Onset of Maikop sedimentation and cessation of Eocene arc volcanism in the Talysh Mountains, Azerbaijan

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Abstract: The Maikop Series forms an important source rock in the former Paratethys. Deposition is often interpreted as anoxic, linked to restriction of the Paratethys. The Pirembel formation in the Talysh Mountains (Azerbaijan) is attributed to the Maikop Series and was deposited above the Eocene volcanic Peshtasar formation. Dating the onset of anoxia could help to distinguish glacio-eustatic from tectonic causes of restriction. We integrated magnetostratigraphy and biostratigraphy to date the onset of Pirembel sedimentation and used geochemistry to characterize the tectonic setting of the Peshtasar volcanic rocks. The onset of Maikop sedimentation in the Talysh was determined to be 37.7 Ma, ruling out a link with the major sea-level drop at the Eocene–Oligocene Transition (33.9 Ma) and favouring a tectonic cause. Extrapolating the average sedimentation rate (34 cm kyr) suggests that the entire Pirembel formation belongs to the Late Eocene. We hypothesize that the end of volcanism is important in the transition to Pirembel sedimentation. The palaeomagnetic and geochemistry results for the volcanic rocks cluster in three groups, suggesting three distinct episodes of volcanism. Volcanic sills within the Eocene Arkevan formation plot exactly on these groups, confirming the relationship between the Arkevan and Peshtasar formations. Volcanic rocks of the Talysh show continental-arc signatures and may be related to an Eocene volcanic belt extending towards southeastern Iran.

Supplementary material: The full analytical data of the Ar–Ar dating are available at http://www.geolsoc.org.uk/SUP18851

The Maikop Series was deposited in the Paratethys, an epicontinental sea extending from Germany to China during the Palaeogene (Fig. 1a). The Maikop Series predominantly consists of black shales that are the most important source rock for hydrocarbons in the South Caspian Basin and are thus of great economic importance. Black shales generally contain a relatively large amount of unoxidized carbon and are usually deposited under anoxic, reducing conditions (Passier et al. 1999; Efendiyeva 2004; Hudson et al. 2008; Sachsenhofer et al. 2009; Johnson et al. 2010). Oxygen levels were lowest during the deposition of the lower part of the Maikop Series (Hudson et al. 2008), but generally ranged from anoxic to suboxic (0.0–0.2 ml/l O2; Hudson et al. 2008). Anoxia is associated with stagnant water columns, which are generally related to decreased basin circulation (Tyson & Pearson 1991; Soták 2010). The onset of Maikop sedimentation thus marks a major change in the palaeoenvironmental conditions in the Paratethys basin, which is most likely to be related to the restriction of water circulation between the Paratethys and the open ocean (Steininger & Wessely 2000; Schulz et al. 2005; Hudson et al. 2008; Jovane et al. 2009; Johnson et al. 2010).

First published online October 29, 2015, https://doi.org/10.1144/SP428.3
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The prevailing hypothesis for the Maikop Series places the onset of anoxia in the Late Eocene to Early Oligocene, with anoxic conditions continuing into the Early Miocene (Abrams & Narimanov 1997; Hudson et al. 2008; Popov et al. 2008). There are two main processes that can explain a restriction event in this time interval: (1) tectonic uplift related to plate convergence between Arabia and Eurasia, leading to the disconnection of the marine gateways; and (2) a major sea-level drop.
at the Eocene–Oligocene Transition (EOT) at 33.9 Ma (e.g. Steininger & Wessely 2000; Zachos et al. 2001; Schulz et al. 2005; Jovane et al. 2009).

Precise dating of the Maikop Series could potentially distinguish between these geodynamic and climatic processes. If the onset of anoxia exactly coincides with the EOT, glacio-eustatic sea-level lowering is a very likely cause. Anoxia might have, in turn, increased carbon burial, leading to further global cooling. If, however, the onset of anoxia had already started during the Eocene, the tectonic collision of Africa and Arabia with Eurasia may be a more likely cause for the restriction of water exchange. The timing of this collision is still poorly constrained, but it probably took place between 39 and 20 Ma (Ballato et al. 2011; McQuarrie & van Hinsbergen 2013). If collision occurred around the EOT, the respective roles of climate and tectonics will be difficult to distinguish.

This research aimed to date the onset of Maikop deposition in the Talysh Mountains of Azerbaijan (Figs 1b, c & 2) to resolve the mechanism that initiated the anoxic conditions in this part of the Paratethys basin.

A thick succession of volcanic rocks is present underneath the Maikop sediments in the Talysh, linked to the Alborz Magmatic Arc (AMA) in the east (Vincent et al. 2005; Asiabanha & Foden 2012) and to the Adjara–Trialet zone of Turkey and Georgia in the west (Golonka 2004; Brunet et al. 2003; Asiabanha & Foden 2012). Much controversy exists about the tectonic setting of these volcanic belts (e.g. Yilmaz et al. 2000). Two contrasting settings are proposed: an arc setting and a back-arc setting.

A multidisciplinary stratigraphic approach based on magnetostratigraphy, biostratigraphy and X-ray fluorescence (XRF) analyses has been used to obtain constraints on the timing of the onset of anoxia and the tectonic setting in which the volcanic rocks were deposited. Furthermore, samples of various lithologies (sandstones, siltstones and shales) were taken for biostratigraphic analyses. Ar–Ar dating was performed on basalts from the two formations that contained volcanic rocks.

