H., 1991, Orogeny associated with anticlockwise P-T-t paths: Evidence from low-P, high-T metamorphic terranes in the Arunta inlier, central Australia: Geology, v. 19, p. 835–838, doi:10.1130/0091-7613(1991)019<0835:OAWAPT>2.3.CO;2.

Forshaw, J.B., and Pattison, D.R.M., 2022, Major-element geochemistry of pelites: Geology, v. 51, p. 39–43, doi:10.1130/G50542.1.

Greenfield, J.E., 1997, Migmatite formation at Mt. Stafford, Central Australia:, https://ses.library.usyd.edu.au/handle/2123/10592 (accessed May 2024).

Greenfield, J.E., Clarke, G.L., and White, R.W., 1998, A sequence of partial melting reactions at Mt Stafford, central Australia: Journal of Metamorphic Geology, v. 16, p. 363–378, doi:10.1111/j.1525-1314.1998.00141.x.

Korhonen, F.J., Brown, M., Clark, C., and Bhattacharya, S., 2013, Osumilite–melt interactions in ultrahigh temperature granulites: phase equilibria modelling and implications for the *P–T–t* evolution of the Eastern Ghats Province, I ndia: Journal of Metamorphic Geology, v. 31, p. 881–907, doi:10.1111/jmg.12049.

Ludwig, K., 2003, User's manual for Isoplot 3.00. A geochronological Toolkit for Microsoft Excel: Berkeley, California, Berkeley Geochronology Center, Special Publication No. 4a.

MacRae, N.D., and Nesbitt, H.W., 1980, Partial melting of common metasedimentary rocks: A mass balance approach: Contributions to Mineralogy and Petrology, v. 75, p. 21–26, doi:10.1007/BF00371886.

Morfin, S., Sawyer, E.W., and Bandyayera, D., 2013, Large volumes of anatectic melt retained in granulite facies migmatites: An injection complex in northern Quebec: Lithos, v. 168–169, p. 200–218, doi:10.1016/j.lithos.2013.02.007.

Olsen, S.N., 1985, Mass balance in migmatites, *in* Ashworth, J.R. ed., Migmatites, Springer US, p. 145–179, http://link.springer.com/chapter/10.1007/978-1-4613-2347-1_4 (accessed June 2013).

Palya, A.P., Buick, I.S., and Bebout, G.E., 2011, Storage and mobility of nitrogen in the continental crust: Evidence from partially melted metasedimentary rocks, Mt. Stafford, Australia: Chemical Geology, v. 281, p. 211–226, doi:10.1016/j.chemgeo.2010.12.009.

Rubatto, D., Hermann, J., and Buick, I.S., 2006, Temperature and bulk composition control on the growth of monazite and zircon during low-pressure anatexis (Mount Stafford, central Australia): Journal Of Petrology, v. 47, p. 1973–1996.

Rudnick, R.L., 1992, Restites, Eu anomalies and the lower continental crust: Geochimica et Cosmochimica Acta, v. 56, p. 963–970, doi:10.1016/0016-7037(92)90040-P.

Rushmer, T., 2001, Volume change during partial melting reactions: implications for melt extraction, melt geochemistry and crustal rheology: Tectonophysics, v. 342, p. 389–405, doi:10.1016/S0040-1951(01)00172-X.

Spicer, E.M., Stevens, G., and Buick, I.S., 2004, The low-pressure partial-melting behaviour of natural boron-bearing metapelites from the Mt. Stafford area, central Australia: Contributions to Mineralogy and Petrology, v. 148, p. 160–179, doi:10.1007/s00410-004-0577-z.

Stepanov, A.S., Allen, C., Jiang, S.-Y., Zhukova, I.A., Duan, D.-F., and Wang, L., 2024, Geochemistry of metasedimentary restitic rocks and implications for melting conditions and metal potential of crustal felsic magmas: Earth-Science Reviews, p. 104799, doi:10.1016/j.earscirev.2024.104799.

Stepanov, A.S., Hermann, J., Korsakov, A.V., and Rubatto, D., 2014, Geochemistry of ultrahigh-pressure anatexis: fractionation of elements in the Kokchetav gneisses during melting at diamond-facies conditions: Contributions to Mineralogy and Petrology, v. 167, p. 1–25, doi:10.1007/s00410-014-1002-x.

Vernon, R.H., Clarke, G.L., and Collins, W.J., 1990, Local, mid-crustal granulite facies metamorphism and melting: an example in the Mount Stafford area, central Australia, *in* Ashworth, J.R. and Brown, M. eds., High-temperature Metamorphism and Crustal

Anatexis, Dordrecht, Springer Netherlands, The Mineralogical Society Series, p. 272–319, doi:10.1007/978-94-015-3929-6_11.

Vielzeuf, D., Clemens, J.D., Pin, C., and Moinet, E., 1990, Granites, Granulites, and Crustal Differentiation, *in* Vielzeuf, D. and Vidal, P. eds., Granulites and Crustal Evolution, Springer Netherlands, NATO ASI Series 311, p. 59–85, http://link.springer.com/chapter/10.1007/978-94-009-2055-2_5 (accessed November 2013).

