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Hydrothermal activity at the ultraslow- to slow-spreading Red Sea Rift traced by chlorine in basalt

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ABSTRACT

Newly formed oceanic crust is initially cooled by circulating seawater, although where this occurs and over what regions fluids enter the crust is still unclear. Differences in the chlorine (Cl) concentrations between mid-ocean ridge basalt and seawater potentially make Cl a sensitive tracer for this hydrothermal circulation, allowing assimilation of hydrothermal fluids or hydrothermally altered crust by rising magma to be traced by measuring excess Cl in erupted lavas. Such excess Cl has been found in basalts from fast-spreading ridges (Cl concentrations up to 1200 ppm), but not so far on ultraslow- and slow-spreading ridges, where lower Cl values in the basalts (~50–200 ppm) make variations harder to measure. The Red Sea, with its relatively saline bottom water (40–42‰, cf. 35‰ salinity in open ocean water), the presence of axial brine pools (up to 270‰ salinity) and thick evaporite sequences flanking the young rift provides an ideal opportunity to study the incorporation of hydrothermal Cl at an ultraslow- to slow-spreading ridge (max. 1.6 cm/yr). Both absolute Cl concentrations (up to 1300 ppm) and ratios of Cl to elements of similar mantle incompatibility (e.g. K, Nb) are much higher in Red Sea basalts than for average ultraslow- and slow-spreading ridges. An origin of these Cl-excesses by seafloor weathering or syn-eruptive contamination can be excluded, as can mineral/melt fractionation during melting or crystallisation, based on trace element data. Instead, the incorporation of Cl at depth derived from hydrothermal circulation either by direct assimilation of hydrothermal fluids or through mixing of magma with partial melts of the hydrothermally altered crust is indicated. We see no influence of local spreading rate, the intensity of seafloor fracturing or the calculated depth of last crystal fractionation on Cl-excess. Seafloor areas with clear evidence of present or recent hydrothermal activity (brine pool temperatures above ambient, presence of hydrothermal sediments) always show Cl-excess in the local basalts and there is a positive correlation between Cl-excess and intensity of local volcanism (as determined by the percentage of local seafloor showing volcanic bathymetric forms). From this we conclude that Cl-excess in basalts is related to high crustal temperatures and hydrothermal circulation and so can be used to prospect for active or recently extinct hydrothermal systems. Samples recovered within 5 km of a seafloor evaporite outcrop show particularly high Cl-excesses, suggesting addition of Cl from the evaporites to the inflow fluids and that this may be the length scale over which hydrothermal recharge occurs.

1. Introduction

Hydrothermal circulation occurs at mid-ocean ridges where seawater interacts with the magma and rocks of the newly formed oceanic crust. The result is both heat- and element-transfer between the circulating fluids and the young oceanic lithosphere. Venting of the hot fluids on the seafloor creates habitats for chemosynthetic communities and can lead to the formation of seafloor massive sulphide deposits (e.g. Rona et al., 1986; Hannington et al., 2005). Hydrothermal activity occurs at all ridges, but the along-axis frequency of high-temperature hydrothermal venting appears to increase with spreading rate (e.g. Baker and German, 2004; Hannington et al., 2011).

Active high-temperature vent fields can be found by detecting their effluent in the overlying water column, while the detection of extinct fields, which are economically more interesting, is at best extremely challenging. As all hydrothermal venting is the seafloor expression of deeper hydrothermal fluid circulation and alteration, an alternative prospection method for finding active and extinct hydrothermal sites is to search for traces of hydrothermal alteration of the crust. The chlorine (Cl) contents of seawater and magma are vastly different and

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interaction with seawater may increase the Cl content of a magma depending on the extent of melt-fluid interaction, so the Cl contents of erupted basalts are sensitive to hydrothermal influence (Michael and Schilling, 1989; Gillis et al., 2003). Models from fast-spreading ridges indicate an interaction zone between water and magma at the roof of shallow axial magma chambers (Coogan et al., 2003; Gillis et al., 2003; France et al., 2009; France et al., 2010; France et al., 2013). Assimilation, by wall-rock stoping or partial melting, of hydrothermally altered oceanic crust with a high Cl-content (Barnes and Cisneros, 2012) will increase the Cl-content of a magma (e.g. Michael and Schilling, 1989; Michael and Cornell, 1998; Kent et al., 1999a,b; Gillis et al., 2003; le Roux et al., 2006; Wanless et al., 2011), irrespective of whether the altered rocks are linked to an active or an extinct hydrothermal system. Direct interaction of the magma with a hydrothermal fluid will similarly raise magmatic Cl-contents (e.g. Kendrick et al., 2013). So magmatic Cl-excess provides a magmatic marker of present and past hydrothermal circulation.

