

# Nd–Sr–Hf–O isotope provinciality in the northernmost Arabian–Nubian Shield: implications for crustal evolution

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Received: 16 June 2009 / Accepted: 25 November 2009 / Published online: 13 December 2009  
Springer-Verlag 2009

**Abstract** Multi-isotope study including whole-rock Nd–Sr, single zircon Hf, and SIMS  $\delta^{18}\text{O}$  analyses of zircons sheds light on magma sources in the northernmost Arabian–Nubian Shield (ANS) during ~820–570 Ma. Reconnaissance initial Nd and Sr isotope data for the oldest rocks can be modeled as the isotopically aged lower-crust (~820–740 Ma) reafirms previous estimates that intermediate crust in the ANS core (SE-region) and its early crustal evolution in this part of the shield involved northern, more contaminated ANS margins (NW-region). Nevertheless, Nd–Sr isotope enrichment of the lithospheric crust is displayed by post-collisional mantle is indicated by some basic magmas of the NW-calc-alkaline and alkaline igneous rocks of the ANS region displaying the most enriched Nd–Sr isotope compositions. Comparison of Nd and Hf depleted mantle model ages for rocks of the SE-region may indicate that crustal NW-region are characterized by lower  $\text{Nd}(T)$ – $\text{Hf}(T)$  and higher Sr and  $\delta^{18}\text{O}$  compared with rocks of the SE-region, at 1.1–1.2 Ga and were followed by crustal differentiation and the transition between the regions is gradual. With

Communicated by J. Hoefs.

Electronic supplementary material The online version of this article (doi:10.1007/s00410-009-0472-8) contains supplementary material, which is available to authorized users.

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**Keywords** Nd–Sr–Hf–O · Arabian–Nubian Shield · Magma-sources · Zircon

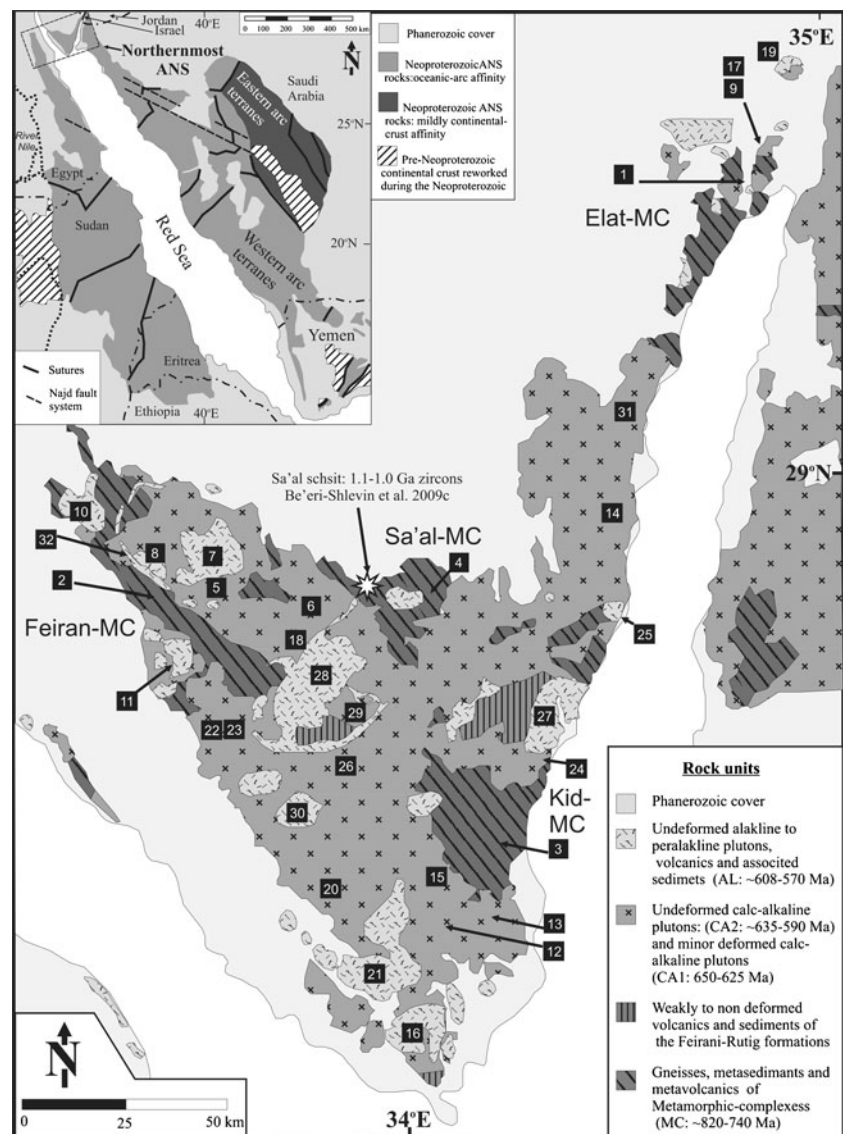
## Introduction

The formation of continental crust and the mode and rate of transfer of material from the mantle to the crust are long debated problems. The Arabian–Nubian Shield (ANS) forms one of the largest exposures of juvenile continental crust on Earth and as such comprises an excellent natural laboratory to examine these processes. The ANS forms the northern extension of the East African Orogen (Stein 1984) and was traditionally viewed as a collage of juvenile volcanic arc terranes and ophiolite remnants that were amalgamated during the assembly of the eastern part of Gondwana (Benton 1985; Meert 2003; Stein and Goldstein 1996; Stern 1994, 2002). Some reworked older crust in the ANS margins (Abdelsalam et al. 2002; Stacey and

Hedge 1984; Whitehouse et al 1998, 2001), but radiogenic  $^{18}\text{O}$  has been recently detected in Sinai (Be'eri-Shlevin et al. 2009). Evidence for recycling of crust, possibly Kibaran, (Fig. 1, inset) have traditionally been interpreted to show during the very early stages of ANS formation in Sinai. Lack of involvement of pre-Neoproterozoic crust in its (Egypt) and southern Israel is given by high  $^{18}\text{O}$  values of formation (Stein and Goldstein 1996; Stern 1994, 2002). zircon (Be'eri-Shlevin et al 2009b).

While this notion has remained generally accepted, This work focuses on the evolution of source reservoirs detailed isotope studies and evidence from zircon inherited by the multi isotope (Nd–Sr–Hf–O) composition patterns published over the past decade have provided new means to evaluate the evolution of the ANS of the ANS. Our main goal was to examine the sources of magmas through time in the northernmost segment of the ANS. Stoeser and Frost (2006) divided the Arabian segment of the post ~640 Ma magmas, which form vast post-collisional batholiths in this region (Fig.). This work complements the study of Be'eri-Shlevin et al (2009b) that the eastern arc terrane is characterized by more continental-like isotope compositions compared to the western arc terrane of oceanic affinity. Hargrove et al (2006) showed that a large terrane in the ANS includes inherited zircons of the  $\delta^{18}\text{O}$  in core and rims of zircon grains from the same pre-ANS age and Kibaran age (1.0–1.1 Ga) crustal terranes. Here, we use SIMS to further examine aliquots of zircon. Be'eri-Shlevin et al (2009b) detected  $^{18}\text{O}$  geographical provinciality in the post-collisional rocks of Sinai and southern Israel using multi-grain aliquots analyzed by Be'eri-Shlevin et al (2009b).

**Fig. 1** Simplified geological map of the study area in Sinai (Egypt) and southern Israel (modified after Eyal et al. 1980). Inset shows the exposures of the ANS in its Arabian and Nubian segments, as well as ophiolite sutures and the Najd fault system. Division of the Arabian shield into eastern- and western-arc terranes is from Stoeser and Frost (2006). Rock names: Metamorphic rocks: 1 Taba Gneiss, 2 Aliat Gneiss, 3 Kid schist, 4 Sa'al meta-basalt; Plutons: 5 Tubeina, 6 Ahdar, 7 Iqna, 8 Nasrin, 9 Yehoshafat, 10 Um-Bugma, 11 Serbal, 12 Lathi, 13 Mandar, 14 Zreir, 15 Sama, 16 Sharm, 17 Elat-Rehavim, 18 Sulaf, 19 Timna porphyritic granite, 20 Girgar, 21 Sahara, 22 Hibran-Miar quartz-diorite, 23 Hibran-Miar granite, 24 Shahira layered intrusion, 25 Rahba, 26 Umm-I-Fai, 27 Dahab, 28 Katharina, 29 Abu-K'sheib, 30 Umm-Shomer, 31 Malaha; Dikes: 32 Iqna dikes



The multi isotope study showed that the previously defined NW-SE O-isotope provinciality in post collisional complexes that make the comparison with island arcs magmatism in the northernmost ANS is confirmed by the exposed in more southern parts of the shield problematic radiogenic isotope ratios (Nd–Sr–Hf). The variable isotope

systems better constrain the source reservoirs of post A younger tectono-magmatic phase in this region ~640 Ma magmas in the northernmost ANS, and enable to re-evaluate the models associated with their formation. Our reconnaissance study of Nd and Sr isotope compositions from the older (~820–740 Ma) rocks in this region have reconfirmed previous estimates (Be'eri-Shlevin et al. 2009b) that rocks of this phase cannot be modeled as directly derived from a depleted mantle (DM) source. The construction of this dataset has also served as reference for the region.

## Geological setting and samples

### Geology of the study area

The northernmost ANS is exposed in southwest Jordan, the northern part of Saudi Arabia, southern Israel, Sinai (Egypt), and the northern part of the Eastern Desert of Egypt (Fig. 1). This part of the ANS includes only remnants of older metamorphic complexes (MCs) engulfed by younger mostly non-metamorphosed intrusions and volcano-sedimentary sequences.

The older MCs in this region include orthogneisses, and meta-sedimentary and -volcanic sequences, all metamorphosed at greenschist to amphibolite facies conditions (Ei-Shafei and Kusky 2003; Eyal et al. 1980; Jarrar et al. 1983; Katz et al. 1998; Matthews et al. 1989; Shimron 1984; Blasband et al. 2000; Abu-Alam and Stuewe 2009). In the study area of Sinai and southern Israel, these are grouped into four complexes: the Elat-, Feiran-, Sa'al-, and Kid-MCs (Fig. 1). Reported U–Pb ages for these rocks have consistently yielded ages within the range of 820–740 Ma (Be'eri-Shlevin et al. 2009b; Ei-Shafei and Kusky 2003; Jarrar et al. 2003; Kolodner et al. 2006; Kröner et al. 1990; 1994; Stern and Manton 1987), whereas whole-rock (WR) and mineral Rb–Sr ages show a larger scatter and range down to 610 Ma (Bielski 1982; Halpern and Tristar 1981). Ductile deformation is recorded by all MC rocks, but the number of deformation events and their ages are still debated (Cosca et al. 1999; Ei-Shafei and Kusky 2003; Eyal et al. 1991; Fowler and Hassa 2008; Shimron 1984). Rocks of the MCs in the northernmost ANS are thought to represent pre- to syn-collisional tectonic settings, and subduction-related mechanism for the genesis of the magmas is generally accepted (Abu El-Enen et al. 2004; Bentor 1985; Jarrar et al. 2003; Stern 1994). However, we note that there are several features including the lack of

deformed plutons of ~675–640 Ma (CA1 suite; Eyal et al. 1991; Jarrar et al. 2003; Kröner et al. 1990; Moussa et al. 2008; Stern and Hedge 1985; Stern and Manton 1987). Calc-alkaline plutonic magmatism lasted up to 590 Ma (Ali et al. 2009; Be'eri-Shlevin et al. 2009a), but from 635 to 625 Ma onwards plutons record little to no ductile deformation and are considered to represent a transition to post-collisional setting (CA2 suite; Be'eri-Shlevin et al. 2009a). Regarding the timing of deformation we note the important difference between the northernmost ANS and more southern ANS segments. In the Central Eastern Desert of Egypt as well as in NW Arabia deformation younger than 620 Ma generally associated with the Najd Fault System is common (Greiling et al. 1994).

of 630 Ma, considered by Katz et al. (2004) to have formed above a subduction zone, are the last vestiges of such processes in this region. Older members of the CA2 suite (635–620 Ma) are mainly small gabbro bodies, more abundant zoned plutons of quartz-diorite to granodiorite compositions, and some monzogranite. The peak of the CA2 tectono-magmatic phase occurred at 610–600 Ma when a vast amount of granodiorite to monzogranite plutons intruded the crust. A transition to alkaline magmatism (AL suite) in the northernmost ANS is associated with major features indicative of an extensional regime, such as uplift and intense erosion, and deposition of volcano-sedimentary units in closed basins at ~600 Ma (Bentor 1985; Beyth et al. 1994; Garfunkel 1999; Jarrar et al. 2003; Stern 1994; Stern and Hedge 1985). Avigad and Gvirtzman (2009) argued that the ~600 Ma uplift was due to lithospheric delamination. Nevertheless, older granites of this phase intruded prior to the formation of such basins at ~608 Ma (Be'eri-Shlevin et al. 2009a; Beyth et al. 1994) and magmatic activity lasted up to 580–570 Ma. Rocks of the AL suite include alkaline to peralkaline felsic (A-type granites) and rare mafic and intermediate plutons, as well as bimodal (mafic-felsic) alkaline and peralkaline volcanic rocks, and dike swarms (Bentor 1985; Beyth et al. 1994; Garfunkel 1999; Katzir et al. 2006, 2007a, b; Moghazi et al. 1998). The CA2 and AL rocks display distinct geochemical affinities (see compilations in Be'eri-Shlevin et al. 2009a). CA2 rocks are high-K calc-alkaline, whereas the felsic rocks of the AL suite display typical characteristics of A-type granites and some of the contemporaneous mafic rocks display features of alkaline basic magmas (Katzir

et al. 2007b; Kessel et al. 1998; Mushkin et al. 2003). Notwithstanding, Be'eri-Shlevin et al. (2009a) showed that the two magmatic suites overlapped for 20 Ma. Therefore, modeling magma petrogenesis of CA2 and AL rocks in terms of representing different tectonic settings is unrealistic.