**Geological background**

The Talysh Mountains are situated in the southernmost part of Azerbaijan (Figs 1a & 2), bordering the Caspian Sea and Iran. The Talysh Mountains are located within the active collision zone between Arabia and Eurasia and continue into the Iranian

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**Fig. 2.** Regional map showing the distribution of Eocene volcanic rocks (after maps of Nalivkin 1976; Allen & Armstrong 2008; Agard et al. 2011) and the names of relevant areas.
Talesh, which form the western part of the Alborz mountain belt. The Alborz Mountains pass into the Kopet Dagh in the east. The lithology of the Talysh is composed of a succession of predominantly Eocene–Oligocene volcanic rocks and sediments. The volcanic rocks are mainly high-K alkali basalts, which, according to Asiabanha & Foden (2012), Golonka (2004) and Vincent et al. (2005), formed in a back-arc system. Their geochemical signatures (Nb troughs, Th spikes, large ion lithophile element (LILE) and Sr enrichment, and relative heavy rare earth element depletion), however, are characteristic of magmatic arcs (Vincent et al. 2005; Verdel et al. 2011; Asiabanha & Foden 2012).

To the SE of the Talysh (in Iran), another magmatic belt of Eocene age has been reported. This is the Urumieh–Dokhtar Magmatic Arc (UDMA), which runs from Urmia in the NW of Iran to Bazman in the SE (Fig. 2). This belt consists of subduction-related magmas (e.g. Ghorbani 2006; Verdel et al. 2011). The subduction of the Arabian plate underneath the Eurasian plate took place along the Zagros–Bitlis suture zone (Agard et al. 2011; Asiabanha & Foden 2012; Mouthereau et al. 2012).

In the Talysh, the Paleocene and Eocene rocks mostly consist of volcanic rocks and volcanogenic sediments, but a consensus on their age is still lacking (Alizadeh et al. 2005; Vincent et al. 2005) (Fig. 1b, c). According to Alizadeh et al. (2005), the lowermost Palaeogene unit is the Astara formation of Palaeocene age. It consists of tuffaceous sandstones, siltstones and mudstones, which are unconformably overlain by the volcanic rocks of the Kosmalyan formation (Lower Eocene). However, Vincent et al. (2005) infer that the Astara formation is a lateral equivalent of the Kosmalyan formation. $^{40}$Ar–$^{39}$Ar dating on samples from a sill intruded into the Astara formation constrains its lower part as older than 40.7 Ma (Vincent et al. 2005).

The volcanic rocks of the Kosmalyan formation are overlain by sedimentary rocks (sandstones, siltstones and shales) of the Neslin formation and a second volcanic unit; the Peshtasar formation (Fig. 3). The volcanic Peshtasar rocks are considered to be partly equivalent to the sandstone-dominated Arkevan formation. The thickness of the Arkevan formation varies considerably, showing a thickening trend towards the east (Vincent et al. 2005). The transition from the sandstone-dominated Arkevan formation to the mudstone-dominated Pirembel formation of the Maikop Series is marked by the sudden disappearance of metre-thick sandstone beds (Fig. 4). The Pirembel formation contains many dark grey to black very fine silt layers. Shales in the classical sense (that have a mud grain size) are scarce. The study of Vincent et al. (2005) presents a more detailed (sedimentological) description of the formations in the Talysh Mountains.

Sections and lithology

Several long and continuous outcrops of the Maikop Series are exposed in the Talysh Mountains of Azerbaijan. We sampled three sections and follow the nomenclature of Vincent et al. (2005). The locations of all sampled sections are plotted on the geological maps of Vincent et al. (2005) and Alizadeh et al. (2005) (Fig. 1b, c). This shows that section AZ15 represents the onset of Maikop sediments in both stratigraphic concepts. Section AZ16 covers the entire volcanic Peshtasar formation, with the lowest sampled flow, AZ16A, just above the sediments of the Neslin formation (Fig. 3).

AZ14 is located near Tilyakend (38.948605°N, 48.509431°E) and consists of sandstones, siltstones and basalts. The section consists of several volcanic sills intercalated in the sands of the Eocene Arkevan formation (Vincent et al. 2005). The same section is considered by Alizadeh et al. (2005) to be of Oligocene age (Fig. 1b, c).

Section AZ15 (38.953597°N, 48.381920°E) is a sedimentary succession with sandstones, siltstones and shales (Fig. 4). The Arkevan–Pirembel transition is marked by a change in lithology at 118 m in our sampled section. Above this level, hardly any thick sand layers are present, whereas they are abundant below. Sands that are 1 m thick or more are horizontally exaggerated in Figure 4 to show that the thick sand beds decrease in thickness and abundance higher up in the stratigraphy.

Section AZ16 (for GPS coordinates, see Fig. 5) consists mostly of Peshtasar basalts, with very few sedimentary beds in between. No lithostratigraphic log has been made because the different volcanic units are hard to distinguish in the field. The section is made up of basalts (pillow basalts according to Vincent et al. 2005) with some interbedded sediments (allowing correction for bedding tilt). Blocks of sediments and sedimentary intercalations within the section were observed between sites AZ16F, AZ16G and AZ16H (see also Fig. 3) and between AZ16K and AZ16L, possibly indicating periods of magmatic quiescence.

Materials and methods

Over 500 standard palaeomagnetic samples (25 mm diameter cores) were drilled using a gasoline-powered drill and an electrical drill. The cores were cut in the laboratory into standard (22 mm) specimens. Basalt samples with a high natural remanent magnetization were subsequently cut in half.

Palaeomagnetism

The magnetic susceptibility at room temperature of every second sample of AZ15 was measured using
an AGICO KLY-3 Kappa bridge. Measured values for the susceptibility were mass-normalized. Changes in magnetic susceptibility throughout a section may reflect variations in the lithology caused, for example, by variations in detrital input (Hay 1996, 1998; Ellwood et al. 2000) or variations in climate and environment resulting in changing diagenetic conditions (e.g. Da Silva et al. 2012).

