Progressive orocline formation in the Eastern Pontides–Lesser Caucasus

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Abstract: The Eastern Pontides–Lesser Caucasus fold–thrust belt displays a peculiar northwards arc-shaped geometry that was defined as an orocline in earlier studies. The Lesser Caucasus was affected by two main tectonic events that could have caused orocline formation: (1) Paleocene–Eocene collision of the South Armenian Block with Eurasia; and (2) Oligocene–Miocene Arabia–Eurasia collision. We tested the hypothesis that the Lesser Caucasus is an orocline and aimed to time the formation of this orocline. To determine the vertical axis rotations, 37 sites were sampled for palaeomagnetism in rocks of Upper Cretaceous–Miocene age in Georgia and Armenia. In addition, we compiled a review of c. 100 available datasets. A strike test was applied to the remaining datasets, which were divided into four chronological sub-sets, leading us to conclude that the Eastern Pontides–Lesser Caucasus fold–thrust belt forms a progressive orocline. We concluded that: (1) some pre-existing curvature must have been present before the Late Cretaceous; (2) the orocline acquired part of its curvature after the Paleocene and before the Middle Eocene as a result of South Armenian Block–Eurasia collision; and (3) about 50% of the curvature formed after the Eocene and probably before the Late Miocene, probably as a result of Arabia–Eurasia collision.

Supplementary material: Results from rock magnetic experiments, reversal and fold tests and equal area projections of the characteristic remanent magnetizations for each site, as well as biostratigraphic ages and a table with palaeomagnetic results from the literature review (with assigned numbers referred to in the text) are available at http://www.geolsoc.org.uk/SUP18852.

The Eastern Pontides–Lesser Caucasus fold–thrust belt (Fig. 1a) forms part of the former Eurasian margin that deformed as a result of ongoing Africa(–Arabia)–Eurasia convergence. The peculiar arc-shaped geometry of the Eastern Pontides–Lesser Caucasus, which is expressed in the change
of strike of folds and thrusts, typifies a wider region from the Central Pontides to the Talysh and Alborz mountain belts (see red broken line, Fig. 1a). Based on palaeomagnetic data (Bazhenov & Burtman 2002; Hisarli 2011), the northwards arc-shaped geometry in the Central Pontides is an orocline – a bend imposed on a pre-existing, formerly linear orogenic belt (Carey 1955). To the west, the Central Pontides orocline was formed in the latest Cretaceous to Paleocene as a result of the accretion of the Central Anatolian Crystalline Complex to the (rheologically contrasting) Eurasian margin (Meijers et al. 2010; Lefebvre et al. 2013). Likewise, the arcuate fold–thrust belts of the Talysh and Alborz ranges could have formed by the indentation of the Arabian plate into the South Caspian Basin (Agard et al. 2011) or as a result of strong contrasts in rheology between the South Caspian Basin (Fig. 2) and the Talysh and Alborz ranges (Fig. 1a; Allen et al. 2003). The latter was confirmed for the Alborz range by Cifelli et al. (2015), who determined that orocline formation was initiated after c. 7.6 Ma (Late Miocene). The Eastern Pontides–Lesser Caucasus fold–thrust belt, the subject of this study, was defined as an orocline based on palaeomagnetic data from Upper Cretaceous and Eocene deposits (Bazhenov & Burtman 2002; Hisarli 2011). In their compilations, Bazhenov & Burtman (2002) and Hisarli (2011) did not find any significant difference in the vertical axis rotations between their Late Cretaceous and Eocene data. Consequently, they pooled these together and

Fig. 1. (a) Tectonic map of the Black and Caspian Sea region. AF, Araks Fault; ASHS, Amasia-Sevan-Hakari Suture; ATB, Anatolide–Tauride Block; CACC, Central Anatolian Crystalline Complex; CAF, Central Anatolian Fault; EAF, East Anatolian Fault; Ist. Z., Istanbul Zone; IAES, Izmir–Ankara–Erzincan Suture; GC, Greater Caucasus; MM, Menderes Massif; NAF, North Anatolian Fault; R, Lake Rezaiyeh; SAB, South Armenian Block; SSZ, Sanandaj–Sirjan Zone; TC, Transcaucasus; V, Lake Van. Broken red line roughly indicates the general strike and the strike of the fold axes in the Central Pontides, Eastern Pontides–Lesser Caucasus, Talysh and Central Alborz Mountains fold–thrust belts. Black circular arrows in the Central Pontides, Eastern Pontide–Lesser Caucasus and Central Alborz Mountains indicate the sense of rotation in both limbs of the orocline. The Anatolide–Tauride Block, Istanbul Zone, Sakarya, South Armenian Block and Transcaucasus refer to basement units. Modified after Avagyan et al. (2005) and Sosson et al. (2010). (b) Structural map of the Lesser Caucasus–Eastern Pontides–Northeast Anatolides regions after Hässig et al. (2014a).
Fig. 1. Continued.
concluded that oroclinal bending occurred after Late Cretaceous to Eocene times, but without really quantifying the oroclinal test. No data from younger deposits were presented in either study and therefore the timing as well as the cause of the vertical axis rotations remains unconstrained. Orocline formation could have resulted from two latest Cretaceous and younger major deformation events: (1) the collision of the South Armenian Block (SAB) with the Eurasian margin (i.e. the Transcaucasus basement; Fig. 1a); and (2) the collision of Arabia with the Eurasian margin. Estimates of the age of SAB–Eurasia collision range from the end of the Campanian (Rolland et al. 2012) to Paleocene–Eocene (Robinson et al. 1995; Okay & Şahintürk 1997; Sosson et al. 2010; Robertson et al. 2014). A palaeolatitude study on SAB and Eurasian margin sediments (Meijers et al. 2015) could not distinguish between the collision ages proposed by Rolland et al. (2012) and Sosson et al. (2010). Arabia–Eurasia collision estimates vary widely and range from the Late Cretaceous to Early Miocene, although most researchers agree that collision started during or after the latest Eocene (e.g. Berberian & King 1981; Axen et al. 2001; Allen et al. 2004; Ritz et al. 2006; Guest et al. 2007; Vincent et al. 2007; Allen & Armstrong 2008; Ballato et al. 2011; Moutheureau 2011; Rezaeian et al. 2012; Madanipour et al. 2013).