Chlorine-excess is best observed by the use of ratios of Cl against elements of similar magmatic incompatibility such as potassium (K) or niobium (Nb) (e.g. Michael and Schilling, 1989; Sun et al., 2007). Compared to <0.08 Cl/K for uncontaminated MORB (Michael and Cornell, 1998; Stroncik and Haase, 2004), very high Cl/K ratios (up to 0.7) and Cl contents (up to 1200 ppm) have been found at fast-spreading (>10 cm/yr) mid-ocean ridges (Michael and Schilling, 1989; Gillis et al., 2003; le Roux et al., 2006), at back-arc basins (Kent et al., 2002; Sun et al., 2007), in ophiolites (Coogan et al., 2002; Coogan, 2003), as well as in ocean island basalts (Kent et al., 1999a; Kent et al., 1999b). Compared to fast-spreading ridges, the Cl-concentrations of ultraslow-, slow-, and many intermediate-spreading ridges (e.g. Southeast Indian Ridge, Austral-Antarctic Discordance) are generally lower (Cl ~50–200 ppm; Cl/K 0.01–0.09 (e.g. Michael and Cornell, 1998)). Although the depth to which hydrothermal circulation penetrates the crust at ultraslow- and slow-spreading ridges is unknown, calculated depths of last equilibrium crystallisation suggest that magma at ultraslow- and slow-spreading ridges generally cools deeper (>3 kbar: ~10 km; Michael and Cornell, 1998) than at fast-spreading ridges (cf. McCaig and Harris, 2012). This is also deeper than the maximum depth of observed hydrothermal penetration in e.g. the Oman ophiolite (6–7 km in Gregory and Taylor, 1981; Bosch et al., 2004) although a greater depth for hydrothermal circulation at ultraslow- and slow-spreading ridges potentially could apply as faults can penetrate down to the lower crust (e.g. Harper, 1985; Cannat et al., 1991; Escartin et al., 2008). Together, low Cl/K and deep calculated crystallisation pressures were used by Michael and Cornell (1998) to suggest that hydrothermal Cl was generally not entering magmas at slower spreading rates due to the absence of a shallow and continuous magma lens. The only higher Cl/K ratios (up to 0.5) these authors found at slow-spreading ridges were from ridges with a thick crust (and hence presumably higher magma fluxes) and low calculated equilibrium crystallisation pressures, e.g. at the Alvin Mid-Atlantic Ridge (AMAR), Kolbeinsey and Reykjanes Ridges (Michael and Cornell, 1998).

