Pliocene Growth of the Dowlatabad Syncline in Frontal Fars: Folding Propagation across the Zagros Fold Belt

**Mahdi Najafi^{1*}, Elisabet Beamud², Jonas Ruh³, Frédéric Mouthereau⁴, Alireza Tahmasbi⁵,
Gilen Bernaola⁶, Ali Yassaghi⁷, Hossein Motamedi⁵, Shahram Sherkati⁸, Mohammad Ghasem
Hassan Goodarzi⁵, Jaume Vergés⁹**

¹*Department of Earth Sciences, Institute for Advanced Studies in Basic Sciences (IASBS), Zanjan,
Iran*

²*Paleomagnetic Laboratory CCI-TUB-Geo3Bcn, CSIC, Lluís Solé i Sabarís, s/n, 08028 Barcelona,
Spain*

³*Structural Geology and Tectonics Group, Geological Institute, Department of Earth Sciences,
ETH Zurich, Sonneggstrasse 5, 8092 Zurich, Switzerland*

⁴*Geosciences Environnement Toulouse (GET), Université de Toulouse, UPS, Univ. Paul Sabatier,
CNRS, IRD, 14 av. Edouard Belin, F-31400 Toulouse, France*

⁵*National Iranian Oil Company, Exploration Directorate, PO Box 19395-6669, Tehran, Iran*

⁶*Department of Mining-Metallurgy Engineering and Material Science, Faculty of Engineering in
Bilbao, University of the Basque Country (UPV/EHU), Plaza Ingeniero Torres Quevedo 1, Bilbao,
Spain*

⁷*Department of Geology, Tarbiat Modares University, Tehran, Iran*

⁸*PETRONAS, Kuala Lumpur, Malaysia*

⁹*Group of Dynamics of the Lithosphere (GDL), Geosciences Barcelona, Geo3Bcn - CSIC, Lluís
Solé i Sabarís s/n, 08028 Barcelona, Spain*

**Correspondence: mahdinajafi@iasbs.ac.ir*

ABSTRACT

The integration of biostratigraphy, strontium isotope stratigraphy and magnetostratigraphy allowed to precisely date the >3.0-km-thick marine to non-marine foreland sedimentary succession within the Dowlatabad growth syncline along the Frontal Fars in the Zagros Fold Belt. This area was the missing link to complete the ages of syntectonic deposits in the Fars arc to be able to carry out regional quantified analyses of both the migration of sedimentary belts and of the propagation of folding across the Mesopotamian foreland basin. Both are essential for defining the interplay of basin evolution and sequence of folding.

Deposition of the foreland marine marls in the Mishan Fm. started about 11.5 Ma. The transition to a non-marine basin infill occurred at 4.9 Ma by the progradation of thick fluvial deposits of the Aghajari Fm. with a fast accumulation rate of 63 cm/kyr. The beginning of growth strata deposition and thus the onset of folding in the Dowlatabad syncline is dated at 4.65 Ma. The first appearance of carbonate conglomerates sourced from the Guri limestone at 2.8 Ma marked the progressive dismantling of the nearby growing anticlines. Tectonic deformation in the front of the Fars arc was active at least for 2.85 Myr and ceased at 1.8 Ma beneath the Bakhtyari unconformable conglomerates composed of external clasts derived from the Zagros hinterland.

The compilation of magnetostratigraphic ages reveals that both the migration of Aghajari-Bakhtyari sedimentary belts and propagation of folding front was in-sequence towards the foreland with rates close to 20 mm/yr in the Fars arc and to 15 mm/yr in the Lurestan arc in the last 20 Myr. These high rates of folding propagation are about one order of magnitude larger than age equivalent shortening rates (~4 mm/yr in Fars arc, and ~2 mm/yr in Lurestan arc) and thus implying an efficient detachment level at the base of deformed Arabian sedimentary cover.

Numerical experiments including cover and basement are designed to test the influence of inherited basement structures on the deformation propagation within the cover sequence providing clues on

the partly coeval in-sequence deformation of the Zagros Simply Folded Belt and the local out-of-sequence Mountain Frontal Fault system as illustrated by regional and local geology.