Constraining the age of orocline bending may therefore enable speculation about which of these collisional events (i.e. SAB–Eurasia or Arabia–Eurasia collision) was responsible for orocline formation. To constrain the timing of orocline bending in the Eastern Pontides–Lesser Caucasus fold– thrust belt, we sampled 37 new sites in rocks of Late Cretaceous to Late Miocene age in the Lesser...
Caucasus (Georgia and Armenia) for palaeomagnetic analysis. We combined our new data with a compilation of existing data from Cretaceous to Pleistocene strata in the area of interest – including northeastern Turkey, Georgia and Armenia – from the International Association of Geomagnetism and Aeronomy Global Paleomagnetic (GPMDB) Database (http://www.ngu.no/geodynamics/gpmdb/). The data from the GPMDB are often of unknown quality and, in some cases, the data have not been published, which led Bazhenov & Burtman (2002) to exclude all GPMDB data from their review. We included all c. 100 of these datasets in our review and applied a number of reliability criteria. The data that were accepted after applying these criteria were taken into account for further analysis. If a coherent temporal and spatial declination pattern emerges from such a large amount of data, it can be assumed that the data must have regional tectonic significance.

Geology of the Eastern Pontides and Lesser Caucasus

The geological and tectonic history of the Caucasus, situated between the Black Sea to the west and the Caspian Sea to the east (Fig. 1a), is mainly determined by its position between the still-converging Eurasian and African–Arabian lithospheric plates within a wide zone of continent–continent collision. During the Late Palaeozoic to Early Cenozoic, the region belonged to the now-vanished Tethys ocean and its southern Eurasian and northern Gondwana margins. Within the Africa–Eurasia convergence zone, a system of island arcs, intra-arc rifts and back-arc basins formed. Microplates drifted from the African margin in the south and accreted to the Eurasian margin in the north following the successive opening and closure of the oceanic domains of the Proto-Tethys, Palaeo-Tethys and finally Neo-Tethys ocean (e.g. Kazmin et al. 1986; Ustaömer & Robertson 2010; Adamia et al. 2011). During the collisional stages of the late Alpine tectonic cycle resulting from Arabia–Eurasia collision, back-arc basins were inverted into the fold–thrust belts of the Greater and Lesser Caucasus (Fig. 2; Forte et al. 2010; Vincent et al. 2014). In the west and east, the Rioni and Kura Basins separate the Greater and Lesser Caucasus (Fig. 2; Adamia et al. 2011).

Geographically, the Lesser Caucasus is the mountain belt running through Georgia, Armenia and Azerbaijan south of the Kura and Rioni basins and roughly north of the Araks Valley (Fig. 2). Across the Georgian–Turkish border in the west, it continues into the Eastern Pontides. North of the Izmir–Ankara–Erzincan and Amasia–Sevan–Hakari suture zones (Fig. 1a), the basement units underlying the Eastern Pontides–Lesser Caucasus fold–thrust belt are the Sakarya Zone and the Transcaucasus. South of the Izmir–Ankara–Erzincan and Amasia–Sevan–Hakari suture zones, the basement is formed by the Anatolide Tauride Block and SAB of Pan-African affinity (e.g. Belov & Sokolov 1973; Kröner & Şengör 1990). The Sakarya Zone and the Transcaucasus terranes accreted to the Eurasian margin in the Palaeozoic (Saintot et al. 2006; Okay 2008), whereas the SAB of Pan-African affinity started colliding with the Eurasian margin either in the Late Cretaceous (Rolland et al. 2012) or in the Palaeocene–Eocene (Robinson et al. 1995; Okay & Şahintürk 1997; Sosson et al. 2010; Robertson et al. 2014).

The Sakarya and Transcaucasus terranes (Fig. 1a) are often treated separately, which is probably due to their division over several countries: Turkey in the west (Eastern Pontides) and three former Soviet countries in the east (Georgia, Armenia and Azerbaijan; Lesser Caucasus). The Sakarya and the Transcaucasus basalts are both limited by the Scythian Platform (i.e. the thinned margin of the East European Platform) to the north (Yılmaz et al. 2000; Hässig et al. 2014a).

The northwards drift of Africa led to the collision of its Arabian promontory with Eurasia. Thermochronometric studies by Vincent et al. (2007) and Madanipour et al. (2013) provide evidence for Early Oligocene Arabia–Eurasia collision in the Western Greater Caucasus and Talysch mountains, respectively, close to our study area. Ongoing convergence in the region is one of the driving forces for the westwards escape of the Anatolian microplate along the North Anatolian Fault (Fig. 1a; Dewey & Şengör 1979; Şengör et al. 1985) and continuing deformation in the Lesser and Greater Caucasus (Jackson 1992). The global positioning system (GPS) velocity vectors in the collision zone between Arabia and Eurasia (Fig. 2) show the decoupling of Anatolia and the (Eastern) Pontides–Lesser Caucasus, as well as the large amount of deformation that is being accommodated along the major thrust faults of the Greater Caucasus and the Kura Basin (e.g. Jackson 1992; Forte et al. 2010, 2013; Mosar et al. 2010). The increase in GPS velocity from west to east across the Lesser Caucasus clearly shows the present-day counter-clockwise (CCW) rotation of the Lesser Caucasus with respect to fixed Eurasia (Reilinger et al. 2006; Kadirov et al. 2008; Djamour et al. 2011; Karakhanyan et al. 2013).