To assess the magnetic carriers of the samples as well as chemical alteration temperatures, thermomagnetic analyses were carried out in air using a modified horizontal translation Curie balance (Mullender et al. 1993). The magnetization of powdered samples was measured as a function of temperature during six cycles up to increasingly elevated temperatures (700°C maximum; heating and cooling rate 10°C/min). The magnetic susceptibility of the samples was measured as a function of temperature on an AGICO KLY-3S susceptometer with a CS-3 furnace attachment (measurement frequency 976 Hz, field strength 400 A m⁻¹ peak level, noise level 2 × 10⁻⁷ SI). Measurements were performed on crushed basalt samples; again, six heating and cooling cycles were performed on powdered samples up to a temperature of 600°C.

A total of 107 samples were cored at section AZ14: at least seven samples per level were taken from five cooling units and sediments between the cooling units were also sampled. A total of 221 samples from sediments were analysed at section AZ15. A total of 26 levels were sampled in volcanic section AZ16, with at least seven samples per level. Each level was sampled in a different magmatic cooling unit. The specimens were subjected to thermal and/or alternating field (AF) demagnetization. Temperature increments of 20–60°C were applied to thermally demagnetize the samples in a shielded ASC oven up to a maximum temperature of 580°C for basalts and approximately 400°C for sediments to avoid thermal alteration as a result of the presence

![Fig. 3. Schematic logs of the three sampled sections of this study and photograph of two marker beds in the top of the Arkevan formation.](image-url)
of pyrite. Several samples were stepwise AF demagnetized after thermal demagnetization at 150, 175 or 200°C to remove the high coercivity components resulting from possible low temperature oxidation (weathering) (Van Velzen & Zijderveld 1995). Demagnetization steps of 5–100 mT were applied. Another part of the samples was demagnetized using three or five steps (20, 100, 150 and, sometimes, 175 and 200°C) of thermal demagnetization prior to AF demagnetization. The magnetization after each step was measured on a 2G Enterprise horizontal cryogenic magnetometer equipped with three DC SQUIDS (noise level 3 × 10⁻¹² Am²) with a robotized 2G DC-SQUID magnetometer.
The results were analysed using orthogonal vector diagrams (Zijderveld 1967) and the characteristic remanent magnetization (ChRM) directions were determined using principal components analysis (Kirschvink 1980). The mean directions were determined using standard Fisher statistics and standard errors were calculated following Butler (1992) and the criteria of Deenen et al. (2011b).

**40Ar–39Ar dating**

We selected ten cooling units for 40Ar/39Ar dating. The samples were processed using standard mineral separation techniques. The samples were first crushed, washed and dried. Heavy liquid separation was performed on size fractions between 250 and 500 µm with densities of 2.75 and 2.81 g/cm³ to remove the xenocrysts from the groundmass. In a few cases the 125–250 µm fraction was used. All the groundmass separates were finally hand-picked under a microscope. Samples and standards (Fish Canyon tuff sanidine; Kuiper et al. 2008) were wrapped in aluminium foil and irradiated for 18 hours in the High Flux Reactor, Petten, the Netherlands in the Cd-shielded RODEO-P3 position. Groundmass samples were loaded into 21-hole trays in 6 mm pits, placed in an ultra-high vacuum system and baked in two steps: for two days at 250 ± 8°C in a vacuum system at c. 10^{-2} mbar, followed by a one-day bake-out in the extraction line at c. 125 ± 8°C. The samples were heated stepwise using a Synrad 48–5 CO₂ laser and custom-made beam delivery system. The released gas was purified in a sample clean-up line designed in-house (St172, NP10 and Ti getters) and analysed on an MAP215–50 noble gas mass spectrometer fitted with a Balzers SEV217 detector. Mass discrimination was monitored regularly by three replicate runs of air pipettes at least every 12–24 hours. Blanks were run for every three to four unknowns. Ages were calculated using the in-house developed ArArCalc software (Koppers 2002) relative to the Fish Canyon tuff sanidine of 28.201 ± 0.23 Ma (Kuiper et al. 2008) using the decay constants of Min et al. (2001) (see discussion in Kuiper et al. 2008). The 40Ar/36Ar ratio of 298.56 of Lee et al. (2006) was used in the calculations. Errors are reported at the 2σ level.

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<td>-11.2</td>
<td>38.9545</td>
<td>48.07971</td>
<td></td>
</tr>
</tbody>
</table>

**Fig. 5.** Summary of palaeomagnetic results for section AZ16.
Micropalaeontology

Calcareous nannofossils were analysed from 25 samples of section AZ15. Smear slides were made following standard techniques (Bown & Young 1998). Quantitative analyses were unattainable as a result of the extreme scarcity of the assemblage. To classify most of the species, an optical microscope at 1250× magnification was used with both crossed nicols and phase contrast. Analyses were extended for at least 400 fields of view.

Geochemistry

The chemical compositions of all the cooling units of sections AZ14 and AZ16 were determined at Vrije Universiteit in Amsterdam, the Netherlands, using XRF spectrometry. The samples were crushed using a jaw-crusher, then powdered using several agate mills. The powders were dried overnight in an oven at 110°C. For each sample, a bead was made for major element analysis. The samples were oxidized for 30 minutes in a muffle furnace at a temperature of 1000°C. The samples were weighed before and after heating to calculate the loss on ignition (LOI). After oxidation, approximately 1.0 g of each sample was mixed with four times this amount of lithium metaborate (LiBO₂) flux. The samples were subsequently shaken for ten minutes. Beads were created using the Perl’X3 furnace and pellets were created of the dried, unoxidized powder, which was mixed with a binding resin (EMU powder). About 4.5 g of sample were mixed with 10% EMU for 15 minutes. The pellets were pressed at a pressure of 20 tonnes. Beads and pellets were analysed on a Panalytical MagiX Pro (PW2440) instrument using different programmes of XRF. The accuracy and precision of the measurements were based on the measurement of four international standards analysed between 2003 and 2011 (Handley et al. 2007) (see http://www.falw.vu/~petrolab/xrf).