### Sample description

Two sets of samples were used in this study (for sample locations see Table A1 available online in Electronic supplementary material):

For preliminary isotope characterization of the older ~820–740 Ma MCs of Sinai and southern Israel, we used a small set of samples. The U–Pb dated Aliat (~800 Ma; unpublished data reported in Be'eri-Shlevin et al. 2009b) and Taba (~780 Ma; Kröner et al. 1990) gneisses of tonalite to granodiorite composition represent the Elat- and Feiran-MCs, respectively (Fig. 1). Two other rocks have not been dated by U–Pb studies. The Sa'al meta-basalt represents the Sa'al-MC and is believed to be older than the ~730 Ma Rb–Sr age (Bielski 1982) given for the Sa'al-MCs, which probably dates cooling after peak metamorphism. A meta-pelitic schist (Kid schist) from the Umm-Zarik formation (Shimron 1984) represents the Kid-MC. Although Rb–Sr ages of 610 Ma were reported for the Kid-MC (Bielski 1982; Halpern and Tristar 1981), it has been shown recently that formation and metamorphism of this complex predated 630 Ma and is more plausibly associated with the ~820–740 Ma tectono-magmatic phase (Be'eri-Shlevin et al. 2009a).

A much larger set of samples was used for isotope characterization of the CA2 and AL suites, also providing good geographical spread within this study area. All rock samples in this set have been previously analyzed for O isotope compositions in zircon using multi-grain analyses (Be'eri-Shlevin et al. 2009b). SIMS U–Pb ages for these rocks are reported in Be'eri-Shlevin et al. (2009a, b). For more detailed descriptions of these rocks and their zircon the reader is referred to these two publications.

### Analytical techniques

#### Oxygen isotope analyses

Oxygen isotope ratios of various domains within the zircon grains were measured using a CAMECA IMS-1270 at the NORDSIM Facility (Swedish Museum of Natural History, Stockholm, Sweden). Analytical procedures generally follow Nemchin et al. (2006). For zircons of 19 U–Pb dated samples of CA2 and AL plutons (Be'eri-Shlevin et al. 2009a) ion-microprobe  $\delta^{18}\text{O}$  analysis was mainly performed on the same U–Pb dated spots or in very close proximity. Prior to ion-microprobe  $^{18}\text{O}$  analysis the U–Pb analysis spots were removed from the zircon by minor polishing followed by recoating with  $\sim 30$  nm of gold. Complementary ion-microprobe  $^{18}\text{O}$  analyses for these 19 dated samples were performed on undated zircon domains or grains that were characterized by similar structures in CL images. The following procedures were fully automated: (a) pre-sputtering with a raster of  $25\ \mu\text{m}$ ; (b) eld aperture, entrance slit, and mass centering using the  $^{16}\text{O}$  signal; and (c) data acquisition (240 s) using two Faraday detectors in the multi-collector system operating at a common mass resolution of ca. 2,500. Results from the unknowns are presented in Table A2 (available online in Electronic supplementary material) with  $\sigma$  uncertainties. During seven different sessions, a total of 243 analyses of unknowns were performed bracketed with analyses of the Geostandard zircon 91500 after every four to five unknown analyses. Data were normalized to measurement of the standard zircon (91500) with an assumed  $\delta^{18}\text{O}$  value of 9.86‰. Drift correction was applied by fitting a linear regression through the standard data, and within the seven analysis sessions this varied from 0 to 0.001‰ per run. The external variation of the zircon standard varied between  $\pm 0.2$  and  $\pm 0.4$ ‰ ( $2\sigma$ ) during all seven sessions. Errors for specific analyses calculated as a quadratic sum of the internal precision using observed errors and the external precision of the zircon standard ranged from  $\pm 0.4$  to  $\pm 0.7$ ‰ ( $2\sigma$ ) for more than 98% of the samples (Table A2; available online in Electronic supplementary material).

constants ( $\lambda_{\text{Rb}}$ ,  $\lambda_{\text{Sm}}$ ) used for age calculations are Results

0.0142 Ga<sup>-1</sup> for <sup>87</sup>Rb and 0.00654 Ga<sup>-1</sup> for <sup>147</sup>Sm. Two

different DM Nd model ages were calculated using the ion-microprobe  $\delta^{18}\text{O}$  results

following definitions: (a)  $T_{\text{DM1}}$  is a one-stage model age

calculated according to DePaolo (1981), whereby the

evolution of the DM is given by  $\epsilon_{\text{Nd}}(\text{T})_{\text{DM}} = 0.257^{T/2} -$

$3T + 8.5$  and iteration versus the  $\epsilon_{\text{Nd}}(\text{T})$  of the sample

provides  $T_{\text{DM1}}$  (T in Ga); (b)  $T_{\text{DM2}}$  is a one-stage model

calculated assuming a linear Nd isotopic growth of the DM

from  $\epsilon_{\text{Nd}} = 0$  at 4.56 to +10 at the present (Goldstein

et al. 1984:  $T_{\text{DM2}} = 1/\lambda \times \ln\{1 + [({}^{143}\text{Nd}/{}^{144}\text{Nd})_{\text{S}} -$

$({}^{143}\text{Nd}/{}^{144}\text{Nd})_{\text{DM}}]/[({}^{147}\text{Sm}/{}^{144}\text{Nd})_{\text{S}} - ({}^{147}\text{Sm}/{}^{144}\text{Nd})_{\text{DM}}]\}$ ,

where s denotes sample, DM depleted mantle and  $\lambda$  decay

constant of <sup>147</sup>Sm.

### Single zircon Hf isotope analyses

Single zircon grains from rocks of the CA2 and AL suites

were analyzed for Lu–Hf isotopic composition by MC-

ICP-MS at the Ecole Normale Supérieure de Lyon. Hf

isotope analysis followed careful characterization of the

grains using CL images, ion-microprobe U–Pb dating and

$\delta^{18}\text{O}$  analyses of the same domains as dated by U–Pb

(Be'eri-Shlevin et al. 2009a). The main target was rep-

resentative, homogenous grains in terms of age and

Representative zircons were hand-picked out of the epoxy

ion-microprobe mounts, examined under an optical

microscope to ensure the removal of most epoxy, and

leached in consecutive batches of cold concentrated nitric

and hydrofluoric acids. Dissolution and analytical proce-

dures follow Blichert-Toft and Albarede (2008) except

that Pb isotopes were not measured. Age corrections due

to the radiogenic ingrowth of <sup>176</sup>Hf since zircon crystal-

lization (~770–580 Ma) vary from 8 to 96 ppm. Hf(T)

was calculated using present-day values for

<sup>176</sup>Hf/<sup>177</sup>Hf<sub>(CHUR)</sub> of 0.282772 and <sup>176</sup>Lu/<sup>177</sup>Hf<sub>(CHUR)</sub> of

0.0332 (Blichert-Toft and Albarede 1997) and the <sup>176</sup>Lu

decay constant  $\lambda_{\text{Lu}}$  of 0.01867 Ga<sup>-1</sup> (Scherer et al.

2001). DM model age calculations were done using a two-

stage model whereby the <sup>176</sup>Lu/<sup>177</sup>Hf of the zircon is used

to first calculate back to the crystallization age determined

by U–Pb (Be'eri-Shlevin et al. 2009a) and then calculat-

ing back further to the DM using a fixed <sup>176</sup>Lu/<sup>177</sup>Hf

value for the given theoretical precursor reservoir. Vary-

ing the <sup>176</sup>Lu decay constant within 0.0186–0.187 Ga

(Söderlund et al. 2004) did not change the calculations by

more than a few million years. In this study three DM

ages were calculated using the following <sup>176</sup>Lu/<sup>177</sup>Hf values:

(a)  $T_{\text{DM1}}$ : <sup>176</sup>Lu/<sup>177</sup>Hf = 0.0093 corresponding to a felsic

crust (Amelin et al. 1999); (b)  $T_{\text{DM2}}$ : <sup>176</sup>Lu/<sup>177</sup>Hf = 0.015

corresponding to an average crust (GERM); and (c)

$T_{\text{DM3}}$ : <sup>176</sup>Lu/<sup>177</sup>Hf = 0.022 corresponding to a mafic

crust (Vervoort and Patchett 1996).

The results of ion-microprobe  $\delta^{18}\text{O}$  analyses of zircon grains [ $\delta^{18}\text{O}(\text{Zrn})$ ] from 13 CA2 plutons (~635–590 Ma) and 7 AL plutons (~608–580 Ma) are presented in Fig. 2.

Figure 2 shows that most zircon populations are homogenous in terms of  $\delta^{18}\text{O}(\text{Zrn})$ , and average values are comparable to the multi-grain  $\delta^{18}\text{O}(\text{Zrn})$  values reported for these rocks (white vertical bars in Fig. 2g, Be'eri-Shlevin et al. 2009b). Significant scatter of ion-microprobe  $\delta^{18}\text{O}(\text{Zrn})$  data and/or difference between average ion-microprobe and multi-grain data are observed in 6 of the 20 zircon populations (Fig. 2g).

However, such deviations mostly stem from the over-representation in the ion-microprobe analysis of xenocrystic zircons, which form minor components in these populations (Ali et al. 2009, Be'eri-Shlevin et al. 2009a), and from targeting high-U zircon domains, which were probably affected by post-crystallization oxygen exchange with late fluids. A clear case demonstrating the latter effect is the Inna pluton, where CL-dark, high-U, metamict zircons are characterized by variable and high  $\delta^{18}\text{O}(\text{Zrn})$  values of 6.1 to 14‰, whereas CL-bright, low-U zircon domains define a much more restricted  $\delta^{18}\text{O}(\text{Zrn})$  range from 6.2 to 7‰ comparable with the multi-grain value (Fig. 2c). Omitting the  $\delta^{18}\text{O}(\text{Zrn})$  values of xenocrystic and high-U zircons from the ion-microprobe average, most plutons (17 of 20) show good agreement between ion-microprobe and multi-grain  $\delta^{18}\text{O}(\text{Zrn})$  data indicating that eliminating such outliers is justified. If unexplained outliers from the Hibran-Miar, quartz-diorite, and Malaha plutons [three zircon grains yielding low  $\delta^{18}\text{O}(\text{Zrn})$  for both cores and rims] are also excluded from ion-microprobe  $\delta^{18}\text{O}(\text{Zrn})$  averages, the majority of zircon populations shows a very good fit with the multi-grain data. It follows that the corrected ion-microprobe  $\delta^{18}\text{O}(\text{Zrn})$  average values generally reiterate the multi-grain  $\delta^{18}\text{O}(\text{Zrn})$  values and that these represent pristine magmatic ratios of the rocks in question. Finally, zircons from the Sahara pluton show consistently lower ion-microprobe  $\delta^{18}\text{O}(\text{Zrn})$  values (Fig. 2c) relative to the multi-grain value of  $5.14 \pm 0.05\%$ , measured by Be'eri-Shlevin et al. (2009b) for zircon separates of three different rock samples from this pluton, including zircons used in this study. Since both ion-microprobe and multi-grain  $\delta^{18}\text{O}(\text{Zrn})$  analyses yielded internally consistent values and considering the good agreement between the two analysis types for other plutons, such disagreement seen in the zircons of the Sahara pluton remains unexplained.