The basement units of the Eastern Pontides portion of the Sakarya Zone and the Transcaucasus are covered by unmetamorphosed Late Palaeozoic, Mesozoic and Cenozoic sequences (Fig. 1b; Okay & Şahintürk 1997; Yılmaz & Kandemir 2006; Yılmaz et al. 2000; Adamia et al. 2011). Widespread Jurassic to Late Cretaceous (Maastrichtian) volcanics are
interpreted as arc volcanics attributable to the northwards subduction of the Neo-Tethys (Tüysüz et al. 1995; Okay & Şahintürk 1997; Tüysüz 1999; Okay et al. 2006; Rice et al. 2006; Tüysüz & Tekin 2007; Adamia et al. 2011). Subduction resulted in high-pressure metamorphism between 170 and 90 Ma (Rolland et al. 2009a; Topuz et al. 2013). The metamorphic basement of the SAB (Fig. 1b) mainly consists of gneisses, mica schists and diorite–leucogranite intrusions (Aghamalyan 1998), which record a Late Jurassic to Early Cretaceous (160–120 Ma) metamorphic evolution in response to the south-dipping subduction of the Neo-Tethys below the SAB (Hässig et al. 2014b). The basement rocks are unconformably overlain by unmetamorphosed incomplete Palaeozoic to Mesozoic (Santonian) sedimentary sequences (Paffenholz 1959; Karyakin 1989; Sosson et al. 2010). These sedimentary sequences are overlain by Middle to Upper Jurassic ophiolitic sequences (Danelian et al. 2008, 2010, 2012; Galoyan et al. 2009; Rolland et al. 2009b; Asatryan et al. 2012; Hässig et al. 2013), which are, in turn, unconformably overlain by oce- anic island basalts and arc-type volcanic rocks (Galoyan et al. 2007, 2009; Galoyan 2008) of Early Cretaceous age (Belov et al. 1991; Rolland et al. 2011; Asatryan et al. 2012). Santonian and younger sedimentary, volcano-sedimentary and volcanic rocks cover the ophiolites, oceanic island basalts and arc-type volcanic rocks (Sosson et al. 2010).

Strike test

Orogenic bends are classified based on kinematics and the relative timing of curvature (Weil & Sussman 2004; Johnston et al. 2013). Orogenic belts are characterized by primary bends are those in which the curvature is an inherent physiographical feature present prior, during and throughout the for- mation of the orogen, such as an embayment. Oro- clines are map-view curves that developed in response to the bending or buckling of an existing orogenic belt about a vertical axis of rotation (Carey 1955). The strike test (or orocline test; Schwartz & Van der Voo 1983; Eldredge et al. 1985; Yonkee & Weil 2010) evaluates the relationship between variations in the regional structural trend and the orientations of given geological fabric elements (e.g. palaeomagnetic declinations, fractures, cleavage, veins or lineations). The strike test distinguishes between two end-members: (1) a primary arcuate shape of a mountain belt that yields no variation of declination with strike, resulting from the absence of vertical axis rotations; and (2) a secondary orocline, characterized by a unit slope relation between the declination and the structural grain. Any intermediate relation between both parameters is known as a progressive orocline and could indicate one of two scenarios. In the first sce- nario, a partial initial curvature of the fold–thrust belt existed, which was modified by a later tectonic event. Alternatively, in the second scenario, the sampled rocks acquired their magnetization during the process of orocline bending, due either to remagnetization or to their formation during orocline bending. The two end-members and the intermediate field are illustrated in declination v. strike plots of the fold axis diagram in Figure 3. Our new data and the previously published data presented in this study were assessed using the strike test.

This study

Sampling and experimental procedure

The goal of this study was to constrain the timing of rotation previously observed by Bazhenov & Burtman (2002) and Hisarlı (2011) in Upper Creta- ceous and Eocene rocks. We therefore attempted to sample Oligocene to Pliocene geological forma- tions. The mostly coarse, clastic Neogene rocks were, however, often weathered and/or unsuitable for palaeomagnetic sampling. In the Georgian and Armenian segments of the Eastern Pontides–Lesser Caucasus orocline, we collected a total of 734 ori- ented palaeomagnetic cores from 37 sites (Table 1) using a gasoline-powered motor drill. The sampling covered four areas along the curvature of the belt. In Georgia, 15 sites east and west of Tbilisi, six sites in the Akhaltsikhe area and seven sites close to Batumi were sampled. In Armenia, nine sites were sampled in or just north of the SAB (Table 1). In terms of age and lithology, our dataset included four sites in Upper Cretaceous–Palaeocene limestones and silty marls, 20 sites in Eocene silt- stones, turbidites, tuffs and (volcano-)clastic sedi- ments, five sites in Oligocene (carbonatic) clay/siltstones and eight sites in Miocene mud/silt/sandstones, limestones and tuffs (Table 1). Ages were assigned based on nannofossils, molluscs, ostracods, forams or a combination of these species (Sadradze 2015). Only the age of site BB was derived from the geological map. Core orientations were measured with a magnetic compass and corrected for c. 6° present-day declination. Whenever permitted by their length, the cores were cut into multiple specimens in the laboratory. Subsequently, specimens were: (1) thermally demagnetized in a magnetically shielded oven (n = 335); (2) alternating field (AF) demagnetized (n = 83); or (3) ther- mally demagnetized up to 150 or 190°C followed by AF demagnetization (n = 461). We refer to the last group (combined thermal and AF demagnetization) as AF demagnetized, as the characteristic rem- anent magnetization (ChRM) was mostly isolated.
Fig. 3. Figure illustrating two end-members (primary orogenic bend v. secondary orocline) and an intermediate field (progressive orocline). T1, T2 and T3 are arbitrary successive moments in time. $D$, declination; $D_R$, reference declination; $S$, strike; $S_R$, reference strike.
<table>
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<th>Rock type</th>
<th>Site lat.</th>
<th>Site long.</th>
<th>S</th>
<th>Age*</th>
<th>ΔAge</th>
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Site lat., Site latitude; Site long., Site longitude; S, regional strike trend at the sampling location, ∆Age, age error; Ncomp number of of demagnetization diagrams for which the great circle approach was used to determine the ChRM; N/N₄⁵, total number of determined (VGPs); DEC, declination; INC, inclination; ∆Dₓ and ∆Iₓ, error in declination (D) and inclination (I) determined following Butler (1992), tions; K, precision parameter determined from the mean VGP directions; A95, cone of confidence determined from the mean VGP criteria in the text.
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sampled cores; $N_{45}$, number of alternating field demagnetized specimens; $N_{th}$, number of thermally demagnetized specimens; $N_{gc}$, number of ChRM directions/remaining number of ChRM directions after application of a fixed 45° cut-off on the virtual geomagnetic poles $k$, estimate of the precision parameter determined from the ChRM directions; $\alpha_{95}$, cone of confidence determined from the ChRM directions; $\lambda$, calculated palaeolatitude. The numbers in the ‘Exclusion reason’ column correspond to the numbers of the reliability...
from the AF demagnetization steps. Samples were heated to 150–190°C to: (1) remove a viscous remanent magnetization component (often carrying a present-day field direction) commonly recorded in sediments; (2) demagnetize the goethite (Néel temperature = 120°C; Özdemir & Dunlop 1996) present in a significant number of samples and often carrying the aforementioned viscous remanent magnetization; and (3) remove possible stress in magnetite grains caused by surface oxidation at low temperatures (Van Velzen & Zijderveld 1995). On all sites (except for site BB), both AF and thermal demagnetization were carried out to test the efficiency of each treatment in isolating the ChRM and the reproducibility of the results (e.g. site BD; Fig. 4l–m). Natural remanent magnetization demagnetization is displayed using orthogonal vector diagrams (Zijderveld 1967). The sample ChRM directions were determined using principal components analysis (Kirschvink 1980) on five to seven successive demagnetization steps for most of the samples. For datasets with low quality demagnetization diagrams, however, this was not always possible. Samples carrying two magnetization components with overlapping unblocking fields ($H_{ubs}$) or temperatures ($T_{ubs}$) were analysed using the great circle approach (e.g. Fig. 4j; McFadden & McElhinny 1988). This method determines the direction that lies closest on the great circle to the mean from well-determined natural remanent magnetization directions.