Results

Palaeomagnetism

Arkevan formation (AZ14). The palaeomagnetic results for both the sedimentary and volcanic samples of AZ14 are ambiguous. Many of the demagnetization diagrams showed a clustering at the high temperature/high coercivity steps and the magnetization did not decrease towards the origin. For these samples, no reliable directions or polarities could be determined (see Van der Boon 2013).

Arkevan–Pirembel transition (AZ15). The bulk magnetic susceptibility record of section AZ15 does not show any major trend (Fig. 4). The sampling resolution appears to be insufficient to observe any clear cyclic variations and a much higher resolution than the palaeomagnetic sampling (which was targeted only at the most promising lithologies) seems to be required. On average, the susceptibility is slightly higher in the upper half of the section (see the five-point moving average in Fig. 4). However, a lower resolution leading to a sampling bias in the lower half of the section could account for this effect.

Thermomagnetic runs of three different lithologies were performed (siltstones, grey mudstones and sandstones, Fig. 6n–p). The magnetization decreases gradually on heating and shows reversible behaviour up to c. 400°C. The magnetization increases up to a temperature of 500°C, after which it decreases up to a temperature of 580°C. For the siltstone sample (AZ15.26; Fig. 5n), a very slight decrease can be observed at 680°C, suggesting the presence of some hematite. The decrease in magnetization up to about 350–400°C can be explained by the breakdown of iron sulphides. The increase in magnetization from 400 to 500°C is caused by the oxidation of iron sulphides, most probably pyrite, converting to magnetite between 380 and 420°C. These newly formed magnetic minerals are removed at 580°C, confirming the formation of magnetite.

Representative Zijderveld diagrams of samples of section AZ15 (Fig. 6b–m) show variable demagnetization. Many samples show a random ‘spaghetti plot’ behaviour, mainly caused by very low intensities that are of the order of the noise level of the magnetometer (Fig. 6g, i), apart from a viscous component that is removed at the very first demagnetization step. For a number of samples it is not certain whether the normal polarity is of primary origin or results from a present-day field overprint (Fig. 6f). In distinguishing the ChRM from a recently acquired normal polarity, we have used the criterion of ‘gradual’ decay at higher fields or temperatures (probably the ChRM residing in magnetite or an iron sulphide) v. very fast decay at low fields/temperatures (probably a recent overprint). For samples yielding a reversed polarity, the interpretation is more straightforward, varying from ‘probably’ reversed (Fig. 6d, e) to clearly reversed (Fig. 6b, c).

In the magnetostratigraphic column (Fig. 6a), we have indicated when we consider that a sample carries a primary magnetization, depending on the described criteria and on the consistency of the magnetostratigraphy. If the results are inconsistent, of low quality and/or probably overprinted, we have not assigned any polarity signature (grey shaded levels in Fig. 6a). Plotting all the results in stratigraphic order reveals at least three polarity
Fig. 6. Interpretation and examples of characteristic remanent magnetization directions (ChRM) of AZ15 and Curie balance results. (a) Lithostratigraphic log of section AZ15 and interpretation of characteristic remanent magnetization. (b–m) Examples of samples that were assigned a reversed (b, c), reversed? (d, e), unknown (f, g, h, i), normal? (j, k), normal (l, m) polarity. (n–p) Curie balance results of samples of different lithologies: (n) siltstone; (o) grey mudstone; and (p) sandstone.
Fig. 7. Interpretation and examples of characteristic remanent magnetization directions (ChRM) of AZ16. (a) Zijderveld diagrams of samples with arrows showing interpreted components of alternating field demagnetized samples. Thermally demagnetized sample Zijderveld diagrams are shown for comparison. (b) Averages of interpreted declination and inclination per cooling unit in an equal area diagram. Three groups are apparent that match the stratigraphy of the sampled cooling units. A satellite image (Google Earth) of the location of cooling units is also shown. (c) Average declination and inclination v. the stratigraphic position for each cooling unit.
reversals. The basal part of the section shows dominantly normal polarities and a normal/reverse (R/N) polarity transition at 17 m. The next R/N transition occurs at 82 m, in the upper part of the Arkevan formation. Another N/R reversal occurs in the Pirembel formation at 350 m.

Volcanic rocks of the Peshtasar formation (AZ16). Specimens from all 26 levels were demagnetized using AF demagnetization up to 100 mT. Additional thermal demagnetization was applied to two specimens per level. The results of the AF and thermal demagnetization (Fig. 7a) are very similar. Many cooling units have multiple components; a low coercivity or low temperature component is usually removed after demagnetization up to 20 mT or 200–250°C. We consider this to be a viscous overprint, which is usually randomly directed and occasionally carries the present-day field. In thin sections it appears that a number of levels represent sills (Van der Boon 2013). The generally slower cooling rate of shallow intrusive rocks may therefore result in a longer record of the magnetic field. We have plotted the mean ChRM for all sampled levels in AZ16 in Figure 7. Occasionally, the samples contained a high coercivity/temperature component observed above 60 mT or 500°C, of which the character remains elusive.

When we considered all the ChRM directions from AZ16, it became evident that there were three distinct directional groups. These palaeomagnetic groups coincide with equally distinct groups in the stratigraphic order in the sampled sequence (colour-coded in Fig. 7b). The lower group (I) contains units A–K and the middle group (II) contains units L–P, T and U. The upper group (II) contains units Q–S and V–Z. According to the stratigraphic order, unit U belongs to group III, whereas it clearly belongs palaeomagnetically to group II. There is, however, very little stratigraphic distance between the flows in this interval and we sampled partly along-strike (see Fig. 7b). Therefore a slight unnoticed change in the lateral variation in the thickness
of the flows or the presence of sills can lead to a relative stratigraphic position of U in the upper group (II) instead of the middle group (III).