Wang, W.-(RZ), Clarke, G., Daczko, N.R., and Zhao, Y., 2019, Modelling the partial melting of metasediments in a low-pressure regional contact aureole: the effect of water and whole-rock composition: Geological Magazine, v. 156, p. 1400–1424, doi:10.1017/S001675681800078X.

Weinberg, R.F., and Hasalova, P., 2015, Water-fluxed melting of the continental crust: A review: Lithos, v. 212–215, p. 158–188, doi:10.1016/j.lithos.2014.08.021.

White, R.W., Pomroy, N.E., and Powell, R., 2005, An *in situ* metatexite–diatexite transition in upper amphibolite facies rocks from Broken Hill, Australia: Journal of Metamorphic Geology, v. 23, p. 579–602, doi:10.1111/j.1525-1314.2005.00597.x.

White, R.W., Powell, R., and Clarke, G.L., 2003, Prograde metamorphic assemblage evolution during partial melting of metasedimentary rocks at low pressures: migmatites from Mt Stafford, Central Australia: Journal of Petrology, v. 44, p. 1937–1960, doi:10.1093/petrology/egg065.

White, R.W., Powell, R., and Halpin, J.A., 2004, Spatially-focussed melt formation in aluminous metapelites from Broken Hill, Australia: Journal of Metamorphic Geology, v. 22, p. 825–845, doi:10.1111/j.1525-1314.2004.00553.x.

Williams, M.A., Kelsey, D.E., Baggs, T., Hand, M., and Alessio, K.L., 2018, Thorium distribution in the crust: Outcrop and grain-scale perspectives: Lithos, v. 320–321, p. 222–235, doi:10.1016/j.lithos.2018.09.016.

Yakymchuk, C., and Brown, M., 2014, Consequences of open-system melting in tectonics: Journal of the Geological Society, v. 171, p. 21–40, doi:10.1144/jgs2013-039.

FIGURE CAPTIONS

Figure 1. A) Simplified map of the metamorphic zoning of the Mount Stafford complex, Australia (Greenfield et al., 1998), with metamorphic grade increasing eastward from zone 1 to 5. A description of the metamorphic zones is given in the text. B–E) bivariant major oxides plots for protolith (B and D) and rocks from different metamorphic zones (C and E). The crosses show compositions of pelite (PL), psammite (PS) and granite (G) endmembers, lines show change of composition of percentage of melt loss, and arrows show vectors of melt loss.

Figure 2. A) Map of square residuals sum function for sample 2000-MST45 (Palya et al., 2011) for the range of π and φ values. Note that the minimum is unique. B) A section across minimization function surface at constant π and the corresponding reverse minimization function.

Figure 3. A-C) The estimated π and φ values for the rocks of zones 3–5 of Mount Stafford. The ellipses show the minimization function “valley” in Fig. 2A and correspond to uncertainty of the values estimated for particular samples. D-F) The π and φ estimates for zones 3–5 superimposed on Perple_X calculated melt fractions at different temperatures across the psammite-pelite range (such as Fig. S2A, see details in SM) for three fluid regimes: (D) closed system before muscovite breakdown, fluid saturation ≈500 °C; (E) closed system after muscovite breakdown, fluid saturation ≈600 °C; and (F) addition of 2% H_2O to scenario shown in Fig. 3D. At the 800°C peak temperature estimated for Mount Stafford high melt loss observed for multiple samples can be explained only by addition of fluid, perhaps formed by muscovite breakdown in adjacent rocks.

Table 1. Whole-rock compositions of principal components for the mass balance model of psammite-pelite melting at Mount Stafford. Granite composition is average melt from (Spicer et al., 2004). σ is 1 standard deviation, Δ for difference between measured and calculated compositions. The rock compositions have been recalculated to 100% sum of 9 major oxides.

	Pelite	± σ	**Psammite**	± σ	**Granite**	± σ	2000-MST45		
N	**6**		**8**		**19**		Measured	Calculated	Δ
SiO2	57.0	0.98	79.6	1.8	75.7	1.5	55.73	56.42	0.69
TiO2	0.63	0.04	0.39	0.0	0.37	0.2	0.79	0.66	-0.13
Al2O3	25.4	0.59	11.0	1.2	12.6	1.0	24.71	26.1	1.38
FeO*	6.10	0.76	2.96	0.7	2.16	0.8	8.31	6.62	-1.69
MnO	0.07	0.02	0.08	0.1	0.04		0.09	0.09	0.00
MgO	2.28	0.18	1.06	0.3	0.13	0.2	2.98	2.69	-0.30
CaO	0.18	0.09	0.49	0.2	0.23	0.2	0.16	0.24	0.08
Na2O	1.50	0.70	1.19	0.5	1.93	0.6	1.55	1.28	-0.27
K2O	6.81	0.48	3.29	0.9	6.82	0.9	5.67	5.92	0.24

* all Fe recalculated to FeO.