The magnetic mineralogy of representative samples for each site was determined by the interpretation of thermomagnetic runs performed either in low field (0.38 mT) or in high field (150–300 mT) for one or two samples per site. In the

![Fig. 4. (a–o) Orthogonal vector diagrams (Zijderveld 1967) showing characteristic demagnetization diagrams for the sampled sites. Closed (open) circles indicate the projection on the horizontal (vertical) plane. Alternating field and thermal demagnetization steps are indicated. All diagrams are displayed after correction for bedding tilt. Examples of (1) specimens from the same sample that were alternating field and thermally demagnetized are included for comparison of both techniques for sample BD12 (l, m) as well as (2) a specimen (AA9A, j) that was interpreted using great circles (see text for explanation).](image)
former case, the susceptibility v. temperature runs 
\((K/T\) curves) were performed on an Agico KLY-
3S Kappabridge, equipped with a CS-L low-temper-
ature device and a CS-3 furnace apparatus at the 
University of Montpellier II (France). The pow-
dered samples were sieved into 0.125–0.8 mm 
grain size fractions. They were first heated from liq-
uid nitrogen (77 K) to room temperature and then 
heated and cooled in air and/or in argon up to suc-
cessively higher temperature steps \((c. 280, 350, 
420, 590, 700\textdegree C)\) to monitor possible chemi-
ical changes in the magnetic minerals. At the end of 
these cycles, a low-temperature run was repeated 
on the same powder. In the latter high-field case, 
powdered samples were placed in a modified hori-
zontal translation type Curie balance with a sensitiv-
ity of \(c. 5 \times 10^{-9}\) Am\(^2\) (Mullender et al. 1993). A 
35–55 mg mass of powdered rock sample was 
placed into a quartz glass holder and was held in 
place by quartz wool. At heating and cooling rates 
of \(10\textdegree C/\text{min, the temperatures were cycled up}
and down up to a maximum temperature of 700\textdegree C.

Site means and virtual geomagnetic poles (VGP) 
and their means were calculated from the ChRM 
directions. A 45\textdegree fixed cut-off was applied on the 
VGPs per site to remove outliers. The error in decli-
nation \((\Delta D)\) and inclination \((\Delta I)\) were calculated 
from the \(A_{55}\) value following Butler (1992). Because 
directional distributions become more elongated 
towards lower latitudes, this approach more realisti-
cally describes the associated errors (Creer et al. 
1959; Tauxe et al. 2008; Deenen et al. 2011). Where-
ever applicable, the reversal test of McFadden & 
McElhinny (1990) and the fold test of Tauxe & Wat-
son (1994) were carried out.

**Rock magnetic results**

The rock magnetic data show that the magnetic 
carriers are similar in the SAB and in the Georgian 
sites of the Transcaucasus, irrespective of rock 
type. The magnetic carriers can, however, show 
large within-site variations with either ferrimagnetic 
behaviour with a well-defined Curie temperature 
\((T_c, e.g. GE19.17)\) or only paramagnetic behaviour 
\((e.g. GE21.21)\) with no \(T_c\). In addition to these para-
magnetic curves, several types of thermomagnetic 
curve can be distinguished. The majority of the 
data are characterized by reversible (volcanic tuff; 
e.g. GE19.17) or near-reversible curves (turbidites, 
volcanoclastics; e.g. GE05 and BD11, respectively) 
with a single \(T_c\) between 520 and 550\textdegree C, charac-
teristic of unoxidized low-Ti titanomagnetite. Low-
temperature runs down to the LN2 temperature 
\((77 K)\) confirm the presence of an original magnetic 
mineral close to magnetite in composition with a 
Verwey transition around 100–110 K \((-173 \text{ to}
–163\textdegree C; e.g. GE05.10). Site GE23 (volcanoclastic 
turbidites) shows also almost no alteration up to 
500\textdegree C, pointing again to a low-Ti titanomagnetite 
as the original magnetic mineral, but here large trans-
formations occurred when the sample was heated 
to 700\textdegree C. It is probable that some titanomaghemite 
was initially present, which inverted to hematite, 
as shown by the irreversibility of the cooling curve. 
Other curves are characterized by reversible behav-
ior up to \(c. 350\textdegree C, but a large degree of irreversi-
bility above 400–450\textdegree C (e.g. GE07.07). These curves 
also show Curie temperatures of magnetite in the 
range 550–580\textdegree C. Magnetite was originally present 
in GE07 (turbiditic section of tuffs), as shown by 
the Verwey transition in the first low-temperature 
run, probably together with a titanomaghemite that 
inverted on heating above 330\textdegree C. Part of the magne-
tite was formed on heating in the siltstones of site 
GE13, as indicated by the higher susceptibility on 
cooling and the heating/cooling \(T_c\) of magnetite. 
In the samples from sites GE13 and AA, as well as 
from other sites, a mineral with a \(T_c\) around 300–
310\textdegree C appears on the cooling curve. This is possibly 
an iron sulphide created on cooling, after heating 
to 590 and 700\textdegree C, respectively. The white oolithic 
limestone of GE10 shows large irreversibility 
avove 400\textdegree C as a result of newly formed magnetite 
on heating. This newly formed magnetite possibly 
oxidizes to hematite when heated to 700\textdegree C, as sug-
gested by the low susceptibility of the cooling curve.