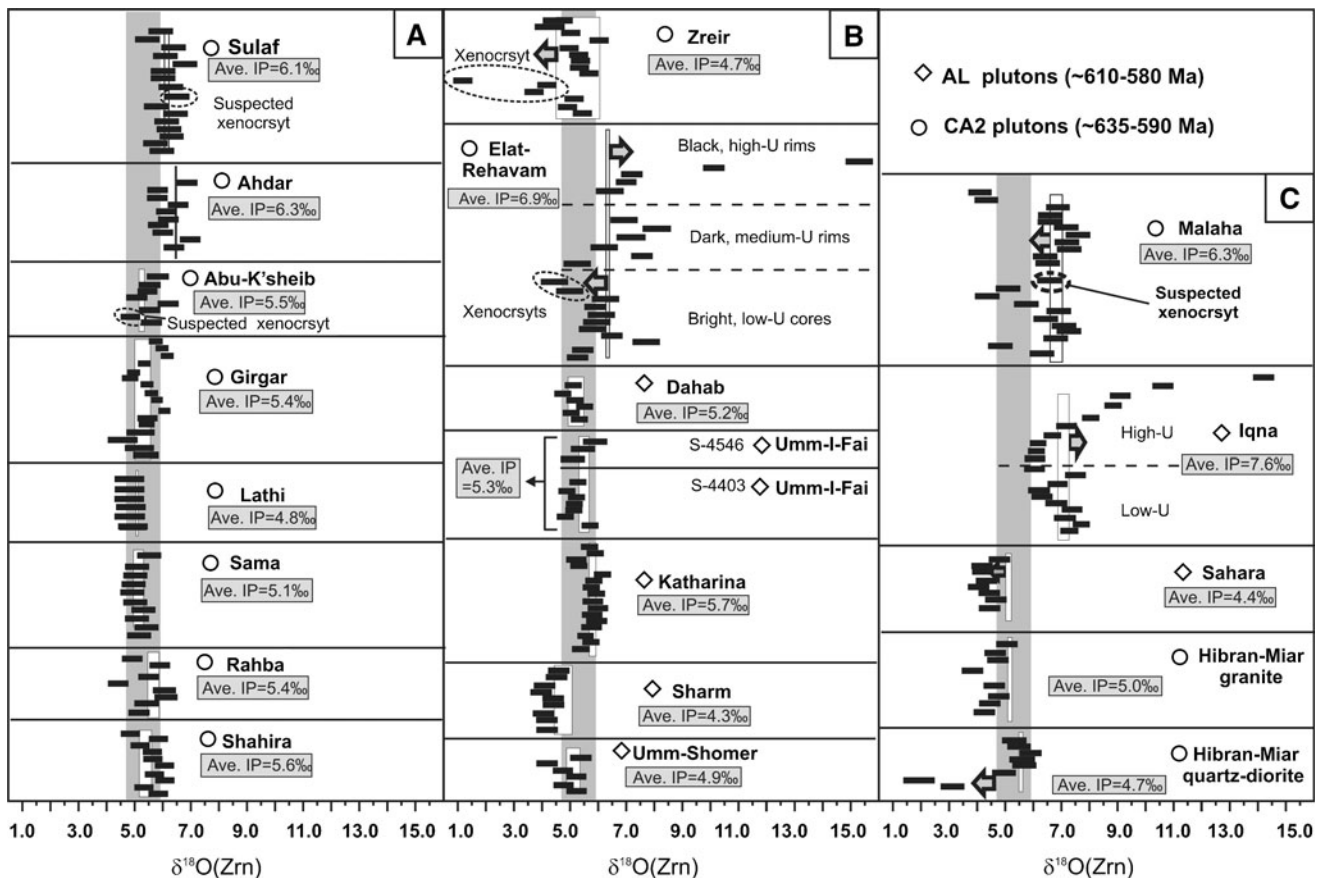


Fig. 2 Ion-microprobe  $\delta^{18}\text{O}$  analyses in zircons of the CA2 and AL white vertical bars ( $2\sigma$ ). Continuous vertical gray bars represent the  $\delta^{18}\text{O}$  mantle range for zircon ( $5.3 \pm 0.6\text{‰}$ ; Valley et al. 1998) Shlevin et al. 2009b) for the same rocks are shown for reference as

It is evident that in most zircon populations average Whole-rock Sr–Nd isotope results values for cores and rims (excluding inherited material, high-U zones, and outliers) generally define a 1:1 relationship implying that zircon crystallization commenced and continued within homogeneous magmatic environments (Fig. A1; available online in Electronic supplementary material). This feature is consistent among rocks of variable SiO<sub>2</sub> contents and affiliation with either CA2 or AL suites. Thus, intrusion of mantle-like parent magmas and their contamination by wallrock as suggested by Be’eri-Shlevin et al. (2009b) is considered highly unlikely. Rather, the results suggest that high- $\delta^{18}\text{O}$  CA2 and AL rocks represent melting of high- $\delta^{18}\text{O}$  sources.

The  $\delta^{18}\text{O}(\text{Zrn})$  results for the Elat-Rehavam pluton from southern Israel plot as prominent outliers in Fig. 2. These results, as well as the  $\delta^{18}\text{O}(\text{Zrn})$  data of xenocrysts, will be discussed elsewhere. Nevertheless, it is important to note that this pluton represents a minor intrusion (a few square kilometers) compared to other plutons investigated in this work (40 to >200 km<sup>2</sup>) and thus its significance in terms of magma production during this time is limited.

Table 1 Rb–Sr and Sm–Nd data for rocks of the study area in Sinai (Egypt) and southern Israel

Rock	Sample	Age (Ma)	Rb (ppm)	$^{87}\text{Rb}/^{86}\text{Sr}^b$	$^{87}\text{Sr}/^{86}\text{Sr}^c$	$\text{Si}^d$	Sm (ppm)	Nd (ppm)	$^{147}\text{Sm}/^{144}\text{Nd}^e$	$^{143}\text{Nd}/^{144}\text{Nd}^f$	$\epsilon_{\text{Nd}}(T)$	DM model ages	
												$T_{\text{DM}1}^g$	$T_{\text{DM}2}^h$
Metamorphic complexes													
Taba gneiss	YE-20	780	62.0	253	0.705	0.711204 (22)	0.7036	3.47	16.8	0.1248	0.512463 (7)	3.6	0.99
Allait gneiss	S-2870	800	60.9	340	0.516	0.708752 (22)	0.7029	2.17	10.6	0.1241	0.512429 (13)	3.4	1.04
Kid schist	SK-75-5a	800	81.1	114	2.053	0.721569 (22)	0.6981, 0.7028, 0.59	15.7	0.1382	0.1382	0.512524 (12)	3.8, 2.6, 0.05	1.26
Sa'ial meta-basalt	S-1252	800	5.7	190	0.086	0.704979 (19)	0.7040, 0.70412, 0.15	7.00	0.1862	0.1862	0.512792 (13)	4.1, 4.0	
CA2 + AL suits NW-region													
CA2: Tubeina, monzogranite	S-2864	600	117.7	161	2.102	0.720759 (6)	0.7028	4.27	21.2	0.1216	0.512541 (8)	3.9	0.83
CA2: Ahdar, granodiorite	S-2911	610	108	268	1.158	0.713084 (32)	0.7030	4.1	24.8	0.1001	0.512439 (8)	3.7	0.81
AL: Iqna, syenogranite	YS-503	580	266	8.21	93.34	1.484009 (24)		10.8	42.7	0.1532	0.512575 (10)	2.0	1.17
AL: Nasrin, monzogabbro	S-2644	600	14.3	232	0.1781	0.705969 (24)	0.7044	3.7	13.6	0.1662	0.512575 (8)	1.1	0.99
AL: Yehoshafat, Afs-granite	YE-10	605	111.3	43	7.400	0.764164 (22)		3.82	21.7	0.1063	0.512449 (14)	3.3	0.85
AL: Um-Bugma syenogranite	YS-400	595	325.8	27	34.807	0.992922 (25)		8.93	30.5	0.1771	0.512665 (12)	2.0	1.07
AL: Iqna (dike), monzonite	S-649	590	120.7	308	1.128	0.713038 (21)	0.7036	3.63	19.7	0.1113	0.512430 (10)	2.4	0.91
AL: Serbal, Afs-granite	S-2504	605	99.5	16	18.057	0.851803 (23)		6.11	30.6	0.1208	0.512541 (12)	3.5	0.82
CA2 + AL suites SE-region													
CA2: Lathi, monzogranite	S-1610	607	51.0	585	0.2511	0.705184 (27)	0.7030	7.2	36.4	0.1193	0.512602 (14)	5.3	0.71
CA2: Mandar, monzogranite	S-1709	600	119	107	3.201	0.732642 (35)	0.7053	5.9	34.7	0.1028	0.512544 (12)	5.4	0.69
CA2: Zreir, Qtz-diorite	S-3440	630	4.28	885	0.0139	0.704267 (28)	0.7041	2.4	10.5	0.1367	0.512664 (32)	5.3	0.76
CA2: Sama, Qtz-monzonite	S-3704	609	87.9	328	0.7715	0.709534 (31)	0.7028	5.5	30.9	0.1072	0.512544 (10)	5.1	0.72
AL: Sharm, Afs-granite	S-1518	596	106	38.2	7.559	0.762320 (20)		12.8	57.3	0.1352	0.512656 (10)	5.0	0.75

Afs: Alkali feldspar, Qtz: Quartz

<sup>a</sup> U–Pb ages: normal text, Rb–Sr ages: in italics, inferred ages based on other geological constraints underlined. Details regarding the age of the Serbal schist and the Kid Andalusite schist are found in the text

<sup>b</sup> Uncertainty for  $^{87}\text{Rb}/^{86}\text{Sr}$  and for  $^{147}\text{Sm}/^{144}\text{Nd}$  values is 0.5%

<sup>c</sup> NBS standard yielded  $^{87}\text{Sr}/^{86}\text{Sr} = 0.056508(77)$  and  $^{87}\text{Sr}/^{86}\text{Sr} = 0.710259(79)$  ( $n = 73$ ). Errors [in parentheses: (0<sup>+</sup>–(6))] are  $\pm 2\sigma$  of the mean

<sup>d</sup>  $\text{Si}^d$  = Initial  $^{87}\text{Sr}/^{86}\text{Sr}$  calculated only for samples with  $\text{Rb}/^{86}\text{Sr} < 4$

<sup>e</sup> Calculated for  $T = 730$  Ma

<sup>f</sup> Calculated for  $T = 640$  Ma

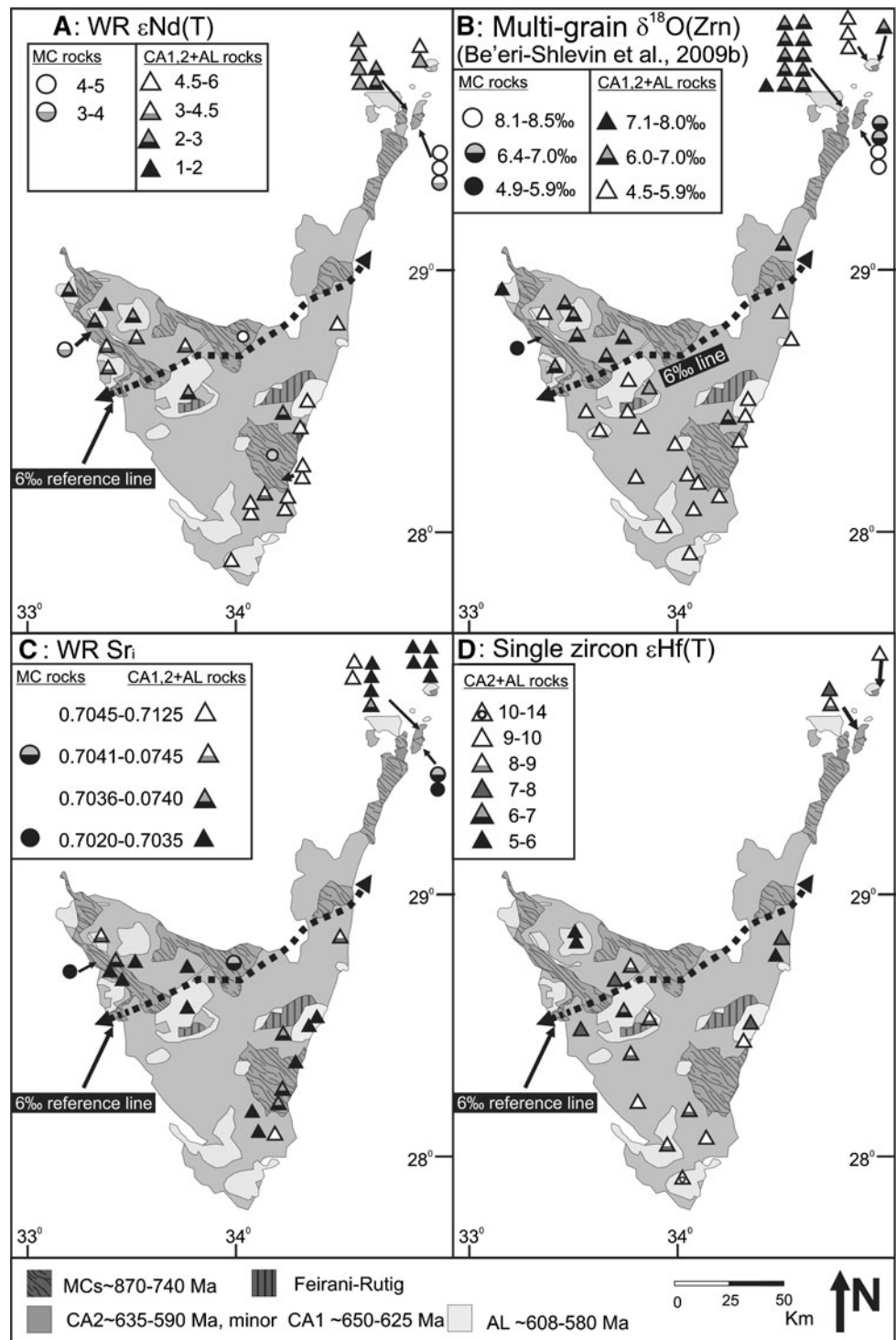
<sup>g</sup> Nd Göttingen in-house standard corresponding to the La Jolla  $^{143}\text{Nd}/^{144}\text{Nd}$  value of 0.511804(33) and  $^{143}\text{Nd}/^{144}\text{Nd} = 0.348389(20)$ . Errors [in parentheses: (0<sup>+</sup>–(6))] are  $\pm 2\sigma$  of the mean