The present-day inclination of the region is approximately 58°. Little northward movement since the Eocene is expected, so the Eocene values should be roughly similar. The palaeomagnetic directions before tilt correction are very shallow for groups I and III (inclination ≈ 20.8° and ≈ 19.5°, respectively) and rather steep for group II (inclination ≈ 68.8°). A striking feature is that the tilt correction (as derived from the sedimentary beds below, within and above the volcanic sequence) causes the directions to become shallower. We are therefore not certain that tilt correction is warranted. This would apply especially to the sills that form the majority of the volcanic rocks in the sequence. It is unlikely that the mean ChRM directions of the three groups are caused by remagnetization because this would have led to similar directions for all three groups.

More striking is the large dispersion of the three groups, which is not often recorded as a result of palaeo-secular variation. We notice, however, that similar groups – each forming a rather tight cluster – with such a wide dispersion have been observed in Central Atlantic Magmatic Province lavas of Early Jurassic age in the High Atlas and the Argana Basin in Morocco (Deenen et al. 2011a; Knight et al. 2004).

40Ar−39Ar dating

Some of the cooling units that we aimed to date had already been analysed by Vincent et al. (2005) and can be directly compared with our data when the 40Ar/39Ar ages are recalculated using the calibration model of Kuiper et al. (2008). The lowermost cooling unit AZ16A is very close to sample TA92.6 of Vincent et al. (2005) with a recalculated age of 41.0 Ma (reported age 40.5 Ma). In the top part, AZ16Y is close to sample TA87.1 with a recalculated age of 38.8 Ma (reported age 38.3 Ma). The most significant species useful for biostratigraphy are Reticulofenestra dictyoda (phase contrast), XN, 1250; Cribrocentrum erbae, Isthmus sp., XN, 1250; Dicytococcos bicornatus, XN, 1250; Calcareous nannofossils. This assemblage is extremely scarce and preservation is moderate to poor. The most significant species useful for biostratigraphic assignment are Cribrocentrum erbae, Isthmolithus recurvus and Cribrocentrum reticulatum. Using these markers, the section can be placed within zones CNE 17 and CNE 19 of the biozonation (Agnini et al. 2014). The rare presence of C. erbae does not allow recognition of the acme of this species that defines the base and top of the CNE17 Zone, which otherwise is proposed as the best nannofossil bioevent to mark the base of the Priabonian (Agnini et al. 2011).

Fig. 9. Photographs of calcareous nannofossils. (1) Cribrocentrum reticulatum, crossed nicols (XN), 1250×, sample AZ15.46; (2) Lanternithus sp., XN, 1250×, sample AZ15.46; (3, 4) Lanternithus sp., CF (phase contrast), XN, 1250×, sample AZ15.46; (5) Dicytococcos bicornatus, XN, 1250×, sample AZ15.46; (6) Ericsonia formosa, CF, 1250×, sample AZ15.46; (7) Lanternithus sp., XN, 1250×, sample AZ15.46; (8) D. bicornatus, XN, 1250×, sample AZ15.46; (9) Dicytococcos bicornatus, XN, 1250×, sample AZ15.103; (10) Reticulofenestra dictyoda, XN, 1250×, sample AZ15.131; (11) D. bicornatus, XN, 1250×, sample AZ15.177; (12) Cyclococcos bicornatus, XN, 1250×, sample AZ15.46; (13–16) C. reticulatum, XN, 1250×, sample AZ15.46; (17) C. reticulatum, CF, 1250×, sample AZ15.46; (18) Lanternithus sp., XN, 1250×, sample AZ15.46; (19, 20) D. bicornatus, XN, 1250×, sample AZ15.46; (21, 22) Reticulofenestra sp., XN, 1250×, sample AZ15.46; (23, 24) Reticulofenestra dictyoda, XN, CF, 1250×, sample AZ15.146.
The presence of rare *I. recurvus* in few samples from 369.85 m is indicative of Zone NP 19 of Martini (1971) in the Priabonian. *Cribrocentrum reticulatum* is distributed from the lowest fossiliferous sample up to 500 m; its highest occurrence defines the top of CNE 19 of Agnini et al. (2014) dated at 35.24 Ma. Therefore the section is included in the interval spanning from Zone CNE17 to CNE19, which is correlated with Chrons 17n.2n to C15r.

Microphotographs of the most significant species are shown in Figure 9 and the results are summarized in Figure 10.

**Large benthic forams.** Several large benthic forams were found in the thick sand beds at the top of the Arkevan formation in section AZ15 (at c. 104 m). The following species were recorded: striate, small *Nummulites* sp., *Nummulites fabianii*, *Operculina* sp., *Orthophragminids* sp. and *Heterostegina reticulata*. Of these, the orthophragminids, *N. fabianii*, and *H. reticulata* all became extinct at the Eocene–Oligocene boundary. *Nummulites fabianii* is a marker for SBZ19–20 (Serra-Kiel et al. 1998), but it is difficult to separate it from its ancestral species. From the four specimens of *H. reticulata* it was possible to obtain details of the internal characteristics that are often used as stratigraphic indicators. The observed proloculus size ranged from 100 to 130 μm and the number of chambers without subdivision from three to four. This compares with similar characteristics seen in populations from the Vedi area in Armenia reported by Less et al. (2008). Photographs of the three species are shown in Figure 11. Uncertainty in the correlation of the SBZ biozones to the geological timescale or other biostratigraphic schemes (e.g. planktonic forams; Wade et al. 2011) prevent a more detailed age assignment than the early part of the Late Eocene. The best preserved specimen is an *Operculina* sp. (Fig. 11), which is an extremely long-ranging taxon. The large benthic forams do not show indications of mixing in the sense of the overlapping ranges of the present species. However, reworking might be an issue, as some of the grains are rather coarse and sands are not indicative of a typical large benthic foram environment.