**Field tests**

The reversal test developed by McFadden & McEl-
hinny (1990) and its classification \((A, B, C, indeter-
minate)\) could be applied to the three sites that 
recorded normal and reverse polarities: BA, GE18 
and BG. The classifications are based on the critical 
angle \(\gamma\) and the angle \(\gamma\) between the means, which 
is equivalent to using the \(V_s\) statistical parameter of 
Watson (1983). The test is indeterminate for site 
BA. The resulting classifications C (for site GE18) 
and B (for site BG) indicate that the ChRM direc-
tions have been properly isolated.

The non-parametric fold test of Tauxe & Watson 
(1994) was carried out on site GE09 and the com-
bined sites GE07 and GE08 based on their proximity 
and similar age. The individual ChRM directions 
of site GE09 were corrected for a plunging fold axis. 
This fold test was positive at the 95% confidence 
level, as the maximum clustering of the dataset 
occurs between 81 and 109% unfolding. For sites 
GE07 and GE08 the fold test is also positive. Max-
imum clustering occurs between 87 and 104% of 
unfolding. The positive fold tests show that the 
recorded remanent magnetizations in sites GE07, 
GE08 and GE09 are of pre-tilting origin and give 
confidence in the stability of the palaeomagnetic 
signal on geological timescales.
**Palaeomagnetic results**

Overall, the retrieval of reliable magnetic records from the Oligocene and Miocene silt/sandstones and mudstones was more difficult than from the Upper Cretaceous, Paleocene and Eocene volcanoclastics and turbidites. This might explain why previous studies concentrated on Upper Cretaceous and Eocene rocks. The site mean directions of the isolated ChRM and secondary magnetization components are listed in Table 1. Among our 37 sites, several give questionable results, which were analysed using a set of reliability criteria. The same reliability criteria were also applied to previously published data.

The calculated palaeolatitudes derived from the palaeomagnetic directions were plotted v. age (Fig. 5). A comparison of the calculated palaeolatitudes with the latitude v. age curves of Eurasia and Africa (Torsvik *et al.* 2012) showed that the great majority of the accepted sites displayed unrealistically low palaeolatitudes. Possible explanations for the low latitudes are discussed in this paper.

**Literature review procedure**

The data reviewed in this study consist of all the available sites from the GPMDB in addition to studies published in the English language scientific literature. Data from c. 100 sites with Upper Cretaceous to Pleistocene deposits were included in the review. The GPMDB contains results included in later studies that were also entered into the database. To avoid including the same dataset twice, we only included the combined and non-superseded studies in our review. Because we aimed to compare the data and associated errors obtained in this study with those of the database and in the literature, we recalculated the reported $\alpha_{95}$ values in the literature.

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**Fig. 5.** Palaeolatitude v. age diagram for the period 110–0 Ma. The Eurasian and African curves (calculated for a reference location in Georgia at $\lambda = 41.3$, $\varphi = 43.7$) with their $\Delta l_e$ error (shaded area) are derived from the global apparent polar wander path of Torsvik *et al.* (2012). Black (grey) circles and their error bars show the palaeolatitude and its error (calculated from $\Delta l_e$) for each accepted (rejected) site of this study. Blue triangles/red inverted triangles/purple diamonds and their error bars show the palaeolatitude and its error (calculated from $\Delta l_e$) for each accepted sediment/volcanic/intrusive site of the review data presented in this study. The dotted lines indicate the position of the Eurasian curve calculated with flattening factors ($f$) of 0.9, 0.6 and 0.3 to show the possible effect of inclination shallowing. Yellow dotted lines subdivide epochs. E. Cret., Early Cretaceous; Paleoc., Paleocene; Pl., Pleistocene; Pl., Pliocene; Pl., Pleistocene.
to $A_{95}$ values using the Creer transformation (Creer 1962). $\Delta D_x$ and $\Delta I_x$ were subsequently calculated from the $A_{95}$ values (following Butler 1992).

**Reliability criteria**

A set of reliability criteria was applied to our data as well as to the literature data presented in this study. However, not all the reliability criteria were applied to the GPMDB data because the available information (e.g. statistics, demagnetization method and tectonic context) was not always sufficient to apply all the reliability criteria. We therefore cannot guarantee that all the GPMDB datasets of low quality (in terms of statistics, demagnetization method and tectonic context) were excluded after applying the reliability criteria. The reliability of this review therefore does not lie in the individual datasets, but in the consistency within the large dataset, i.e. in the spatial and temporal variation of the vertical axis rotations. The reliability criteria listed here were applied to all the data and led to the exclusion of a number of sites (see Table 1).

Sites that did not pass the following criteria were excluded from further analysis.

1. Sites with less than five samples ($N$; e.g. sites GE26 and BB) for sediments, or $N<5$ for the number of lava flows sampled (e.g. dataset 158).
2. Sites with a mean ChRM direction (before tilt correction) that was statistically indistinguishable from the present-day geocentric axial dipole field at the site location ($I = c. 60^\circ$; $D = c. 0^\circ$). This led to the exclusion of sites GE06, GE14, GE24, GE26, GE28 and BB. Site GE10 was not excluded as a result of the presence of one reversed polarity sample, which gives some confidence that the ChRM could be of primary origin.
3. Datasets with suspect directions (e.g. north/up or south/down directions), either indicating $>90^\circ$ rotation or deposition in the southern hemisphere (e.g. sites GE06, GE25 and dataset 166).
4. Datasets that yield an inclination that would have undergone very significant inclination shallowing (flattening factor, $f < 0.3$) or with unrealistically high inclinations ($\tan I_{\text{observed}} = f \tan I_{\text{expected}}$, King 1955). The sampled sediments were probably deposited near or on the Eurasian margin and the dotted lines in Figure 5 indicate the position of the Eurasian margin, for the study area, with three different flattening factors ($f = 0.9$, $f = 0.6$ and $f = 0.3$). This criterion led to the rejection of sites GE14 and GE22.
5. Datasets with an error in declination ($\Delta D_x$) $>15^\circ$ (e.g. sites GE20, BA and dataset 13).
6. Finally, we removed obvious outliers from the general data trend (see Figs 5 & 6). In the review, these are sites 166 (which was also excluded under criterion 3), 167 and 96 from the Late Cretaceous time slice, as well as site 78 from the Paleocene. These sites have probably been affected by local fault...
movements or were not recognized as remagnetized sites. The same holds for our site GE20 (sampled in Upper Eocene rocks), which displays a large rotation that is the opposite of all other reported rotations in the area.