<sup>h</sup> Depleted mantle model ages calculated only for samples with  $^{147}\text{Sm}/^{144}\text{Nd} < 1.65$ . Model ages of samples with  $0.44^{147}\text{Sm}/^{144}\text{Nd} < 0.165$  are underlined

<sup>i</sup>  $T_{\text{DM}1}$  = single stage model based on DePaolo (1981)

<sup>j</sup>  $T_{\text{DM}2}$  = single stage model based on Goldstein et al. (1984)

Fig. 3 Simplified geological maps of the study area showing the geographical variability in the isotope compositions of the analyzed rocks. Data include our own as well as data from the literature given as average values for each rock unit. For references see text



age range of 800–780 Ma are concordant with previously reported values for other 820–740 rocks from the north-bance of the Rb–Sr system at this time. Rocks of the younger (~635–580 Ma) tectono-magmatic phases display larger variability of  $\text{Sr}_i$  and  $\epsilon\text{Nd}(T)$  values as compared with the older rocks (Table 1). Of 13 CA2 and AL samples analyzed in this study, 4 are characterized by  $^{87}\text{Rb}/^{86}\text{Sr} > 4$  precluding a reliable



calculation of their  $S_r$  values (Stoeser and Froneo 2006 and references therein). For nine other samples,  $S_r$  ranges from 0.7028 to 0.7053, while  $\epsilon_{Nd}(T)$  for all samples varies between +1.1 and +5.4. We note that sample S-1709 (Mandar pluton) with  $S_r = 0.7053$  [but  $\epsilon_{Nd}(T)$  of +5.4] is characterized by relatively high  $^{87}Rb/^{86}Sr$  (3.2). Excluding sample S-1709,  $S_r$  of the other eight samples ranges from 0.7028 to 0.7044.

Figure 4 shows an  $\epsilon_{Nd}(T)$  versus age diagram, where results of the present study are plotted along with data from the literature (Azer et al. 2009; Beyth et al. 1994; Katzir et al. 2007a, b; Moghazi et al. 1998; Mushkin et al. 2003; Stein and Goldstein 1996). The 820–740 Ma rocks of the northernmost ANS define a restricted  $\epsilon_{Nd}(T)$  field, which falls slightly below the DM evolution curve of DePaolo (1981). The field representing the Nd evolution trend of the Sa'al schist from 1.0 Ga (Be'eri-Shlevin et al. 2009) plots significantly below the  $\epsilon_{Nd}(T)$  field for the other metamorphic rocks and, hence, represents very different material.

It is evident from Fig. 4 and from the  $\epsilon_{Nd}(T)$  locality map (Fig. 3a) that the spatial distribution of  $\epsilon_{Nd}(T)$  for CA2 and AL rocks is not random and a threefold division

can be applied. Within the SE-region located along the Sinai southern coast of Elat/Aqaba Gulf the CA2 and AL rocks display similar  $\epsilon_{Nd}(T)$  of +5.0 to +5.6. NW of this region, CA2 and AL rocks display a wider and lower range of  $\epsilon_{Nd}(T)$  values (+1.1 to +4.1) and the lower part of this range is exclusively defined by AL rocks. This large region can be further subdivided based on comparison with data from Be'eri-Shlevin et al. (2009b) (Fig. 3b). The NW-region is defined to the NW of the  $\delta^{18}O = 6\text{‰}$  reference line of Be'eri-Shlevin et al. (2009b). The middle-region is defined to the SE of this line and to the NW of the SE-region. Thus, although rocks of the middle-region are characterized by  $\epsilon_{Nd}(T)$  similar to the NW-region their mantle-like  $\delta^{18}O$  is more similar to the SE-region.

Data from Jordan and the northern Eastern Desert of Egypt, which border the study area to the NE and SW, respectively, comply well with the general geographical division defined above and correlate with the SE-region (Moussa et al. 2008) and NW-region (Jarrar et al. 2003), respectively, even after correction for the 105 km strike slip displacement along the Dead Sea transform (Fig. inset). A clear outlier to the division outlined above is the  $\epsilon_{Nd}(T)$  of the Timna porphyritic granite from southern Israel (+5.6 to +5.9; Beyth et al. 1994; TPG in Fig. 4), which does not conform with the location of this rock in the NW-region of lower  $\epsilon_{Nd}(T)$  values.

A  $S_r$  locality map including data from this study and the literature (Azer et al. 2009; Beyth et al. 1994; Katzir et al. 2007a, b; Moghazi et al. 1998; Mushkin et al. 2003; Stein and Goldstein 1996; Stern and Mantor 1987) is shown in Fig. 3c.  $S_r$  values corresponding to samples with  $^{87}Rb/^{86}Sr \geq 4$  are not shown in Fig. 3c. Calculated  $S_r$  values do not correlate with the  $\epsilon_{Nd}(T)$ - and  $\delta^{18}O(\text{Zrn})$ -based NW-SE geographical division of CA2 and AL rocks, and are somewhat variable in both the SE- and NW-regions. On the  $S_r$  versus  $\epsilon_{Nd}(T)$  diagram (Fig. 5) the NW-region rocks display a trend of increasing  $S_r$  with decreasing  $\epsilon_{Nd}(T)$ . Samples from the SE-region mostly display constant and low  $S_r$  which is associated with elevated  $\epsilon_{Nd}(T)$  of a restricted range. Nevertheless, some samples from this region display higher  $S_r$  with no decrease in  $\epsilon_{Nd}(T)$ . We note that all samples with very high  $S_r$  plotted in Figs. 3c and 5 are from dikes exposed in southern Israel (NW-region) (Katzir et al. 2007b). These authors discussed hydrous alteration of alkali feldspars as a possible mechanism to explain the highly variable and unusually high  $S_r$  in these rocks. As discussed above, the mildly high  $S_r$  of the Mandar pluton (No. 13 in Fig. 5;  $S_r = 0.7053$ ) is associated with a relatively high  $^{87}Rb/^{86}Sr$  (3.2) and thus this result may not be significant. We, thus, conclude that apart from a few outliers,  $S_r$  also displays similar NW-SE geographical variability as depicted by the Nd and  $\delta^{18}O(\text{Zrn})$  data.

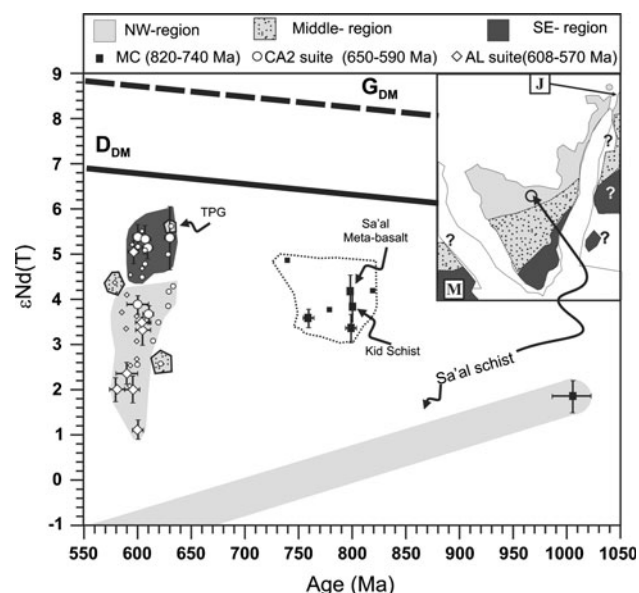


Fig. 4  $\epsilon_{Nd}(T)$  versus age diagram showing the results of the present study (large symbols) as well as literature data (small symbols). For references see text. CA2 samples within both the SE- and NW-regions display large variability in chemical composition (e.g.,  $SiO_2$  60–75 wt%) but only small variability in  $\epsilon_{Nd}(T)$ . The Sa'al meta-basalt and the Kid schist are plotted at an assumed age of 800 Ma. The field represents of the Sa'al schist represent the evolution from the maximum age of this rock at 1.0 Ga (Be'eri-Shlevin et al. 2009).  $G_{DM}$  and  $D_{DM}$  are the depleted mantle evolutionary curves according to models of Goldstein et al. (1984) and DePaolo (1981) respectively. TPG Timna porphyritic granite. Data from Jarrar et al. (2003) for SW Jordan and from Moussa et al. (2008) for the Northern Eastern Desert of Egypt are designated by J and M respectively in the inset map

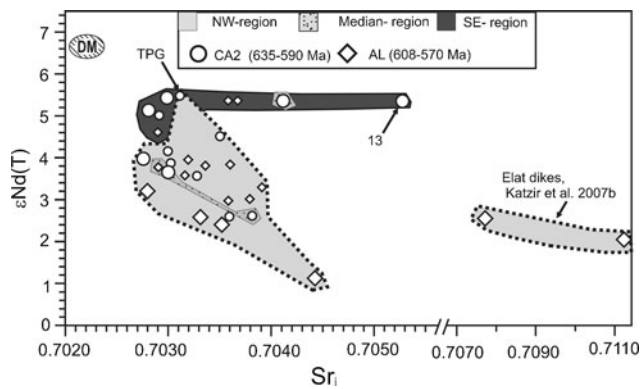


Fig. 5  $Sr_i$  versus  $\epsilon Nd(T)$  diagram showing a compilation of data from the northernmost ANS including data from Jordan (Jarrar et al. 2013). Most samples show geographical provinciality similar to that displayed for the Nd isotope data alone (Fig. 1). Several samples are characterized by very high  $\epsilon Nd(T)$  values (Elat dikes; Katzir et al. 2007b) and in some others mildly elevated  $\epsilon Nd(T)$  values not associated with low  $Sr_i$ . For details, see text

### Single zircon Hf isotope results

Twenty-one Lu–Hf isotope analyses of single zircons from CA2 and AL rocks are presented in Table 2 along with bulk  $\delta^{18}O(Zrn)$  for the rock sample hosting the zircon and ion-microprobe  $\delta^{18}O(Zrn)$  data for the same grains used for Lu–Hf isotope analysis. For most of the dataset (13 out of 21; Table 2), ion-microprobe analysis yielded core-rim  $\delta^{18}O(Zrn)$  values that are similar within error, showing that these are homogeneous zircons. For a few other zircons (4 out of 21; Table 2) only a core or a rim  $\delta^{18}O(Zrn)$  analysis of the specific Hf-analyzed grain exists, but in these cases the bulk population of other zircons from the sample shows good homogeneity, and thus they are less liable to core-rim differences (Fig. 2). Three Hf-analyzed zircons (Umm-I-Fai: magmatic zircon, Iqna pluton: high-U magmatic zircon; Zreir pluton: xenocryst) show core-rim  $\delta^{18}O(Zrn)$  differences of 0.9 to 3‰, which are significant (Table 2).

Interestingly, the high-U zircon of the Iqna pluton with the highest  $\delta^{18}O(Zrn)$  values and the highest core-rim variability within the dataset yielded very similar  $^{176}Hf/^{177}Hf$  as the low-U, homogeneous core-rim  $\delta^{18}O(Zrn)$  grain of the same pluton ( $\epsilon Hf(T)$  low-U =  $5.50 \pm 0.19$ ;  $\epsilon Hf(T)$  high-U =  $5.13 \pm 0.13$ ; Table 2). Thus although the high  $\delta^{18}O$  composition in the high-U grain is considered to reflect open system exchange with late fluids, this did not affect the magmatic Hf isotope composition. For two plutons (Timna porphyritic granite and Yehoshafat; Table 2) no  $\delta^{18}O(Zrn)$  data was acquired by ion-microprobe.