KEY WORDS: magnetostratigraphy, fold growth, folding propagation, numerical modelling, Zagros Fold Belt, Mesopotamian foreland basin

INTRODUCTION

The Zagros Mountains form a large segment of the Alpine–Himalayan orogenic system between Arabia and Eurasia evolving during a long-lasting and multi-episodic orogeny by a combination of thin-skinned folding and basement-involved thrusting (Fig. 1). The first episode of orogenic deformation was the Late Cretaceous emplacement of the Neo-Tethys ophiolites on top of the Arabian passive margin, interpreted as a volcanic arc collision and resulted in a restricted early foreland flexure in front of the emplaced ophiolite slices (Ricou, 1971; Agard et al., 2005; Saura et al., 2011). The continental collision between Arabia and Eurasia initiated by a soft collision during the Oligocene, characterized by underthrusting of the distal portion of the continental margin (Ballato et al., 2011; Mouthereau et al., 2012; Madanipour et al., 2017). The final closure of the Neo-Tethys occurred as a hard collision phase characterized by increasing crustal thickening since the early Miocene during which deformation propagated towards the craton interior to build the Zagros Mountains and their coupled foreland basin (Hessami et al., 2001; Mouthereau et al., 2007, 2012; Ruh et al., 2014; Pirouz et al., 2015; Vergés et al., 2019; Etemad-Saeed et al., 2020).

During the Arabia-Eurasia collision, the tectonic evolution of the Zagros Fold Belt was shaped by the interplay of thin-skinned detachment folding of the Phanerozoic sedimentary cover (McQuarrie, 2004; Sherkati et al., 2006; Jahani et al., 2007; Mouthereau et al., 2007; Vergés et al., 2011a), as well as by a significant contribution of reactivated faults in the crystalline Precambrian basement, inherited from Permo-Triassic Neo-Tethys rifting (Figs. 1B and 2) (Jackson et al., 1981;

Berberian, 1995; Talebian and Jackson, 2004; Mouthereau et al., 2006, 2007; Karasözen et al., 2019).

The Zagros Mountain Front has a sinusoidal shape, forming salients and recesses including two well-developed arc-shaped folded domains, the Lurestan arc in NW Zagros and the Fars arc in SE Zagros (e.g., Falcon, 1961; Berberian, 1995; McQuarrie, 2004; Sepehr and Cosgrove, 2004; Emami et al., 2010; Tavani et al., 2020). Presently, there is general consensus on the forelandward sequence of folding across both Lurestan and Fars arcs in particular and in the Zagros Fold Belt in general based on the extensive database on timing of deformation (see discussion in Vergés et al., 2019). These data include Strontium isotopic ages (Ehrenberg et al., 2007; Pirouz et al., 2015, 2017) and a significant number of magnetostratigraphic studies (Homke et al., 2004; Khadivi et al., 2010; Ruh et al., 2014; Pirouz et al., 2017; Vergés et al., 2019; Lashgari et al., 2020). They have been particularly useful to determine the stratigraphy of the Mesopotamian foreland basin sedimentary fill and the timing of onset and duration of evolution of growth synclines and coeval adjacent anticlines (Fig. 1B). The folding age of the outermost domain of the Simply Folded Belt is however not well constrained. To date, only the Changleh syncline along the front of the Lurestan arc, has yielded constraints on the age of folding at the mountain front at 8.1–7.2 Ma (Fig. 1B) (Homke et al., 2004). Because this fold developed above the blind basement thrust of the Mountain Frontal Fault, it is still unclear whether the age determined corresponds to thin-skinned folding or deep-seated basement deformation. Cover folding and basement thrusting were partially coeval as documented in Emami et al. (2010), Saura et al. (2015), and Le Garzic et al. (2019) on a large scale although the Mountain Frontal Fault in Lurestan ends up deforming the folds located in its hangingwall as discussed in Tavani et al. (2020).

Hence, to obtain age constraints on the outermost folding in the Zagros Fold Belt, we analysed the Dowlatabad syncline in the front of the Fars arc, near the Persian Gulf, which exhibits growth patterns and is situated far away from any step formed by blind basement faults. The growth

syncline comprises, in its SW flank, a continuous and well-preserved >3.0-km-thick foreland sedimentary succession that begins with the shallow marine carbonates and marls of the Guri Mb. and Mishan Fm., sampled for calcareous nannofossils and $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic dating, and continues with the typical fluvial and alluvial continental deposits of the Aghajari and Bakhtyari formations (2.2-km-thick log), sampled for magnetostratigraphic analysis. Furthermore, the first appearance of Guri monomictic clasts as well as polymictic external clasts in the syncline and thus local and regional tectonics and drainage evolution is interpreted and dated. To examine both the timing of folding and of migration of sedimentary belts across the Zagros foreland, these ages were combined with a regional structural cross-section extending from the Main Zagros Thrust to the SW of the Persian Gulf, and with a compilation analysis of published magnetostratigraphic studies. In addition, numerical experiments of basement-involved fold-and-thrust belts were conducted to test the effect of inherited basement faults on the propagation of folding in the sedimentary cover. Numerical results are compared to regional patterns of deformation front propagation based on magnetostratigraphic results presented here.