The data from the literature review yielded very low inclinations, especially for the post-Middle Eocene time interval (c. 45 Ma and younger; Fig. 5). Possible reasons for the low latitudes are discussed in the following section.

**Accepted sites after application of the reliability criteria**

*This study*

Nine of the 37 sites were excluded from further analyses based on the listed reliability criteria. The criteria that led to their exclusion are indicated for each site in Table 1. Site GE02 was rejected because no result could be retrieved from the 26 demagnetized specimens.

The finite rotation vectors relative to stable Eurasia are plotted in Figure 6 for the 26 accepted sites presented in this study. The rotation vector \( \mathbf{R} \) was calculated by subtracting the expected declination for stable Eurasia from the observed declination \( R = D_{\text{obs}} - D_{\text{exp}} \), Butler (1992). The expected declinations have been calculated from the global apparent polar wander path (Torsvik et al. 2012) in the Eurasian reference frame at our reference location (\( \lambda = 41.3^\circ \text{N}; \varphi = 43.7^\circ \text{E} \)). The following corrections \( D_{\text{exp}} \) were applied: +10° for the Late Cretaceous and Paleocene; +11° for the Eocene; +8.5° for the Oligocene; and +4° for the Miocene. Overall the rotations vary from anticlockwise in the western part to slightly clockwise in the eastern part of the orocline, in concordance with earlier findings (Bazhenov & Burtman 2002; Hisarli 2011). The limited geographical spread of the data displayed per time interval does not allow an immediate assessment of the variation in the amount of rotation along the orocline. Our sites sampled from Upper Cretaceous and Paleocene rocks were mostly limited to the eastern part of the study area. The Miocene has only been sampled in the northern part of the orocline because there were no outcrops suitable for sampling in Armenia.

**Literature review**

The application of the reliability criteria led to the exclusion of 12 sites. Using the same procedure as described in the preceding sections, we calculated the finite rotations for the remaining 88 datasets (following Butler 1992). The number of data in each time slice allowed us to conclude that the rotations not only change from westerly to easterly (from west to east), but also that the amount of rotation is lower in younger than in older rocks. The declinations in the Oligocene, Miocene and Plio-Pleistocene intervals are visibly lower than in the Late Cretaceous and Eocene.

The general trend since the Late Cretaceous therefore shows a decrease in the rotation with time.

**Strike test results: a quantitative approach**

For the strike tests, we determined a regional strike for each site based on satellite images from the sampling area in combination with geological maps. The regional strikes were assigned a fixed 10° uncertainty. The strike tests in Figure 8 are based on linear statistics and the 95% error bands associated with the slopes \( m \) of the regression lines were calculated following Tukey’s range test. We performed the strike test for four time intervals based on the available number of datasets per time interval and the tectonic chronology. The first interval (Late Cretaceous–Paleocene) includes the rocks formed presumably before and during the SAB–Eurasia collision. The datasets from rocks deposited in this time interval \( (\Delta t) \) cover c. 40 myr. The second interval covers the Eocene, which represents the time span after initial SAB collision and before the initiation of Arabia–Eurasia collision. Because we have only one dataset from Lower Eocene rocks, all but one dataset in this time interval cover a period of c. 12 myr in the Middle and Late Eocene. The third, Oligocene to Pleistocene, interval includes the Arabia–Eurasia collision \( (\Delta t \approx 33 \text{ myr}) \). The limited number of available datasets for the Oligocene did not allow us to perform the orocline test for these sites only. We do, however,
present a separate oroclone test for data from Late Miocene to Pleistocene rocks. The time interval for this test is therefore limited to c. 12 myr.

The Late Cretaceous to Paleocene strike test (Fig. 8a) gave a slope \( m = 0.604 \pm 0.255 \) (correlation coefficient, \( R^2 = 0.395 \)). The result shows that (within 95% confidence limits) some pre-Late Cretaceous curvature must have existed because \( m < 1 \). We performed two different tests for the Eocene period (Fig. 8b): one with all accepted data (\( m = 0.324 \pm 0.147, R^2 = 0.328 \)) and a second test that excluded five points (\( m = 0.479 \pm 0.133, R^2 = 0.608 \)) that were obvious outliers (Figs 6 & 7c). The latter test resulted in a better fit and much higher \( R^2 \) value. For both Eocene tests, the slope \( (m) \) is lower than for the Late Cretaceous to Paleocene strike test. Both results are, however, not statistically distinguishable within the 95% confidence limits from the Late Cretaceous to Paleocene strike test. The Oligocene to Pleistocene strike test (Fig. 8c) gave a slope \( m = -0.092 \pm 0.163, R^2 = 0.032 \) results in a slope that is indistinguishable from \( m = 0 \). We note, however, that the number of datasets from Oligocene rocks is limited and the Oligocene to Pleistocene strike test covers a very long time interval (c. 33 myr). We therefore performed a strike test on all the data from Upper Miocene to Pleistocene rocks (Fig. 8d), which represents only a c. 12 myr time slice. The Late Miocene to Pleistocene strike test results in \( m = -0.078 \pm 0.187 \) and \( R^2 = 0.023 \).

**Discussion**

**Oroclinal bending**

The results from the strike tests indicate that the Eastern Pontides–Lesser Caucasus oroclone falls within the field of progressive oroclines (Fig. 3; Weil & Sussman 2004; Johnston et al. 2013). Some pre-existing Late Cretaceous curvature (c. 40 ± 25%) must have been present because the slope of the Late Cretaceous to Paleocene strike test is significantly less than unity. The slopes of the two strike tests performed on data from Eocene rocks are slightly less steep than for the Late Cretaceous to Paleocene, indicating that oroclone formation probably initiated before the Eocene as a result of SAB–Eurasia collision. The results are, however, not statistically distinguishable for the two time intervals. Our preferred Eocene strike test, which excludes five outliers, indicates that c. 48 ± 13% of oroclone formation occurred after the Eocene. The Oligocene to Pleistocene strike test, with a slope that was statistically indistinguishable from zero, seems to indicate that oroclone formation occurred entirely before the Oligocene. However, the number of data entries for the Oligocene is limited (\( N = 8 \)), the data are very scattered and the sites are not homogenously distributed over the study area (Figs 6 & 7d and open circles in Fig. 8c). Therefore we conclude that the data from Oligocene rocks do not allow us to constrain the timing of oroclone formation. The strike test for the Late Miocene to Pleistocene interval (\( \Delta t \) c. 12 myr) results in a slope that is statistically indistinguishable from zero. We note, however, that the regression line is poorly defined (Fig. 8d). We thus conclude that oroclone formation probably ended before the Late Miocene. A study of pre-Late Cretaceous rocks is necessary to determine whether the pre-existing Late Cretaceous curvature resulted from oroclinal bending or whether this curvature is an inherited physiographical feature (i.e. a primary orogenic bend; see Fig. 3).