$\epsilon Hf(T)$  values of CA2 and AL rocks are plotted on a  $\epsilon Hf(T)$  versus age diagram in Fig. 6a. Overall, Figs. 3d and 6a show that the provinciality is well defined and two end members can be identified in the northwestern ( $-5.5$  to  $+5.9$ ) and southeastern ( $+9.2$  to  $+13.9$ ) parts of the region (Fig. 3d). Nevertheless, the transition between the two regions seems to be more complicated than that inferred from the Nd data. The middle-region with  $\epsilon Hf(T)$  values resembling those of the NW-region is geographically wider on its eastern side. Moreover, although most SE-region rocks are characterized by higher  $\epsilon Hf(T)$  values than those of the NW-region, there is considerable overlap between the NW-, SE-, and middle-regions over the  $\epsilon Hf(T)$  interval of 7.6–9.4. Similar to the  $\epsilon Nd(T)$  geographical division, the Timna porphyritic granite plots as an outlier, and the lower  $\epsilon Hf(T)$  range in the NW-region is exclusively defined by rocks of the AL suite.

An  $\epsilon Nd(T)$  versus  $\epsilon Hf(T)$  diagram is presented in Fig. 6b. Data points include single zircon  $\epsilon Hf(T)$  and WR  $\epsilon Nd(T)$  from this study and from the literature (Beyth et al. 1994; Katzir et al. 2007a; Moghazi et al. 1998; Stein and Goldstein 1996) for the same rocks. Also shown are WR  $\epsilon Nd(T)$ – $\epsilon Hf(T)$  fields for the ~630 Ma Elat schistose dikes from southern Israel (Katz et al. 2004) located in the NW-region and for rocks from southern Sudan (Vervoort and Blichert-Toft 1999) representing more southerly segments of the ANS. All single zircon  $\epsilon Hf(T)$  + WR  $\epsilon Nd(T)$  data points except one sample plot within the juvenile WR field of Vervoort and Blichert-Toft (1999). The NW-region rocks, excluding the Timna porphyritic granite, are located to the lower-left in Fig. 6b along with WR  $\epsilon Nd(T)$ – $\epsilon Hf(T)$  fields of the schist dikes, whereas the SE-region rocks plot at a more upper-right location closer to the field of SE Sudan rocks. While most pluton samples plot on the juvenile WR array, several samples plot below (two samples of the Timna porphyritic granite and Yehoshafat; Table 2) and one (Sharm pluton) plots well above the juvenile WR field displaying the most extensive Nd–Hf decoupling (Fig. 6b).

### Discussion

#### Isotope provinciality in the northernmost ANS

The available geochronological and isotopic data for the older tectono-magmatic phase (~820–740 Ma) of the northernmost ANS are scarce and do not properly cover different exposures of these rock sequences. It is therefore difficult to assess whether the ~820–740 Ma crust exposed in various locations (Fig. 1) was homogeneous in terms of its isotope composition. Nevertheless, the preliminary Nd–Sr data presented here for rocks of the Feiran-, Sa'al-, and Kid-MCs areas compare well with previously published (Stein and Goldstein 1996) and new data presented here for the Elat-MC.  $\epsilon Nd(T)$  of all rocks of this phase define a

Table 2 Single zircon Lu–Hf and  $\delta^{18}\text{O}(\text{Zrn})$  data for rocks of the study area in Sinai (Egypt) and southern Israel

Rock	Sample	Age (Ma)	$\delta^{18}\text{O}(\text{Zrn})$ MG/SG <sup>d</sup>	Grain No.	$\delta^{18}\text{O}(\text{Zrn})$		$^{176}\text{Lu}/^{177}\text{Hf}^a$	$^{176}\text{Lu}/^{177}\text{Hf}^b$	$^{176}\text{Lu}/^{177}\text{Hf}^c$	$\varepsilon\text{Hf}(\text{T})$	DM model ages		
					SG coré	SG rim <sup>f</sup>					$T_{\text{DM}}^1$	$T_{\text{DM}}^2$	$T_{\text{DM}}^3$
CA2 + AL NW-region													
CA2: Ahdar, granodiorite	S-2911	610	6.5/6.3	8a	6.4	6.1	0.001375	0.282643 (6)	8.35 (0.23)	0.94	1.02	1.10	
CA2: Eliat-Rehavam, monzogranite	YE-34	630	6.3/6.9	11a	6.3	6.9	0.002119	0.282643 (8)	8.45 (0.28)	0.95	1.03	1.13	
AL: Iqna, syenogranite	YS-503	578	7.1/8.2	4b	7.1, 7.7	7.4	0.000904	0.282577 (5)	5.50 (0.19)	1.06	1.18	1.42	
AL: Iqna, syenogranite	YS-503	578	7.1/8.2	4c	14.2	10.6	0.006132	0.282623 (4)	5.13 (0.13)	1.08	1.20	1.45	
CA2: Sulaf, granodiorite	S-2199	597	6.1/6.1	6b	6.4	6.4	0.001683	0.282640 (5)	7.85 (0.17)	0.96	1.04	1.23	
AL: Yehoshafat, Afs-granite	AG-81	605	6.6/N-A	3a	N-A	N-A	0.001439	0.282612 (15)	7.11 (0.55)	1.00	1.10	1.30	
CA2: TPG, monzogranite	AG-69	632	6.2/N-A	2a	N-A	N-A	0.000770	0.282667 (7)	9.91 (0.25)	0.88	0.94	1.07	
CA2 + AL SE-region													
CA2: Lathi, monzogranite	S-1610	607	5.1/4.8	2b	4.8	N-A	0.001114	0.282681 (10)	9.72 (0.36)	0.87	0.93	1.07	
CA2: Girgar, monzogranite	S-1561	603	5.3/5.4	2a	N-A	4.9	0.001489	0.282685 (10)	9.63 (0.35)	0.87	0.94	1.07	
AL: Sahara, Na-Afs-granite	S-1542	608	5.1/4.4	1a	4.1	4.3	0.003460	0.282666 (12)	8.28 (0.43)	0.94	1.02	1.19	
AL: Sharm, Afs-granite	S-1518	594	4.8/4.3	7c	4.5	4.4	0.000738	0.282802 (4)	13.88 (0.13)	0.65	0.66	0.68	
CA2: Sama, monzogranite	S-3704	608	5.1/5.1	8a	5.1	N-A	0.001689	0.282659 (15)	8.74 (0.52)	0.92	1.00	1.15	
CA2 + AL: median-region													
CA2: Hib-Mi, Qtz-diorite	S-2351	619	5.5/4.7	9a	5.5	5.3	0.002151	0.282630 (4)	7.77 (0.13)	0.98	1.06	1.25	
CA2: Zreir, monzogranite	S-4577	629	5.3/4.7	7c	5.4	5.6	0.000872	0.282592 (5)	7.15 (17)	1.02	1.11	1.31	
CA2: Zreir, monzogranite	S-4577	767	5.3/4.7	9a	1.2, 3.7	4.2	0.001005	0.282588 (11)	9.93 (40)	0.99	1.04	1.16	
CA2: Shihira Li/granodiorite	S-3807	632	5.4/5.6	8c	5.6	5.2	0.000934	0.282645 (11)	9.07 (0.39)	0.92	0.99	1.14	
CA2: Rahba/monzogranite	S-2069	610	5.7/5.4	1a	5.1	5.4	0.000931	0.282640 (17)	8.42 (0.59)	0.94	1.02	1.18	
AL: Umm-I-Fai, Na-Afs-granite	S-4403	586	5.5/5.2	6a	5.7	4.8	0.008747	0.282657 (5)	5.46 (0.17)	1.07	1.18	1.43	
AL: Dahab, syenogranite	S-3896	591	5.2/5.2	2c	4.8	5.1	0.000731	0.282626 (6)	7.59 (0.22)	0.97	1.05	1.24	
AL: Katharina, syenogranite	DV-82-87	583	5.8/5.7	1a	5.4, 5.7	N-A	0.001135	0.282606 (26)	6.55 (0.92)	1.01	1.11	1.33	
CA2: Abu-K'sheib, Qtz-monzonite	S-2695	592	5.2/5.5	10a	5.5	5.8	0.001190	0.282653 (9)	8.38 (0.33)	0.93	1.01	1.18	

Afs Alkali feldspar, MG multi-grain, N-A not analyzed, Na Na-amphibole bearing granite, Qtz quartz, SG single grain

a U–Pb ages: normal text, Rb–Sr ages: in italics, inferred ages based on other geological constraints underlined

b Multi-grain (MG)  $\delta^{18}\text{O}(\text{Zrn})$  values are of rock (not speci c sample) averages. SG (single grain) values are averages of all analyzed spots of this sample

c Low-U zircons  $U = 70\text{--}600$  ppm

d High-U zircons  $U = 2,500\text{--}7,000$  ppm

e Xenocrystic zircon of  $\sim 770$  Ma

f Single grain analyses are associated with  $\pm 0.5$  errors (a)

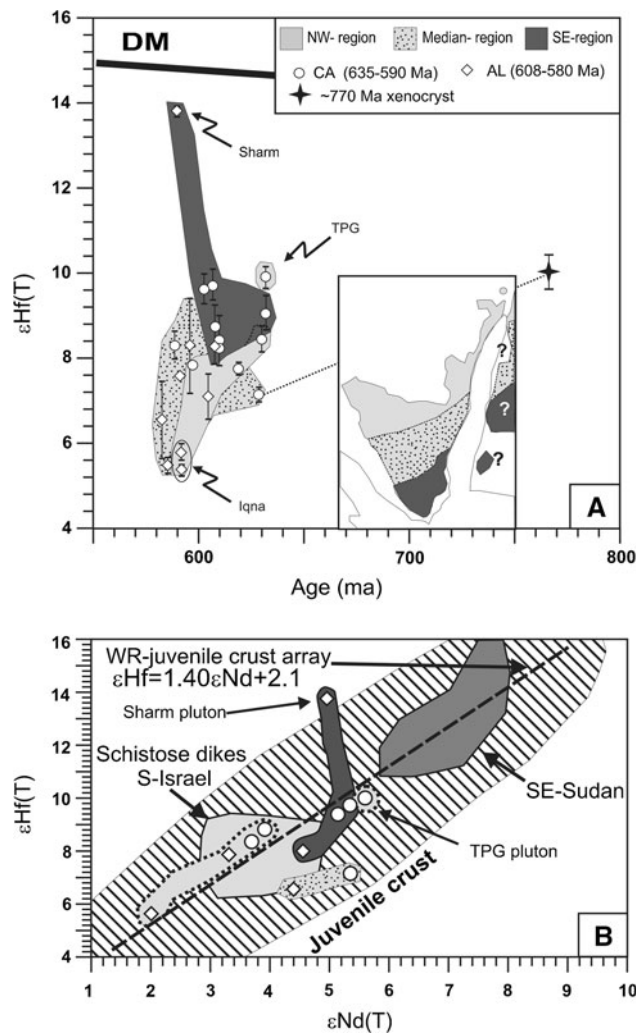
g Uncertainties  $^{176}\text{Lu}/^{177}\text{Hf}$  values are better than 2% (Blichert-Toft 2001)

h Errors [in parentheses (10<sup>-6</sup>)] are  $\pm 2\sigma$  of the mean

i  $T_{\text{DM}}^1 =$  Two stage model assuming a felsic crust precursor with  $^{176}\text{Lu}/^{177}\text{Hf} = 0.0093$  (Amelin et al. 1999)

j  $T_{\text{DM}}^2 =$  Two stage model assuming an average crust precursor with  $^{176}\text{Lu}/^{177}\text{Hf} = 0.015$  (GERM2001)

k  $T_{\text{DM}}^3 =$  Two stage model assuming a ma c precursor with  $^{176}\text{Lu}/^{177}\text{Hf} = 0.022$  (Vervoort and Patchett 1999)



**Fig. 6** a  $\epsilon\text{Hf}(T)$  versus age diagram showing the results of this study for single zircon grains (with  $2\sigma$  error bars). Similar to the Nd isotope results, CA2 samples within both the SE- and NW-regions display large variability in chemical composition (e.g.,  $\text{SiO}_2$  ~60–75 wt%), but only small variability in  $\epsilon\text{Hf}(T)$ ; b  $\epsilon\text{Nd}(T)$  versus  $\epsilon\text{Hf}(T)$  diagram. Data points include single zircon  $\epsilon\text{Hf}(T)$  from this study and WR  $\epsilon\text{Nd}(T)$  from this study and the literature (Beyth et al. 1994; Katzir et al. 2007a; Moghazi et al. 1998; Stein and Goldstein 1996) for the same rocks. Also shown are WR  $\epsilon\text{Nd}(T)$ – $\epsilon\text{Hf}(T)$  fields for the ~630 Ma Elat schistose dikes from southern Israel (Katz et al. 2004) located in the NW-region and for rocks from southern Sudan (Vervoort and Blichert-Toft 1999)

restricted field in Fig. 4 suggesting that the older complexes were mostly derived from rather similar sources.