**Geochemistry**

The generally low LOI, between 0.65 and 1.99 wt%, indicates that the samples were fresh. Four samples showed LOI values of 2.90, 2.33, 2.04 and 2.06 wt%. However, these values are still within the range of samples that are considered suitable for geochemical analyses (e.g. Handley et al. 2007).

The XRF results show that all the samples were very similar in chemical composition (Table 1). According to the IUGS classification of alkaline rocks, the samples are trachybasalts and basaltic trachyandesites. The samples have no normative
quartz, low Fe enrichment, a high alkali content (Na$_2$O + K$_2$O > 5%), high K$_2$/Na$_2$O, and are enriched in P, Rb, Sr, Ba, Pb and light rare earth elements. Furthermore, the samples show a low TiO$_2$ (<1.3%) content and have a high, variable Al$_2$O$_3$ (14–19%) content. Thus the samples show shoshonitic compositions (Morrison 1980). The samples do not show a trend in major element composition because the trends (plotted either v. SiO$_2$ or stratigraphic level) of all the major elements are generally flat (see Fig. 12a).

The N-MORB normalized spider diagram (see Fig. 12c) shows that (compared with the average N-MORB; Sun & McDonough 1989), all samples exhibit an Nb trough, a Th spike, an enrichment in the LILEs (Rb, Ba, K, Pb and Sr) and a relative depletion in heavy rare earth elements (Dy, Y, Er and Yb).

Samples from AZ14 and AZ16 plot within the same reach and we can therefore assume that they represent the same formation, which is consistent with the observations of Vincent et al. (2005). Notably, the three groups that were observed in AZ16 in terms of magnetic signal are also observed in the plots of many elements and even more so from ratios. However, they are most clear in the plots of metals such as Th, U, Zr and Nb when plotted v. SiO$_2$ content (see Fig. 12b for the example of Th). From these plots we concluded that the volcanic rocks from AZ14A and AZ14B belong to the middle unit (II) of section AZ16. Furthermore, AZ14C, AZ14D and AZ14F belong to the upper unit (III) of section AZ16. Sample AZ16A is an outlier in many geochemical plots. This lowermost flow was sampled just above the sands of the Neslin formation, so it might have experienced some crustal or sediment contamination.

Discussion

Onset of Maikop sedimentation in the Talysh Mountains

In the Talysh Mountains, the onset of the Maikop Series is marked by the Arkevan–Pirembel transition, located at the level of 118 m in section AZ15. The Arkevan–Pirembel transition occurs in the second interval of normal polarity. Our interpretation of the combined palaeomagnetic and biostratigraphic dataset correlates the lowest reversed interval to C17n.1r (Fig. 13), as this is the only reversed interval within CNE17. This yields an age for the corresponding stratigraphic level at the R/N transition (at 82 m) of 37.8 Ma. The base of common *I. recurvus* occurs within Zone CNE18 in the upper part of upper Chron C17n.1n (Fornaciari et al. 2010) or in the lowermost part of Chron C16r (Agnini et al. 2014). In our AZ15 record, *I. recurvus* first appears in samples 15.103 and MP7, which correspond to a reversed polarity interval. Consequently, we consider it most likely that the N/R polarity transition at c. 350 m correlates to the base of Chron 16r at an age of c. 37 Ma. It should be noted, however, that we cannot fully exclude alternative correlations as the distribution of the calcareous nannofossil marker species theoretically spans the interval from Zones CNE17 to CNE19, which are correlated with Chrons 17n.2n to C15r.

There is roughly 270 m of stratigraphy within the interval between these polarity transitions, leading
to an average sedimentation rate of approximately 34 cm ka$^{-1}$. Taking the sedimentation rate of 34 cm ka$^{-1}$, the age at the top of the sample section can be calculated through extrapolation, leading to an age of around 36.5 Ma. Consequently, the section is estimated at an age of around 36.5 Ma. The base of the section can be calculated through extrapolation, leading to an age of around 38 Ma. Vincent et al. (2005) mention a thickness for the entire Maikop Series in the Talysh Mountains of 1500 m, which suggests it should be almost entirely of Eocene age. The detailed overview of the stratigraphy of the Talysh by Vincent et al. (2005) does not report any evidence for Oligocene fauna. A Late Eocene age for the base of the Pirembel formation is in agreement with the study of Amini (2006), which concludes that the Ojagheshlagh formation just across the border in Iran is equivalent to the Pirembel formation and starts during the Late Eocene.

The age of 37.7 Ma for the onset of Maikop sedimentation in the Talyshev proves that it is not related to the major sea-level lowering at the EOT (33.9 Ma, Vandenberghe et al. 2012). The age instead indicates a tectonic cause, further evidenced by the end of a major episode of volcanism within the Arabia–Eurasia collision zone. This may also indicate anoxic conditions in the Paratethys is most probably a complex combination
of regional tectonic activity and eustatic sea-level changes. For a better understanding of the controls on widespread anoxia in the Paratethys, the onset of Maikop sedimentation should be dated in high resolution in other regions. The question remains as to whether these sediments truly represent the equivalent of the Maikop sediments as stated in the studies of Azizbekov et al. (1979), Alizadeh et al. (2005) and Vincent et al. (2005).