Despite some limitations in the number of datasets – per time interval, as well as in terms of geographical spread – a coherent chronological trend emerges from the strike test (Fig. 8). We argue that about 40% of the curvature of the Eastern Pontides–Lesser Caucasus fold–thrust belt was acquired before the Late Cretaceous and c. 60% between the Paleocene and probably the Late Miocene. Even though the Late Cretaceous–Paleocene and Eocene strike tests (Figs 8a, b) result in slopes that are statistically indistinguishable from each other, these two tests suggest that c. 10% of the curvature was acquired after the Paleocene and before the Middle Miocene, most probably as a consequence of SAB–Eurasia collision. The remaining c. 50% (Fig. 8b, excluding outliers) occurred after the Late Eocene and probably before the Late Miocene (Fig. 8d) and therefore includes the time interval of Arabia–Eurasia collision. Consequently, we conclude that c. 50% of the belt’s curvature developed in response to Arabia–Eurasia collision. The interpretation of the strike tests of the present study is slightly different from that given by Bazhenov & Burtman (2002) and Hisarli (2011). All three studies agree that the curvature of the belt developed mostly after the Late Cretaceous to Eocene, but Bazhenov & Burtman (2002) and Hisarli (2011) did not conclude that the belt had an existing curvature before the Late Cretaceous to Paleocene. The slopes of the regression lines presented by Hisarli (2011) after the removal of several outliers are, however, slightly lower than unity and lower for the Eocene than for the Late Cretaceous, suggesting progressive oroclinal bending as in our study. We stress that the main difference between this study and the two earlier studies is that we performed a more rigorous strike test on a much larger number of data (including the data from both earlier publications).

After the Eocene, the ongoing convergence and collision between Arabia and Eurasia led to the
exhumation of the western Greater Caucasus (Vincent et al. 2007) – which has rapidly accelerated since the earliest Pliocene in the central Greater Caucasus (Avdeev & Niemi 2011) – and exhumation in the Bitlis–Zagros thrust zone in the Early to Middle Miocene (Okay et al. 2010). Over

Fig. 7. Shaded relief digital elevation models of the Caucasus region showing all rotations (calculated following Butler 1992; see text for explanation) with their associated $\Delta D_0$ error for the accepted (white) and rejected (grey) sites from the presented review for the (a) Late Cretaceous, (b) Paleocene, (c) Eocene, (d) Oligocene, (e) Miocene and (f) Plio-Pleistocene. Numbers refer to the numbers that were assigned to each study in Supplementary Data Table 2.
the last few million years, the region has been characterized by localized deformation, as evidenced by the significant amount of shortening in a narrow zone between the Greater and Lesser Caucasus (Forte et al. 2013), as well as by Anatolian escape tectonics since the Late Miocene (Dewey & Şengör 1979; Şengör et al. 1985). Oroclinal bending most probably ended before the Late Miocene and therefore potentially predates the transformation of distributed to more localized deformation in the region. The entire Lesser Caucasus region continued to experience deformation.
after the Late Miocene, as shown by Copley and Jackson (2006), who investigated how active faulting accommodates Arabia–Eurasia collision in the Turkish–Iranian Plateau. Copley and Jackson (2006) proposed a model of CCW rotating blocks bounded by right-lateral strike-slip faults in the central part of the Lesser Caucasus. This is consistent with the overall CCW rotation of the Lesser Caucasus region as shown by geodetic displacement vectors (Reilinger et al. 2006; Kadirov et al. 2008; Djamour et al. 2011; Karakhanyan et al. 2013).

Fig. 7. Continued.
The timing of orocline formation in the Eastern Pontides–Lesser Caucasus is not coeval with orocline formation in the Central Pontides and Central Alborz Mountains. In the Central Pontides, orocline formation occurred in the latest Cretaceous or Paleocene in response to collision of the Central Anatolian Crystalline Complex with Eurasia (Fig. 1a; Meijers et al. 2010; Lefebvre et al. 2013) and had ended by the Eocene. It therefore predated orocline formation in the Lesser Caucasus region. In contrast, oroclinal bending in the Central Alborz Mountains occurred after c. 7.6 Ma (Late Miocene) and had ended by the Pleistocene. Although orocline bending in the Central Alborz Mountains–Lesser Caucasus belt. Although orocline bending in the Eastern Pontides–Lesser Caucasus seems to have been a process that went on for more than 70 myr, orocline bending in the Central Alborz Mountains and the Central Pontides was of much shorter duration.

The results of the strike tests are highly influenced by the assigned regional strike for each individual site, based on satellite images in combination with geological maps. Therefore we emphasize that detailed structural mapping of the area surrounding each site would probably improve the accuracy of our strike tests and allow a better comparison between our results and the proposed tectonic models for the region. Detailed structural mapping around all our sampling locations, as well as around the sites from the literature review, was, however, beyond the scope of this study.

**Low palaeolatitudes**

All the sampled rocks from this study and the literature data were deposited on the Eurasian margin or close to the Eurasian margin (the sites sampled in the SAB; Meijers et al. 2015). All the presented datasets should therefore plot within error above the African or below the Eurasian palaeolatitude v. age curve of Figure 5. The estimated shortening within the fold–thrust belt and unbending of the Lesser Caucasus orocline must be taken into account, however, which was estimated to be...
>400 km by Bazhenov & Burtman (2002). The correction for this shortening, which occurred mostly in a north–south direction, positions the Eurasian margin c. 4° further south (given that 110 km corresponds to c. 1° in latitude) and thus reduces the palaeolatitudinal gap. The palaeolatitudes in the compilation in this study are often more than 4° lower than the calculated palaeolatitudes from the apparent polar wander path (Fig. 5) and can therefore only partially be explained by shortening in the Caucasus region between stable Eurasia and Arabia. We therefore explore the possibility of other processes that could result in low inclinations.