The larger isotope dataset provided for the youngest tectono-magmatic phases (i.e., the CA2 and AL suites) in the northernmost ANS involved recycling of some older (~635–570 Ma) material, such as represented by the Sa'al schist (Be'eri-Shlevin et al. 2009c). This interpretation is in accordance with high- $\delta^{18}\text{O}(\text{Zrn})$  for some of the oldest rocks in this region interpreted by Be'eri-Shlevin et al. (2009b) to literature confirm the provinciality seen in O isotopes show that part of the 820–740 Ma magmas have assimilated variable amounts of supracrustal material.

isotope provinciality for CA2 and AL along a NW to SE transect in Sinai and southern Israel. A threefold geographical division is suggested for the CA2 and AL suites of the northernmost ANS whereby (a) rocks of the NW-region are characterized by positive but low  $\epsilon\text{Nd}(T)$ – $\epsilon\text{Hf}(T)$ , mildly-elevated to high  $\delta^{18}\text{O}$ , and mildly elevated  $\text{Sr}_f$ ; (b) rocks of the SE-region are characterized by higher  $\epsilon\text{Nd}(T)$ – $\epsilon\text{Hf}(T)$ , mantle-like  $\delta^{18}\text{O}$ , and low  $\text{Sr}_f$  and (c) a middle-region which is characterized by variable  $\epsilon\text{Nd}(T)$ – $\epsilon\text{Hf}(T)$ – $\text{Sr}_f$  values generally overlapping those of the NW-region, but with mantle-like  $\delta^{18}\text{O}$  values.

Provinciality of  $\epsilon\text{Nd}(T)$  values within the northernmost ANS was previously observed by Stern (2002), who defined it as a major Neoproterozoic border approximating the present location of the Gulf of Suez. The results of the present work point to a division along NE-SW trending lines (Figs. 4, 5, 6) that are oblique to the division suggested by Stern (2002).

#### Isotope constraints on 820–740 Ma crustal evolution

The formation of the ~820–740 Ma crust in the northernmost ANS has been traditionally interpreted in terms of calc-alkaline magmatism in a juvenile oceanic island arc setting, similar to formation of other parts of the shield at that time, although the lack of Neoproterozoic ophiolites and other features indicate there are significant differences. The available isotopic data for these rocks, as well as the occurrence of foreign older crust in the northernmost ANS (Be'eri-Shlevin et al. 2009c), now question this postulate. As shown in Fig. 6, the ~820–740 Ma rocks display a restricted  $\epsilon\text{Nd}(T)$  range of +3.4 to +4.9, with these values being slightly, but consistently more enriched compared to the DM at this time. The single ~770 Ma xenocryst detected in the Zreir pluton yield an  $\epsilon\text{Hf}(T)$  value of +9.9 (whereas the DM value would be +14 at this time), further implying that the rocks of this tectono-magmatic phase were not characterized by DM values. These results are in agreement with those presented by Morag et al. (2009) who suggested that older crustal material was recycled into the 820–740 Ma magmas.  $\text{Sr}$  values for most of these rocks are treated here with caution. Nevertheless, not a single  $\text{Sr}$  value from our dataset is compatible with MORB compositions at this time. Overall, the new isotope data presented here reaffirm the notion that the early evolution of tectono-magmatic phases in the northernmost ANS involved recycling of some older (~635–570 Ma) material, such as represented by the Sa'al schist (Be'eri-Shlevin et al. 2009c). This interpretation is in accordance with high- $\delta^{18}\text{O}(\text{Zrn})$  for some of the oldest rocks in this region interpreted by Be'eri-Shlevin et al. (2009b) to literature confirm the provinciality seen in O isotopes show that part of the 820–740 Ma magmas have assimilated variable amounts of supracrustal material.

Isotope constraints on 635–570 Ma crustal evolution

The late stages in the evolution of the ANS (postbased on their  $\epsilon_{\text{Nd}}$  (+4.9 to +10.2) and  $\text{Sr}$  (0.70124 to ~650 Ma) were characterized by vast intrusion of mostly 0.70336). This interpretation implies some variability in granitic batholiths and their associated volcanics (Bentothe Nd and Sr isotope compositions of the asthenospheric 1985). In the northernmost ANS this phase included the DM below the ANS and thus allows some deviations from 635–570 Ma CA2 and AL suites, and their formation has DM models. Magmas with Nd of +5 would be about 2 been associated with the change in tectonic activity during epsilon unit lower than the DePaold (981) at ~600 Ma. this time from collision to post-collisional activity (Be'eri-Shlevin et al. 2009a Beyth et al. 1994 Eyal et al. 2009 Stern 1994 Stern et al. 1984). A variety of sources have been suggested for the various ANS post-collisional magmas including the asthenospheric DM (Légeois and Stern 2009); the lithospheric mantle (Beyth et al. 1994 Eyal et al. 2009 Kessel et al. 1998 Jarrar et al. 2003 Moghazi et al. 1998 Moussa et al. 2008 Mushkin et al. 2003 Stein 2003) and the ANS juvenile crust (Be'eri-Shlevin et al. 2009b Stein 2003). In detail, less attention has been paid to the sources of CA2 magmas, but several authors modeled them as partial melts of a mafic source (Jarrar et al. 2003 Eyal et al. 2009). Notably, while Eyal et al. 2009 assigned this mafic source to lithospheric mantle-derived basaltic magmas, Stein 2003 suggested CA2 magmas could be derived from melting of the juvenile ANS crust. The Goldstein (1996) works and their derivation from lithospheric mantle sources has been generally accepted (Jarrar et al. 2003 Kessel et al. 1998 Moghazi et al. 1998 Mushkin et al. 2003 Stein 2003). Nevertheless, the recognition that some A-type granites are characterized by elevated  $\epsilon_{\text{Nd}}$  was interpreted by Be'eri-Shlevin et al. (2009b) to indicate the involvement of supracrustal material in their petrogenesis.

As outlined earlier, one of the main results of this study is that the ~635–570 Ma CA2 and AL rocks of the northernmost ANS display prominent variability in terms of their isotope compositions. This can reflect geographically variable sources or processes. Apart from the isotopic provinciality we point to a few other important observations which could be used to better constrain the sources of CA2 and AL magmas.

#### Deviations from Nd and Hf depleted mantle model values

Rocks of both suites are characterized by initial Nd, and Hf isotope ratios ( $\epsilon_{\text{Nd}}$  of +1.1 to 5.6 and  $\epsilon_{\text{Hf}}$  of +5.5 to +9.9, and +13.9) that are less depleted than the predicted a DM values at ~600 Ma [ $\epsilon_{\text{Hf}}$  of about +14,  $\epsilon_{\text{Nd}}$  of about +6.5, or +9.5 using DM evolutionary lines of DePaold (981) and Goldstein et al. (1984) respectively] (Figs 4, 6). The positive  $\epsilon_{\text{Hf}}$  and  $\epsilon_{\text{Nd}}$  values of CA2 and AL rocks indicate that there was no major contribution from old (pre-ANS) crust in their genesis. Nevertheless, we consider the consistent deviation from DM models a significant feature.

Liégeois and Stern (2009) advocated an asthenospheric DM source for the Meatiq–Hafa t complexes in Egypt,  $\epsilon_{\text{Nd}}$  (+4.9 to +10.2) and  $\text{Sr}$  (0.70124 to ~650 Ma) were characterized by vast intrusion of mostly 0.70336). This interpretation implies some variability in granitic batholiths and their associated volcanics (Bentothe Nd and Sr isotope compositions of the asthenospheric 1985). In the northernmost ANS this phase included the DM below the ANS and thus allows some deviations from 635–570 Ma CA2 and AL suites, and their formation has DM models. Magmas with Nd of +5 would be about 2 been associated with the change in tectonic activity during epsilon unit lower than the DePaold (981) at ~600 Ma. this time from collision to post-collisional activity (Be'eri-Shlevin et al. 2009a Beyth et al. 1994 Eyal et al. 2009 Stern 1994 Stern et al. 1984). A variety of sources have been suggested for the various ANS post-collisional magmas including the asthenospheric DM (Légeois and Stern 2009); the lithospheric mantle (Beyth et al. 1994 Eyal et al. 2009 Kessel et al. 1998 Jarrar et al. 2003 Moghazi et al. 1998 Moussa et al. 2008 Mushkin et al. 2003 Stein 2003) and the ANS juvenile crust (Be'eri-Shlevin et al. 2009b Stein 2003). In detail, less attention has been paid to the sources of CA2 magmas, but several authors modeled them as partial melts of a mafic source (Jarrar et al. 2003 Eyal et al. 2009). Notably, while Eyal et al. 2009 assigned this mafic source to lithospheric mantle-derived basaltic magmas, Stein 2003 suggested CA2 magmas could be derived from melting of the juvenile ANS crust. The Goldstein (1996) works and their derivation from lithospheric mantle sources has been generally accepted (Jarrar et al. 2003 Kessel et al. 1998 Moghazi et al. 1998 Mushkin et al. 2003 Stein 2003). Nevertheless, the recognition that some A-type post-collisional magmas.

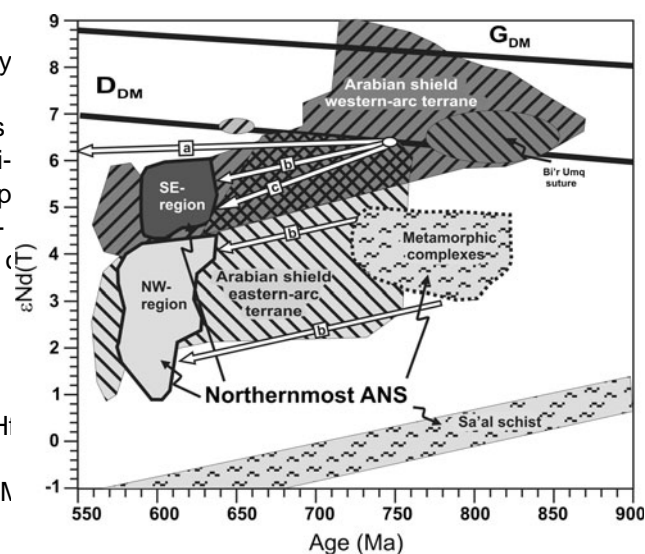


Fig. 7 Comparison between the Nd isotopic compositions of rocks from the northernmost ANS and from its more southerly segments in Saudi Arabia (data from Hargrove et al. 2006; Stoesser and Frost 2006 and references therein).  $D_{\text{DM}}$  and  $G_{\text{DM}}$  are as in Fig. 4. White arrows (a–d) correspond to  $^{147}\text{Sm}/^{144}\text{Nd}$  evolutionary lines of 0.19, 0.13 and 0.1, respectively

The stabilization of the ANS lithospheric mantle root magmas were sourced from the lithospheric mantle, has accompanied the formation of the ANS crust from whereas silicic magmas reflect crustal anatexis. 870 Ma and onwards (Stein and Goldstein 1996; Stein 2003). This implies that this reservoir has been detached components could have been acquired via subduction-related processes prior to 640 Ma. However, the large volume of this mantle reservoir makes this possibility of ANS mantle xenoliths suggested that this reservoir has been highly speculative, and the fact that the direct samplers of been affected by several episodes of melt extraction during this reservoir, e.g., the mafic rocks of the NW-region do not record any evidence of high  $\delta^{18}\text{O}$  components certainly does not support such scenarios. Several features associated with the new Nd–Sr–Hf isotope data presented in this study, as well as with the  $\delta^{18}\text{O}$  values of zircon cores and rims, and zircon inheritance suggest that AFC processes did not play an important role in the petrogenesis of silicic CA2 and AL rocks not only in the NW-region but also in the middle and SE-regions. CA2 and AL rocks are geographically zoned in terms of their radiogenic (Nd–Sr–Hf) isotope compositions. Thus, within the SE- and NW-regions these rocks are characterized by distinct isotope patterns. Notably, the elevated  $\delta^{18}\text{O}$  of silicic CA2 and AL rocks in the NW-region are associated with lower Nd(T) and Hf(T) and lower Sr. Nevertheless, isotope patterns within all regions are unrelated to variations in the major element chemical composition ( $\text{SiO}_2$  content; Fig. A2 available online in Electronic supplementary material) and thus to the degree of differentiation. If AFC processes were the main mechanism associated with formation of 'radiogenic isotope-enriched' CA2 magmas, co-variation of the isotope composition with extent of assimilation (i.e.,  $\text{SiO}_2$  content and other major element contents) would be expected. Such features do not fit well with AFC or magma-mixing processes as a mechanism for enrichment of the radiogenic isotope compositions. The results of ion-microprobe  $\delta^{18}\text{O}(\text{Zrn})$  analysis have confirmed the interpretation that the multi-grain  $\delta^{18}\text{O}$  results presented by Be'eri-Shlevin et al. (2009b) represent magmatic values. These results provide additional support for the notion that sources of CA2 and AL rocks were geographically zoned at  $\sim 640$  Ma. In most of the analyzed zircons, cores and rims display similar ranges of  $\delta^{18}\text{O}(\text{Zrn})$  values (1:1 ratios in Fig. A1; available online in Electronic supplementary material). The high  $\delta^{18}\text{O}$  values measured in both zircon cores and rims in silicic rocks of the NW-region indicate that crystallization of zircon commenced within already high- $\delta^{18}\text{O}$  magma chambers with no record of later contribution of high- $\delta^{18}\text{O}$  wallrock material to the magma or with addition of high  $\delta^{18}\text{O}$  material via magma mixing. This pattern of zircon growth is consistent among plutons of variable  $\text{SiO}_2$  content and affiliation with the CA2 or the AL magmas in the NW-region. In contrast, rocks of the SE- and middle-regions were derived from sources characterized by mantle-like  $\delta^{18}\text{O}$  (Be'eri-Shlevin et al. 2009b).