Figure 1

STRUCTURE OF THE SE ZAGROS FOLD BELT

The Zagros foreland Fold Belt extends along 2000 km from eastern Turkey to southern Iran, and is divided into different tectonic domains based on its structural and stratigraphic variations (Berberian, 1995; Sepehr and Cosgrove, 2004) (Fig. 1B). The arc-shaped Fars province in the SE Zagros is the largest tectonic domain of the Zagros Fold Belt, with more than 280 km of width (Fig. 1B). Several studies presented geological cross-sections from the Fars province (McQuarrie, 2004; Sherkati et al., 2006; Alavi, 2007; Mouthereau et al., 2007; Oveisi et al., 2009; Yamato et al., 2011; Najafi et al., 2014), although they do not present a consensual interpretation of structural geometries at depth or of mechanisms of folding. In order to better define the structure of SE

Zagros, we integrated field data with new seismic lines and well data to construct a revised structural cross-section. The section extends from the Main Zagros Thrust (MZT) to the south of the Persian Gulf in Saudi Arabia. It is oriented NE-SW, with a length of ~540 km and extending downwards to Moho depths (Fig. 2). The Zagros structural features cropping out along the cross section are described from hinterland to foreland.

The Main Zagros Thrust is the SW boundary of the Sanandaj-Sirjan metamorphic-magmatic belt and is characterized by the exposure of ophiolite and radiolarite oceanic remnants, partially over-thrusted as tectonic nappes on top of the High Zagros Imbricated Zone (Fig. 2). The High Zagros Imbricated Zone constitutes highly deformed imbricated and duplex thrust systems (Tavakoli-Shirazi et al., 2013; Sadeghi et al., 2016).

The Fars arc is characterized by remarkably regular and long NW-trending concentric folds with homogenous wavelengths of about 15 km (McQuarrie, 2004; Sherkati et al., 2006; Mouthereau et al., 2007; Fig. 2). This Simply Folded domain is limited by the Imbricated Zone to the north, and by the Persian Gulf to the south. Folds in the Fars arc involve a 10–12-km-thick pile of Phanerozoic sedimentary cover. At the base of the succession, ~2 km salt-bearing evaporites of the Ediacaran–Early Cambrian Hormuz series decouple the deformation from the Precambrian crystalline basement and detachment folding is the main deformation mechanism of the sedimentary cover in the Fars arc (Fig. 2).

In addition to the thin-skinned detachment folding, the contribution of inherited basement reverse faults in the folding of Fars arc is well-constrained for the High Zagros Fault, Surmeh Fault, and Mountain Front Fault (Figs. 1B and 2), based on their seismicity (Jackson and Fitch, 1981; Berberian, 1995; Talebian and Jackson, 2004; Karasözen et al., 2019), structural relief (Homke et al., 2010; Najafi et al., 2018) and magnetic anomalies (Teknik and Ghods, 2017). The Mountain Front Fault is also inferred from the trace of the Mountain front flexure, which is strikingly evident

in the Kirkuk, Lurestan, and Dezful domains (Fig. 1B). In the Fars arc, however, the frontal flexure is less evident since structural steps are smaller, and therefore there is not a full agreement in the exact location of such faults at depth (e.g., Jackson et al., 1981; Berberian, 1995; Mouthereau et al., 2007).

The total shortening across the Fars arc is about 65 km (21%), calculated by comparing the present length of layers (~241 km) between the pin and loose lines in the A–B cross-section, and their pre-folding original length once restored to flat (~306 km). The pin line was located at the deformation front near the Persian Gulf coastal line and the loose line was located in the footwall of the High Zagros Fault (Fig. 2). The near top Bangestan Group (Upper Cretaceous carbonate unit) was selected for unfolding since it is well-preserved at surface and well-imaged on seismic lines. Our calculation represents a minimum value, since we do not account for internal deformation as layer parallel shortening.

The Zagros deformation front can be traced subparallel to the Iranian coast in the Persian Gulf. To the SW of this deformation front structures are not directly related to the frontal deformation of the Zagros Fold Belt (Fig. 2). The Persian Gulf along the section is characterized by N-S-trending low amplitude and long wavelength anticlines formed by dextral reactivation of inherited transverse faults (Burberry et al., 2011; Soleimany et al., 2011), and by the Hormuz and Fars salt diapirs (Jahani et al., 2007; Stewart, 2018; Ezati-Asl et al., 2019).