Typical Maikop lithology is represented by laminated dark grey to dark brown clays, often containing jarosite and fish fossils (Saint-Germes et al. 2000; Hudson et al. 2008; Johnson et al. 2010). These characteristics are not observed in section AZ15 as these consist mostly of dark grey siltstones. Fish fossils and jarosite were not observed during our study. Stratigraphic correlation of the Maikop sediments in general is difficult, as there is an absence of levels that can be used as marker beds throughout the Eastern Paratethys. Cocolith-bearing limestone levels that are reported from the Western Paratethys (e.g. Haczewski 1996; Schulz et al. 2005; Ciurej & Haczewski 2012) are, to our knowledge, not reported from the Talysh.

Three volcanic episodes of the Peshtaslar formation

Three groups are evident in the Peshtaslar formation from a palaeomagnetic, stratigraphic and geochemical point of view. The palaeomagnetic results show that, within each group, there is dispersion that can be explained by secular variation based on the A95 values that fall within the N-dependent A95 envelope of Deenen et al. (2011b). Hence the three directional groups must each represent a
Fig. 12. Geochemistry results. (a) Weight percentage of oxides v. magnesium number (the ratio of magnesium to iron). Trends in this diagram are flat, indicating a lack of magma evolution. (b) Abundance of Th, which clearly shows three groups, as well as a plot of Th/Ta v. Nb/Ta. (c) Spider diagram, with arc- and back-arc compositions in grey (Pouclet et al. 1994).
Fig. 13. Magnetostratigraphy and nannoplankton correlation to the geological polarity timescale. The nannoplankton zones of Agnini et al. (2014) have been correlated with the polarity timescale of Cande and Kent (1995). Reversal ages have changed slightly for the polarity zones, according to the new the geological polarity timescale of Gradstein et al. (2012). This does not influence the correlation itself. For legend of palaeomagnetism, see Figure 6.
period long enough (on the order of several kyr) to record some secular variation. Because the three groups yield very different mean ChRM, it is unlikely that the Peshtasar formation was deposited by a process of uniform extrusion over 2 myr (based on the Ar–Ar ages of Vincent et al. (2005)). Instead, the Peshtasar formation was probably deposited during three short-lived magmatic pulses, with a maximum duration of several thousands of years, considering there is some secular variation in each group. This results in relatively high eruption rates because group I spans 1.5 km stratigraphically.

Arc volcanism in the Talysh

Magmas very rarely reach the surface without experiencing processes that modify their chemical composition. Magmas are usually affected by processes such as crustal assimilation, fractional crystallization and partial melting or magma mixing, processes that lead to evolved compositions. As all the trends of major elements in our observations of the Peshtasar lavas are almost flat (Fig. 12), these processes are unlikely to have severely affected the composition of the magma. This indicates a lack of magma evolution, which is consistent with the hypothesis that

Fig. 14. (a) K/Rb v. Ce/Pb for the Mariana arc and back-arc compared with Gribble et al. (1998). (b) Th/Yb v. Nb/Yb compared with the Japan arc and back-arc of Poulet et al. (1994). For legend of samples, see Figure 12.

Fig. 15. Distribution of Eocene volcanic rocks in Iran (modified after Agard et al. 2011). (a) Locations of the Urumieh–Dokhtar Magmatic Arc (UDMA) and Alborz Magmatic Arc (AMA). (b) Our hypothesis in which Eocene volcanic rocks of the UDMA and AMA are part of one large belt. For legend, see Figure 2.
the three groups represent short-lived episodes of volcanism.

The N-MORB normalized trace element abundance patterns (see Fig. 12) of the samples are very similar to those of arc volcanoes found in other parts of the world (e.g. Handley et al. 2007). The anomalies in Nb, Ta and Zr are also clear evidence for magma generation related to subduction (Baier et al. 2008). Back-arc volcanism would potentially have the same signature (as back-arcs are always related to subduction). Back-arc basin basalts (BABBs) usually vary in geochemical composition anywhere between MORB-like signatures and arc-like signatures.

We compared our geochemical data with data from other arc/back-arc systems. In Figure 14a, the relevant trace element ratios (K/Rb and Ce/Pb) are plotted, together with data from the Mariana arc and the Mariana back-arc (Gribble et al. 1998). The Mariana arc was chosen because of the clear relative position of the arc and back-arc. Furthermore, late Cretaceous volcanism along the active margin of Eurasia had a geochemical signature resembling that of the Mariana arc (Kazmin et al. 1986). The comparison shows that all the samples fall entirely within the range of samples from the arc and none of the samples from the Talysh plots within the range of the Mariana back-arc. The Ce/Pb ratios may reflect a contamination with sediment, additional isotope studies are required for more insight into crustal contamination.

In Figure 14b, the ratios of Th/Nb v. Nb/Yb are plotted to compare different types of basalts (arc basalts, BABBs and oceanic island basalts) from the Japan Sea, in which the relative position of basalts, BABBs and oceanic island basalts are plotted to compare different types of basalts (arc pestetas samples all plot in or near the ARC field, characteristic of continental-arc basalts (Pouclet et al. 2009). Back-arc volcanism would potentially have the same signature (as back-arcs are always related to subduction). Back-arc basin basalts (BABBs) usually vary in geochemical composition anywhere between MORB-like signatures and arc-like signatures.