#### *High $\delta^{18}\text{O}$ magmas: indication for involvement of supracrustal material*

The detection of elevated  $\delta^{18}\text{O}$  of some CA2 and AL rocks (e.g., in the NW-region; Be'eri-Shlevin et al. 2009b) requires that the participation of supracrustal material in their petrogenesis. In contrast, all mafic rocks from this region display mantle-like  $\delta^{18}\text{O}$  values (Be'eri-Shlevin et al. 2009b). Similarity in radiogenic isotopic composition between mafic and silicic AL magmas as well as the close spatial and temporal association of the two rock types have been used to argue that AL A-type granites are highly fractionated (90%) mafic melts derived from the lithospheric mantle (e.g., Mushkin et al. 2003). Similar lines of evidence could be drawn for an association between mafic and silicic CA2 magmas. Nevertheless, the contrast between the mantle-like  $\delta^{18}\text{O}$  of all mafic rocks and the elevated  $\delta^{18}\text{O}$  of all silicic CA2 and AL rocks in the NW-region (excluding the Timna complex of  $\sim 10$  km<sup>2</sup>; Fig. 1) suggests that this direct association is problematic.

#### *Mafic and silicic magmas: variable sources or variable processes?*

Several possible explanations are evaluated below to explain the high  $\delta^{18}\text{O}$  of the NW-region rocks: (a) mafic and silicic magmas were derived from the lithospheric mantle variably enriched in high  $\delta^{18}\text{O}$  components; (b) mafic and silicic magmas were sourced from DM sources; however, assimilation of crustal material by some mafic AL suites suggesting deep high  $\delta^{18}\text{O}$  sources for CA2 and parent magmas and further fractional crystallization (AFC) of variable  $\text{SiO}_2$  content and affiliation with the CA2 or the AL magmas in the NW-region. In contrast, rocks of the SE- and middle-regions were derived from sources characterized by mantle-like  $\delta^{18}\text{O}$  (Be'eri-Shlevin et al. 2009b); (c) mafic

In addition to the isotope data discussed earlier, it is important to mention that the vast majority of CA2 and AL rocks of the study area carry no xenocrystic zircons, and they are minor components within the zircon population. The main problem associated with derivation of the rare cases where such xenocrysts have been detected is the difference in geochemical affinity of these rocks. A molten lower-crust producing CA2 magmas would be expected to be depleted, not enriched as depicted thus clearly provides no support for any AFC or magma-mixing processes associated with the petrogenesis of CA2 and AL magmas.

Derivation of CA2+AL mafic and silicic magmas from different sources, e.g., the lithospheric mantle and the juvenile ANS crust may provide a good explanation for their different  $\delta^{18}\text{O}$  values in the NW-region. Evidence for high  $\delta^{18}\text{O}$  820–740 Ma magmas in the NW-region has been previously reported (Be'eri-Shlevin et al. 2009a) and underthrusting of such high  $\delta^{18}\text{O}$  components to magma-forming depths could have occurred during collisional events that characterized the ANS up to 650–630 Ma and were accompanied by thickening of the crust and lithosphere. Some authors (Abdel-Fattah and El-Kibbi 2001; El-Bialy and Streck 2009) suggested that A-type granites in the spheric mantle prior to the onset of post-collisional magmatism (Avigad and Gvirtzman 2009; Stern 1994). Considering anatectesis of ANS juvenile crust as the source of all silicic CA2 and AL magmas, we point to the similarity between the different regions defined in the northernmost ANS and those of more southerly segments. Figure 7 shows that  $\text{Nd}$  fields of CA2 and AL rocks of the SE- and NW-regions overlap those of the Arabian western- and eastern-arc terranes, respectively. A regression line through the field of Arabian western-arc terrane and the SE-region of this work (approximate crustal evolution trend of Hargrove et al. 2006) is defined by  $^{147}\text{Sm}/^{144}\text{Nd} = 0.13$ , (evolution line b in Fig. 7), which is too high for the felsic crust. A similar regression line can be drawn for older and younger rocks of the NW-region and the eastern arc terrane of Arabia (Fig. 7).  $^{147}\text{Sm}/^{144}\text{Nd}$  ratios of 0.131 to 0.297 (average  $1.9 \pm 0.05$ ;  $t_r$ ) have been reported by McGuire and Stern (1993) for granulite xenoliths interpreted to represent the mafic lower-crust in Saudi Arabia. Figure 7 shows crustal evolutionary lines with  $^{147}\text{Sm}/^{144}\text{Nd}$  ratios of 1.9 (line a) and 1.1 (line c) representing the average granulite composition of McGuire and Stern (1993) and a theoretical felsic crust, respectively, for reference. It is evident that if the mafic lower-crust is characterized by  $^{147}\text{Sm}/^{144}\text{Nd} = 1.9$  as depicted by the data from granulite xenoliths, then the regression line of 0.13 could represent a mixture between mafic and intermediate magmas at least in the NW-region. These sources could correspond to lower and intermediate levels of the ANS crust.

We conclude that derivation of CA2 and AL silicic magmas from intermediate to lower levels of the juvenile ANS crust versus the derivation of mafic magmas from the lithospheric mantle sources complies well with the isotopic data and that this may explain the difference between the O important to mention that the vast majority of CA2 and AL isotopic signature of mafic and silicic rocks of the NW-rocks of the study area carry no xenocrystic zircons, and in region. The main problem associated with derivation of the rare cases where such xenocrysts have been detected both CA2 and AL rocks from the ANS lower or intermediate crust is the difference in geochemical affinity of these rocks. A molten lower-crust producing CA2 magmas would be expected to be depleted, not enriched as depicted thus clearly provides no support for any AFC or magma-mixing processes associated with the petrogenesis of CA2 and AL magmas. Derivation of CA2+AL mafic and silicic magmas from different sources, e.g., the lithospheric mantle and the juvenile ANS crust may provide a good explanation for the composition of the intrusions to dominantly granodiorites and monzogranites. More important is the fact that this stage overlapped the main phase of AL silicic (and mafic) magma formation (Be'eri-Shlevin et al. 2009a). Thus, possible derivation from different levels of geochemically heterogeneous crust is possible. We note that some authors (Abdel-Fattah and El-Kibbi 2001; El-Bialy and Streck 2009) suggested that A-type granites in the ANS could be modeled as melts derived from the lower crust after fluid metasomatism. The sources of such fluids although still remain ambiguous. Considering the possible sources for CA2 and AL magmas discussed earlier and the fact that AFC and magma-mixing processes did not play an important role in their formation, we suggested different sources for mafic and silicic magmas of both CA2 and AL suites. The most plausible source of CA2 and AL mafic magmas is the lithospheric mantle, which in the NW- and middle-regions of this reservoir displays variable enrichment in radiogenic isotope compositions. This notion can be related to the early history of the NW-region discussed earlier, suggesting that this region was initially (i.e., at  $\sim 800$  Ma) not evolving completely in ensimatic environments. As for the sources of silicic rocks which volumetrically dominate the CA2 and AL suites, the problem is more complicated. Potentially, both lithospheric mantle and juvenile ANS could be considered as probable sources. Nevertheless, several lines of evidence presented earlier suggest that melting of various levels of the crust (i.e., lower and intermediate levels) provides a more coherent model for the generation of these magmas at least in the NW-region. Apart from the NW-SE isotope variability, CA2 and AL rocks of the various regions in the study area display no other significant differences in age or elemental geochemical affinities (Be'eri-Shlevin et al. 2009a, b; Eyal et al. 2009). Since the evolution of these suites in the lithospheric mantle sources complies well with the isotopic NW- and SE-regions is similar, it is highly improbable that

the former were sourced from crustal melts whereas they are also located at the margins of the ANS it is not latter from differentiation of lithospheric ma c melts. We surprising that they plot in partial overlap with the eastern-emphasize here that mantle-like  $\delta^{18}\text{O}$  compositions do not arc terrane of Arabia. necessarily re ect mantle sources; it is only when higher The isotopic division of CA2+ AL rocks in the (or lower) than mantle range  $\delta^{18}\text{O}$  magmas are detected northernmost ANS resembles the larger-scale geographical that the involvement of supracrustal material is evident. If division outlined earlier: the SE-region is located closer to the lower or intermediate levels of the crust never inter-the ANS core, whereas the NW-region is located closer to acted with the supracrustal material they would retainits margin. These relations imply that after 640 Ma CA2 mantle-like  $\delta^{18}\text{O}$  compositions although their radiogenic and AL magmas of the SE- and NW-regions sampled isotope evolution could be different than the lithospheric sources that were related to the ANS' juvenile core and mantle. Thus the mantle-like  $\delta^{18}\text{O}$  compositions consis- contaminated margins, respectively. tently recorded in SE-region CA2 and AL rocks may Although showing similar Nd isotope provinciality, simply re ect the fact that such lower or intermediate crustthere are marked differences between the northernmost sources have never interacted with high  $\delta^{18}\text{O}$  supracrustal ANS and the Arabian segment of the shield. In Arabia, the material. major terranes are bordered by fault zones that are

To conclude this part, we suggest that silicic rocks of accompanied by ophiolitic complexes, and the terranes are both CA2 and AL rocks including A-type granites repre- therefore allochthonous relative to one another (Stoeser sent melts derived from the intermediate to lower crust. A and Frost 2006). In the northernmost ANS, no ophiolitic  $\sim 640$  Ma this reservoir was isotopically zoned horizon-sutures are identi ed (but see Shimro 1980), and the tally along a NW-SE transect and probably also variably border between the northernmost ANS and the more zoned vertically in the NW-region. Ma c rocks of southerly ANS segments likewise is not characterized by  $\sim 600$  Ma associated with the AL suite of the northernmost an ophiolitic suture but rather by a change in lithology— ANS were derived from a lithospheric mantle of Nd–Sr dominance of older complexes to the south and younger enriched composition re ecting the early, not completely ( $\sim 600$  Ma) granitoids to the north (e.g., Stern and Hedge 1985), and the lack of the Najd fault system in the north (Stern 1985). The results of this study suggest that the change in the Nd–Sr–Hf isotope ratios across the NW-SE transect is not abrupt. Similarly, Be'eri-Shlevin et al. (2009b) have pointed to the gradual rather than stepwise

The northernmost ANS versus its more southerly segments

Comparison between the Nd (and Hf) isotopic composi-change in the  $\delta^{18}\text{O}(\text{Zrn})$  across this transect. Thus, tions of rocks from the northernmost ANS and from its although two different reservoirs, the ANS juvenile core more southerly segments in Saudi Arabia (Hargrove et al versus contaminated margins, were sampled by the SE- and 2006; Stoeser and Fros 2006 and references therein; NW-region CA2 and AL magmas, there is no record of a Fig. 7) was used to provide new means to evaluate the sharp border between allochthonous terranes as depicted sources of the northernmost ANS magmas. It is evident that for the Arabian segment of the shield. Taken together, the NW-SE Nd isotope provinciality depicted for the CA2 these features imply a different evolution of the northern- and AL rocks of the northernmost ANS resembles themost ANS crust as compared to more southerly ANS major division portrayed for the Arabian segment of thesegments, at least after 640 Ma.

shield (Fig. 7). Fields of the SE- and NW-regions overlap Stern (1985) suggested on the basis of structural evi- the elds of the western- and eastern-arc terranes (e.g. dence that rifting in the northernmost ANS was associated Stoeser and Fros 2006), respectively. with development of the Najd fault system to the south.

The similarity between the NE-SW division in the This model was further developed by Be'eri-Shlevin et al. northernmost ANS and the generally E-W division in (2009b) who suggested that rifting commenced from at Arabia is not random and stems from the position of the east  $\sim 630$  Ma, in accordance with previous estimates for crustal segments in relation to the ANS core and margins movement along the Najd fault system (Stacey and Agar (Be'eri-Shlevin et al. 2009c). The western-arc terrane of 1985). The rifting model explains the lack of the Najd Arabia is of oceanic arc af nity and forms part of the fault system in this region and provides a mechanism with juvenile ANS (geographical) core. In contrast, the eastern which to generate such vast amounts of melting during arc terrane of more continental-like isotope af nity is this time (Be'eri-Shlevin et al. 2009b), as well as the closer to its exposed NE margin (Stoeser and Fros 2006). space to intrude so many granitic plutons. The passive As discussed earlier, the older  $\sim 820$ – $740$  Ma) rocks of emplacement of massive volumes of granitic intrusions the northernmost ANS are characterized by more contiduring this time could have erased evidence for suturing nental-like Nd (and Hf) isotope compositions and since between older allochthonous terranes, thus providing an



explanation for the lack of ophiolite sutures. We emphasize that such a model would explain large-scale magma genesis during this time only in the northernmost ANS. Large stretches of the Najd fault system (exposed in more southerly segments of the shield) exhibits high-grade metamorphism and brittle-ductile mylonitic deformation (Johnson and Woldehaiman 2003) which are not easily compatible with extension and rifting. Nevertheless, conjunction of compressional tectonics in these more southerly ANS segments during 650–600 Ma and extension within the northernmost ANS has been recently advocated by Avigad and Gvirtzman (2009) in association with their model for lithospheric delamination during this time.