The Cl measurements on which these conclusions were based have relatively large analytical uncertainty (15–20 ppm (Michael and Schilling, 1989; Jambon et al., 1995; Michael and Cornell, 1998)) and high detection limits, which makes it difficult to resolve the small variations which characterize Cl-excess in the intrinsically low magmatic Cl-concentrations of ultraslow- and slow-spreading MORB. This also applies to more recent Cl data (e.g. Jenner and O'Neill, 2012), which show several basalts from ultraslow- and slow-spreading ridges (including Red Sea samples) with slightly higher Cl/K values of up to 0.27. Hence, the conclusion that there is no Cl-excess in magmas from ultraslow- and slow-spreading ridges may be due, in part at least, to the low analytical precision of the measurements. In this paper we apply our newly developed, high precision (±1–2 ppm SD) low detection limit (<10 ppm) Cl measurement method (van der Zwan et al., 2012) to measure small variations in Cl contents between samples and so investigate whether the incorporation of hydrothermal Cl in magmas is also taking place at ultraslow to slow-spreading ridges. To improve our chances of observing excess Cl in magmas, we studied basalts from the ultraslow to slow-spreading (<1 to max. 1.6 cm/yr (Chu and Gordon, 1998)) Red Sea (Fig. 1) which has a high intrinsic seawater salinity (40–42‰ compared to 34.5‰ for average ocean water), (hot) saline brine pools in several within-axis bathymetric 'Deeps' (up to 270‰ salinity (e.g. Pierret et al., 2001; Gurvich, 2006)), and thick evaporite sequences flanking the active rift (e.g. Whitmarsh et al., 1974; Mitchell et al., 2010). All of these have the potential to increase the Cl contents of circulating hydrothermal fluids and, potentially, their alteration products, thus increasing the amount of Cl that could contaminate a magma. Note that although this situation is relatively unique for the present-day Earth, episodes of slow ocean opening cutting through evaporitic basins have occurred throughout Earth's history and the processes here may be relevant for older passive margins (e.g. Augustin et al., 2014). The magmatic system in the Red Sea Rift (RSR) does not appear to differ from those found on other ultraslow- and slow-spreading ridges, with basalts showing an average Na_8 of 2.3 (Na corrected for crystallisation to 8 wt.% MgO (Klein and Langmuir, 1987); PetDB database (Lehnert et al., 2000)) and similar major element characteristics (Michael and Cornell, 1998). This implies that any processes responsible for Cl-incorporation which can be identified in the Red Sea should be relevant to ultraslow- and slow-spreading ridges globally, even if elsewhere those processes are less easily traceable by Cl.

Our data show hydrothermal Cl contamination in the Red Sea basalts and we use them to propose areas where the crust may be hydrothermally altered and so where past and/or present hydrothermalism was or is most intense. This is important for the Red Sea as, so far, only fragments of massive sulphides and black smokers have been found at two places, but without any further indications of recent activity (Kebrit Deep; RSR 18°N; Blum and Puchelt, 1991; Monin et al., 1982), although three active hydrothermal systems have been inferred from high temperatures in brines and underlying sediments (Nereus Deep; Discovery-Atlantis II Deeps; Port Sudan Deep; e.g. Pierret et al., 2001; Gurvich, 2006; Swift et al., 2012). This number of occurrences is significantly less than expected over the ~1900 km length of the RSR (at this spreading rate we would expect ca. 10 systems (cf. Hannington et al., 2011)), which implies that there are more vent fields to be discovered.

1.1. Geological setting of the Red Sea rift

The Red Sea is a young ocean with a <60 km wide axial trough with maximum depths of 2850 m (Laughton, 1970; Fig. 1). Current spreading rates vary between <10 and ~16 mm/yr, with the highest rate near 18°N and slower spreading rates N and S of this (Chu and Gordon, 1998). The oldest (~5 Ma) basaltic seafloor present in the axial trough occurs at 17°N (e.g. Courtillot, 1982; Cochran, 1983), but seafloor spreading may have begun earlier (8–12 Ma; Izzeldin, 1987; Girdler, 1991; Augustin et al., 2014) with its magmatic products buried beneath the thick (up to 7 km) cover of Miocene evaporites and younger hemipelagic sediments which have flowed and slumped towards the axis from the basin margins.

The southern part of the RSR (south of 19.5°N) consists of a continuous and well-developed axial valley with parallel normal faulting and extensive volcanism (e.g. Roeser, 1975; Augustin et al., 2014). Further north, between 19.5°N and 23.5°N, basalt occurs in wide troughs and basins, called the Red Sea Deeps (e.g. Bäcker and Schoell, 1972; Pautot, 1983; Bonatti, 1985; Fig. 1). The Deeps are separated by shallower 'Inter-Trough Zones' where the sediment sequences from the flanks cover the axial trough (e.g. Tramontini and Davies, 1969; Searle and Ross, 1975; Izzeldin, 1989; Ligi et al., 2012; Augustin et al., 2014). North of 23.5°N the Deeps are more widely separated and basalt is only found in isolated places (Bannock, Mabahiss and Shaban Deeps; Bonatti et al., 1984; Pautot et al., 1984; Guennoc et al., 1988; Cochran, 2005, see also Fig. 1). The seafloor in the Deeps is more tectonized than that found in the southern Red Sea, although abundant evidence for volcanism is still present, including the localised occurrence of