The spider diagram patterns (Fig. 12) are characteristic of continental-arc basalts (Pouclet et al. 1994; Asiabana et al. 2009; Verdel et al. 2011; Asiabana & Foden 2012). Arc-like signatures have typical Nb troughs, Th spikes, LILE and Sr enrichment and relative HREE depletion. All the samples in this study exhibit these characteristics. Back-arc basin characteristics include a low LILE content, no Th anomaly and a flat REE pattern. None of the basalts in this study therefore represents BABBs in the geochemical sense. When the patterns are compared with the arc lavas and BABBs of Pouclet et al. (1994) (see Fig. 12c), they are much more similar to the arc lavas than the back-arc basalts. Spider diagrams from Asiabana & Foden (2012), Vincent et al. (2005) and Verdel et al. (2011) show the same patterns (note that normalization with N-MORB or primitive mantle has little effect on the patterns). The scenario that Golonka (2004) proposes, in which the volcanism of the Talysh and AMA is linked to ocean spreading in a back-arc basin, is therefore unlikely. These results are consistent with the continental-arc setting of the AMA as proposed by Asiabana et al. (2009). All the samples from the Peshtasar formation exhibit shoshonitic compositions. Shoshonitic compositions of Eocene volcanic rocks have also been reported from the UDMA and the NW of Iran (e.g. Torabi 2009; Dilek et al. 2010; Verdel et al. 2011). Shoshonites are usually found in subduction-related arcs, which often exhibit a progression along the tholeiite–calc-alkaline–shoshonite path. Shoshonites are generally formed later, stratigraphically higher and are associated with greater heights above the Benioff zone. Therefore they are usually found in mature arcs. In regions that are tectonically active and unstable, shoshonites are associated with deformation of the arc, near the termination of subduction, or when there is a transition between two subduction regimes that have a different orientation (Morrison 1980). Shoshonites are sometimes associated with slab break-off (Dilek et al. 2010; Verdel et al. 2011). It is possible that the shoshonites of the Talysh represent short-lived episodes of volcanism caused by the end of subduction or break-off of the Neotethys slab. This is in agreement with the scenario as proposed by Agard et al. (2011), which involves Paleocene–Eocene slab break-off, large shifts in arc magmatism and episodes in which upper crustal extension takes place.

Much controversy exists about the distribution of Eocene volcanic rocks in Azerbaijan and Iran. Many researchers infer the presence of two volcanic belts, the AMA (extending from the Talysh Mountains via the Alborz to the Kopet Dagh) and the UDMA (extending from near Lake Urmia to the SE of Iran) (Fig. 15a). These two belts do not entirely explain the distribution of the Eocene volcanic rocks, as they are lacking in parts of both the AMA and UDMA, especially in the regions near Urmia and east of Tehran (Fig. 15). Regarding the distribution patterns of Eocene volcanic rocks and geochemical arc signatures, we hypothesize that a different scenario may be plausible. The distribution of Eocene volcanic rocks can be traced almost continuously from the Talysh via the western part of the Alborz towards the southeastern part of Iran (Fig. 15b). In this scenario, Eocene volcanic rocks in the Kopet Dagh are not part of the same system, and may have different geochemical signatures.

Conclusions

We have used magnetostratigraphy and nanoplankton biostratigraphy to determine the age of the transition from the Arkevan formation to the
Pirembel formation. This transition has been assumed to represent the onset of Maikop sedimentation within the Talysh Mountains of Azerbaijan. A magnetostratigraphy was obtained for the upper part of the Arkevan formation and the lower part of the Pirembel formation, showing a polarity pattern of two normal and two reversed polarity intervals. This magnetostratigraphy was combined with a biostratigraphy based on nannoplankton assemblages that are characteristic for zones CNE17 and CNE18/19 (Late Eocene). The onset of Maikop sedimentation occurs in the lower part of C17n. In at an age of 37.7 Ma. The estimated sedimentation rate for the Late Eocene sediments of the Talysh is 34 cm kyr⁻¹, which suggests that Oligocene sediments are absent in the Talysh.

The onset of Maikop sedimentation in the Talysh started almost 4 myr before the occurrence of the EOT, which indicates a tectonic cause for the decrease in grain size observed at the Arkevan–Pirembel boundary. Also indicative of an active tectonic regime is the presence of approximately 2 km of subduction-related volcanic rocks of the Peshtasar formation, stratigraphically just below the Arkevan and Pirembel formations. The question remains whether the Pirembel sediments truly represent the equivalent of the classic Maikop Series that are deposited in the more distal parts of the Paratethys. Palaeomagnetic analyses of the volcanic rocks of the older Peshtasar formation in section AZ16 show three distinct groups that are also recognised as subsequent intervals in the stratigraphy. They are interpreted to represent three short time intervals (c 10⁶–10⁷ years), each interval recording some secular variation over a relatively short time. Deposition rates must have been very high because the lower group (I) has a stratigraphic thickness of 1.5 km. The three groups are also distinct in their geochemistry. Based on equal geochemical composition, we confirm the lateral relationship between the Arkevan and Peshtasar formations. Volcanic rocks of the Peshtasar formation are all trachybasalts and basaltic trachyandesites with trace element signatures characteristic of arc lavas. The volcanic rocks probably formed in a continental-arc setting. The major element geochemistry shows that the samples have shoshonitic compositions.

We would like to thank two anonymous reviewers and the Editor, M. Sosson, whose comments have greatly improved this paper. This work was financially supported by the Netherlands Geosciences Foundation (ALW) with support from the Netherlands Organization for Scientific Research (NWO) through the VICI grant of WK. This research was partly financed by the Molengraaff Fund and the DARIUS Programme. MIMM acknowledges funding and support from the Henri Poincaré Fellowship (Observatoire de la Côte d’Azur, Nice), the DARIUS Programme, and the College of Science and Engineering and the Department of Earth Sciences at the University of Minnesota. The fieldwork was carried out with the help of Eldar Huseynov, Gingiz Aliyev and Kamram Aliyev (Geological Institute of the Azerbaijan National Academy of Sciences, Baku). R. van Elsas, S. Matveev and C. Bontje (VU Amsterdam) are thanked for their help with the geochemical analyses and Ar–Ar preparation.

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