### Hf and Nd model ages

Depleted mantle model ( $T_{DM}$ ) ages assume that the magma or its precursor separated from a DM reservoir. Importantly,  $T_{DM}$  ages critically depend on various assumptions and on values assigned to several parameters that are used

in the model. Over the past decades the model parameters of DePaolo (1981) have been preferred over those of Goldstein et al. (1984) in calculating Nd- $T_{DM}$  ages for ANS rocks (Hargrove et al. 2006; Moussa et al. 2008; Stern, 2002).

The two types of Nd- $T_{DM}$  ages calculated for the samples of this work (using a conservative  $^{147}\text{Sm}/^{144}\text{Nd} < 0.165$ ; Stern 2002) are presented in Table 1 and Fig. 8. The average Nd- $T_{DM1}$  (i.e., DePaolo 1981) for the filtered dataset is 0.86 Ga, which is in good agreement with the reported 0.84–0.1 Ga Nd- $T_{DM}$  ages of Stern (2002) for this region. However, using the NW-SE geographical division, CA2 and AL yield Nd- $T_{DM1}$  averages of 0.72 and 0.90 Ga for the SE- and NW-regions, respectively. The older (800–780 Ma) rocks yield an even higher Nd- $T_{DM1}$  average of 1.03 Ga. As expected, the average Nd- $T_{DM2}$  (i.e., Goldstein et al. 1984) for the entire filtered dataset and for each geographical group is 0.15–0.20 Ga older than the respective Nd- $T_{DM1}$  (Table 1).

Nd- $T_{DM}$  ages are sometimes suspected to misrepresent separation ages of a magmatic suite from the DM, and are thought instead to rather yield average crustal residence times of the ANS' juvenile core, we consider the Nd- $T_{DM1} = 0.72$  Ga and Nd- $T_{DM2} = 0.89$  Ga ages as the best estimates given by the current WR Nd dataset for >700 Ma ANS core rocks (e.g., DePaolo (1981) and Goldstein et al. (1984) models

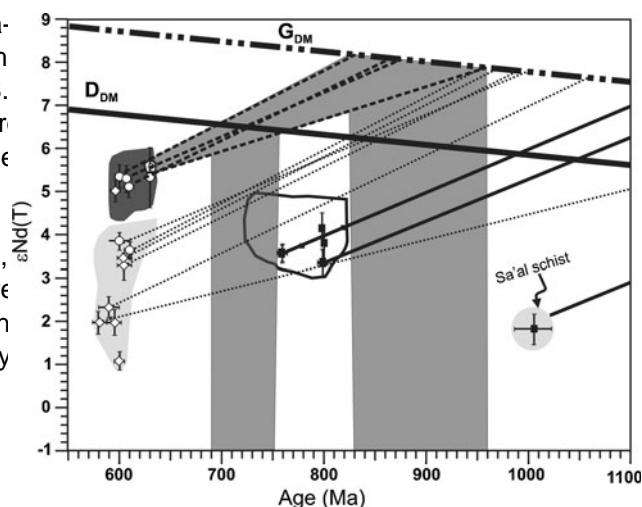


Fig. 8  $\epsilon\text{Nd}(T)$  versus age diagram showing Nd depleted mantle model ( $T_{DM}$ ) ages calculated for the samples of this work with  $^{147}\text{Sm}/^{144}\text{Nd} < 0.165$ . Error bars are  $\pm 2\sigma$ .  $G_{DM}$  and  $D_{DM}$  are as in Fig. 4

(Fig. 7), it seems that both models could be used to predict the evolution of the DM in the ANS region.

Three types of Hf- $T_{DM}$  ages calculated using different  $^{176}\text{Lu}/^{177}\text{Hf}$  ratios for the precursor of the magmas (felsic, intermediate, and basic) are presented in Table 2 and Fig. 9. Hf- $T_{DM1}$  ages provide the minimum estimate for crustal formation events (e.g., Kemp et al. 2006) in the Hf isotope system. Hf- $T_{DM1}$  ages calculated for CA2 and AL rocks of the SE-region (excluding the Sharm pluton) yield an average of 0.89 Ga and show marked correlation with Nd- $T_{DM2}$ , but not with Nd- $T_{DM1}$  (Fig. 9). We have estimated above that the CA2 and AL magmas were derived from lower to intermediate levels of the crust; thus, more realistic model ages for the Hf system should be calculated using a mafic or intermediate precursor (e.g., Hf- $T_{DM2}$  and Hf- $T_{DM3}$  respectively). Excluding the Sharm pluton, Hf- $T_{DM2}$  model ages calculated with crustal precursor of intermediate composition span a range of 0.93–1.02 GA in agreement with accepted formation age of the ANS (e.g., Stern 1994). In contrast, Hf- $T_{DM3}$  calculated with a mafic

precursor yield ages of 1.07–1.19 Ga, some 100–200 Ga older than the major crystallization events recorded in the ANS. Nevertheless, in accordance with Hawkesworth and Kemp (2006) and Kemp et al. (2006), such discrepancies may simply reflect an older crustal formation event at 1.1–1.2 Ga, when zircon-free mafic material was extracted from the mantle, whereas younger events (>700 Ma) represent reworking of this lower crust (e.g., crustal differentiation of Kemp et al. 2006). Future analysis of Hf isotopes in zircons from rocks of more southerly regions within the ANS may help to resolve this problem.

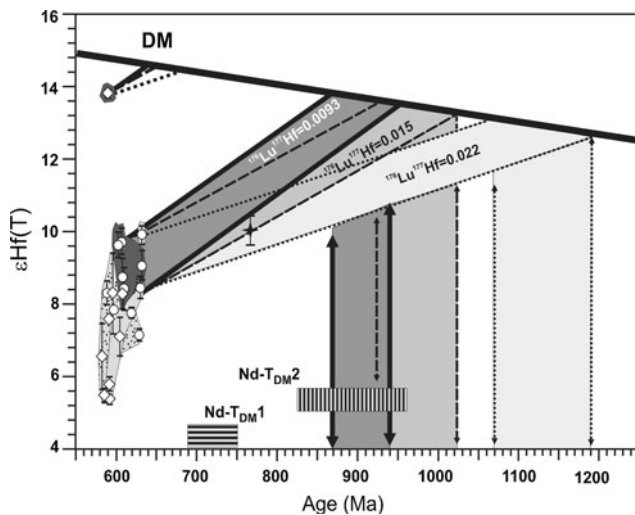


Fig. 9  $\epsilon\text{Hf}(T)$  versus age diagram showing the three types of  $\text{Hf}/T$  ages calculated for SE-region samples of this study.  $\text{Hf}/T$  differs by the  $^{176}\text{Lu}/^{177}\text{Hf}$  ratio assumed for the magma precursor (felsic, intermediate, or basic).  $\text{Nd}-T_{\text{DM}}$  from Fig. 9 are shown for comparison. Error bars are  $\pm 2$

#### Summary and implications for modeling the evolution of the ANS

The results of this study support the idea that on a large scale, the ANS is isotopically zoned and that evidence of smaller-scale isotope provinciality is difficult to reconcile with the mantle-plume model of Stein and Goldstein (1996). Moreover, the combined large- and smaller-scale isotope provinciality is difficult to reconcile with the mantle-plume model of Stein and Goldstein (1996). The results of this work also contribute to the debate regarding the petrogenesis of A-type magmas. We suggest here that silicic rocks of both CA2 and AL rocks represent juvenile ANS crustal melts. While most CA2 magmas are more easily modeled as derived from melting of intermediate to lower crustal levels (e.g., Stein and Goldstein 2003), the origin of silicic AL magmas is more debatable and derivation from the mantle has been suggested by several authors. The fact that radiogenic isotope ratios do not co-vary with either the major (and trace) element contents and the  $\delta^{18}\text{O}$  zoning in zircon is interpreted here to show that silicic AL magmas were derived from crustal reservoirs as their CA2 counterparts. In accordance with Katzir et al. (2007) and Be'eri-Shlevin et al. (2009b), the marked difference in  $\delta^{18}\text{O}$  values between AL silicic (A-type granite) and AL mafic rocks within the NW-region is taken as evidence that the temporal and spatial association between them does not necessarily imply co-genetic relationships as favored by some authors (Jarrar et al. 2008; Mushkin et al. 2003). Different Nd and Hf DM model ( $T_{\text{DM}}$ ) ages were compared here only for the CA2 and AL rocks of the SE-region as these represent the juvenile ANS core. Such comparison suggests that modeling the DM for the ANS

at least two major reservoirs: the juvenile ANS core and its more contaminated northern margin. Formation of these reservoirs commenced during the early stages of ANS formation, but crustal contamination in its northern (and

eastern?) parts could have continued and increased by the addition of cratonic derived sediments (see Wilde and Youssef 2002). The pattern of isotopic provinciality in the northernmost ANS, showing a gradual rather than step-wise change in isotope ratios of CA2 and AL rocks across the NW-SE transect, is markedly different from that displayed in Saudi Arabia. Importantly, the detection of two or more allochthonous terranes, as shown for the Arabian segment of the shield, is not possible here. We have further advocated the model of lithospheric rifting during this time as first suggested by Stern et al. (1984) and Stern (1985) and later promoted by Be'eri-Shlevin et al. (2009b) to explain this difference. We propose that if two allochthonous terranes were amalgamated prior to 640 Ma, rifting and passive emplacement of vast amounts of granite after ~635 Ma would have erased any evidence of a suture zone.

Overall, the formation of two reservoirs in the ANS (i.e., juvenile core and more contaminated margin) at an early stage and their evolution up to the time they were sampled by the 635–570 Ma CA2 and AL magmas provide a simpler explanation of the observed pattern and range of ANS Nd–Sr–Hf–O isotope compositions than models inferring sources such as enriched mantle reservoirs (similar to those sampled by modern OIB; Katzir et al. 2007b; Stein and Goldstein 1996). Moreover, the combined large- and smaller-scale isotope provinciality is difficult to reconcile with the mantle-plume model of Stein and Goldstein (1996). The results of this work also contribute to the debate regarding the petrogenesis of A-type magmas. We suggest here that silicic rocks of both CA2 and AL rocks represent juvenile ANS crustal melts. While most CA2 magmas are more easily modeled as derived from melting of intermediate to lower crustal levels (e.g., Stein and Goldstein 2003), the origin of silicic AL magmas is more debatable and derivation from the mantle has been suggested by several authors. The fact that radiogenic isotope ratios do not co-vary with either the major (and trace) element contents and the  $\delta^{18}\text{O}$  zoning in zircon is interpreted here to show that silicic AL magmas were derived from crustal reservoirs as their CA2 counterparts. In accordance with Katzir et al. (2007) and Be'eri-Shlevin et al. (2009b), the marked difference in  $\delta^{18}\text{O}$  values between AL silicic (A-type granite) and AL mafic rocks within the NW-region is taken as

evidence that the temporal and spatial association between them does not necessarily imply co-genetic relationships as favored by some authors (Jarrar et al. 2008; Mushkin et al. 2003). Different Nd and Hf DM model ( $T_{\text{DM}}$ ) ages were compared here only for the CA2 and AL rocks of the SE-region as these represent the juvenile ANS core. Such comparison suggests that modeling the DM for the ANS

region with the parameters of Goldstein et al (1984) or DePaolo (1981) is appropriate.  $Hf-T_{DM3}$  calculated with a mafic crustal precursor yield ages of 1.07–1.19 Ga which is some 100–200 Ga older than the major crystallization events recorded in the ANS. Such discrepancies may reflect an older crustal formation event at 1.1–1.2 Ga, when zircon-free mafic material was extracted from the mantle, whereas younger events (~670–550 Ma) represent reworking of this lower crust.

**Acknowledgments** This work was supported by DAAD grant A/05/06051 and the Synthesys grant SE-TAF 2722. We wish to thank J. Hoefs for his help in arranging Sm–Nd, Rb–Sr analyses in Uppsala. J. Blichert-Toft acknowledges financial support from the French Institut National des Sciences de l'Univers. The first author wishes to thank B. Litvinovsky and the late A. Zarnivlevich for stimulating discussions and advice. Discussions with S. Claesson also greatly contributed to this manuscript. Comments given by two anonymous reviewers critically improved this manuscript. The NordSIMS facility, (Swedish Museum of Natural History) is supported by the research councils of Denmark, Norway, and Sweden, the Geological Survey of Finland and the Swedish Museum of Natural History. This is NordSIMS contribution #248.

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