The calculated palaeolatitudes (Fig. 5) underscore three main points.

1. The data distribution is heterogeneous (e.g. there are a sizeable number of datasets for the Middle to Late Eocene, but not for the Early Eocene, very few data for the Oligocene and almost no data for the Early and Middle Miocene).
2. The palaeolatitude scatter for some time intervals is considerable (e.g. for the Middle and Late Eocene).
3. The palaeolatitudes are often much lower than expected for the Middle to Late Eocene and the Late Miocene to Pleistocene.

The only time interval for which nearly all results display the expected palaeolatitudes is the Upper Cretaceous (c. 100–60 Ma). Many datasets in the 100–60 Ma interval plot between the African and Eurasian curve and are statistically indistinguishable from the Eurasian curve.

One of the possible causes for low palaeolatitudes is a bias in the recording of the remanence by the rock itself, in particular inclination shallowing, which only affects sedimentary rocks. Inclination shallowing is a (post-)depositional process in sedimentary rocks that can lead to an underestimation of the palaeolatitude. Typical flattening factors (f) range from 0.4 to 0.83 for hematite-bearing rocks and from 0.54 to 0.79 for magnetite-bearing rocks (Bilardello & Kodama 2010). To test for inclination shallowing, we subdivided all the data into three groups: datasets from: (1) sedimentary rocks (64 accepted sites; including all sites from this study); (2) extrusive volcanic rocks (19 accepted sites); and (3) intrusive rocks (five accepted sites; Fig. 5). Whenever a site consisted of a combination of volcanic and sedimentary rocks it was labelled as sedimentary rock. We distinguished between extrusive volcanic and intrusive rocks because, even though intrusive bodies are unlikely to experience significant post-emplacement tilting, they cannot be corrected for tilting. The subdivision of the data in the three groups shows that the volcanic rocks (in red, Fig. 5) yielded expected inclinations for Eocene and older rocks, unlike the sedimentary rocks that mostly displayed palaeolatitudes that were too low from c. 60 Ma to recent. A portion of the Pliocene to Pleistocene magmatic (extrusive volcanics + intrusives) rocks displays low as well as relatively high palaeolatitudes.

Inclination shallowing can thus be a possible explanation for the low latitudes of sedimentary rocks younger than c. 60 Ma, but not for the low (N = 5) and high (N = 1) palaeolatitudes observed in Pliocene to Pleistocene recent magmatic rocks.

We note that the lowest palaeolatitudes in these magmatic rocks come from old datasets and are possibly caused by an error in the determination of the ChRM direction. We must keep in mind, however, that often no information is provided on the demagnetization procedure or on the bedding tilt corrections performed on the sites from the literature data, even though the datasets passed our reliability criteria. In addition, insufficient averaging of secular variation for the intrusive and volcanic rocks and unrecognized remagnetizations (for the literature data) could be a cause for palaeolatitudes that deviate from the Eurasian curve. Not averaging the secular variation of the Earth's magnetic field can lead to a direction that represents an instantaneous field direction instead of that of the geocentric axial dipole hypothesis. Palaeolatitude errors related to the latter possibility would not be biased towards low palaeolatitudes and could therefore explain the spread between too low as well as too high palaeolatitudes in the data from the magmatic rocks for the Pliocene to Pleistocene time interval.

A last possible explanation for the low palaeolatitudes is the contribution of a significant octupolar field to the mostly dipolar Earth’s magnetic field. This hypothesis has been proposed by several researchers for the inclination shallowing observed in Cenozoic rocks from Central Asia (Si & Van der Voo 2001; Dupont-Nivet et al. 2010), although it has been disproved by others (Camps & Prévot 1996; Camps et al. 2007). The effect of an octupolar contribution to the Earth’s magnetic field that would influence the inclination in the Lesser Caucasus region would, however, affect all datasets similarly, disregarding the sampled rock types. The large scatter of palaeolatitudes between c. 45 and c. 30 Ma rules out this hypothesis; differential shallowing of the inclination (depending on the lithology of the sampled rocks) is a more viable explanation for the spread in palaeolatitudes.

A correction for c. 4° of shortening within the Caucasus region (for the Eocene and, to a lesser degree, for younger rocks), variable amounts of inclination shallowing in the sedimentary rocks and insufficient averaging of secular variation in the Pliocene to Pleistocene volcanic rocks can explain
the majority of the low inclination data of Eocene and younger rocks in Figure 5. Inclination shallow- ing would, however, not or hardly affect the declination, whereas insufficient averaging of secular variation would result in erroneous declinations.

Conclusions

Our newly presented palaeomagnetic data, supplemented with published data from Upper Cretaceous to Pleistocene rocks, allow us to confirm that the Eastern Pontides—Lesser Caucasus fold–thrust belt is a progressive orocline. To date oroclinal bending and relate it to tectonic events, a strike test was performed on four intervals that were sub-divided based on tectonic chronology: (1) the Late Cretaceous—Paleocene, a period that includes the rocks formed during SAB–Eurasia collision; (2) the Eocene, representing the time span between the Late Eocene and before the Late Miocene, most probably as a consequence of Arabia–Eurasia collision; (3) the Oligocene to Pleistocene, a time span that includes Arabia–Eurasia collision; and (4) the Late Miocene to Pleis- tocene. The strike test shows that c. 40% of the arcuate shape was acquired before the Late Cretaceous and c. 60% after the Late Cretaceous to Paleocene time period. The pre-Late Cretaceous curvature could have been a pre-existing physiographical feature or formed by oroclinal bending due to an earlier tectonic event. Possibly c. 10% of the curvature developed between the Paleocene and the Mid- dle Eocene as a result of SAB–Eurasia collision. The remaining c. 50% of curvature occurred after the Late Eocene and before the Late Miocene, most probably as a consequence of Arabia–Eurasia collision. The palaeolatitudes calculated from the palaeo- magnetic data are generally low in the c. 45 Ma to present time interval. These low palaeolatitudes are probably caused by a combination of factors: (1) shortening within the fold–thrust belt north of the orocline; (2) inclination shallowing (for sedi- ments only); and/or (3) an under-sampling of palaeo-secular variation in Pliocene to Pleistocene volcanic rocks from older studies.

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Lesser Caucasus Orocline Formation

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