Figure 2

THE DOWLATABAD GROWTH SYNCLINE

The Dowlatabad syncline, is located close to the front of the Fars arc, 40 km to the northeast of the Persian Gulf coastal line, and 250 km to the southwest of the Zagros suture zone (Figs. 1B and 2 for location). Characteristic pre-, syn- and late-growth strata patterns allow to define the relative

166 timing of the deformation in this key area of the frontal portion of the Fars arc (Figs. 3 and 4).
167 Furthermore, deposition of about 3.5-km-thick syntectonic sediments make the syncline an ideal
168 structure to date the foreland succession. The foreland stratigraphy in the Dowlatabad syncline is
169 composed of a marine to continental mixed evaporite-carbonate-siliciclastic sequence including
170 Gachsaran Fm. evaporites, Mishan Fm. marls and carbonates (including the Guri Mb. limestone),
171 and the upward coarsening Aghajari Fm. mudrocks, sandstones and conglomerates, ending with the
172 Bakhtyari Fm. conglomerates at top (Fig. 5).

173 **Structure of the Dowlatabad Syncline: Pre-growth, Growth and Post-growth strata**

174 The 12-km-wide Dowlatabad syncline is limited by the Sefid anticline to the north and the NW
175 segment of the Pazan anticline to the south (Fig. 3). The Sefid anticline is an approximately 45-km-
176 long and 11-km-wide, NW-SE trending, doubly plunging anticline, with an amplitude of 2.9 km.
177 The anticline plunges 10° toward 305° at the NW nose, and 5° toward 112° at the SE nose. The stiff
178 limestone of the Guri Mb. exposed in the core of the Sefid anticline forms the fold shape at surface
179 that shows a simply rounded geometry with an opening angle of ~115°. The SW flank of the Sefid
180 anticline is partially covered by unconformable Bakhtyari conglomeratic units (Fig. 3).

181 The Pazan anticline is located to the south of the Dowlatabad syncline. The anticline is formed
182 by two originally separate anticlines labelled as NW Pazan and SE Pazan that have an en-echelon
183 oblique linkage together (Fig. 3A), as described in several examples in Zagros (Bretis et al., 2011;
184 Fernandez and Kaus, 2014; Collignon et al., 2015). The NW Pazan is a 30-km-long and 14-km-
185 wide, SW-verging fold, with an amplitude of 3.85 km. At surface, the anticline shows an
186 asymmetric and rounded geometry with an opening angle of ~125°. It plunges 15° toward 308° at
187 the NW nose and 7° toward 100° at the SE nose. The SE Pazan anticline shows a symmetric
188 geometry and plunges 10° toward 290° at its NW nose. The formerly NW and SE Pazan anticlines

could grow with slightly different ages, as deduced by their change in direction to the south of Dowlatabad and by the distribution of growth strata units as described below.

The integration of our field and remote-sensing mapping, together with seismic interpretations by Motamedi et al. (2012) and Najafi et al. (2014) permits to construct a geological cross-section across the Dowlatabad syncline (section CD; Fig. 3B). The syncline shows a closed asymmetric chevron-type geometry at surface whereas its deeper geometry is open and displaying a box-fold shape as imaged in the published seismic lines across the Frontal Fars (Fig. 3B).

The Dowlatabad syncline, especially at the Aghajari level, shows a smooth topography related to recent river erosion that does not favour the identification of pre-growth, growth and post-growth strata in a cross section. The change between pre-growth and growth units, and therefore the local folding onset, is identified by comparing the thicknesses in the sampled section CD with the EF section located ~10 km to the NW along the SW flank of the syncline, clearly showing a reduction in bedding thickness towards NW (location in Figs. 3A and 4).

Along the SW flank of the Dowlatabad syncline the Aghajari Formation shows two units, a lower unit with constant thickness, and therefore pre-growth, and an upper unit that displays a significant thinning towards the NW termination of the NW Pazan anticline (Fig. 4). The unconformity at the base of growth strata is labelled U1 (Figs. 3A and 4). The correlation of bedding along the flank of the syncline has been done by remote-sensing mapping and field measurements indicating a 1700 m thick middle and upper parts of the Aghajari Fm. growth strata in section CD thinning to ~950 in section EF (SW of Dowlatabad village) and eventually to about 500 m towards the NW towards the fold ending (Fig. 3A and 4). The middle-upper part of the growth strata unit is characterized by an internal ~5° angle unconformity in the centre of the syncline (labelled U2 in Figs. 3A and 4). This unconformity does affect only few meters of Aghajari deposits and evolves to a parallel contact with underlying Aghajari beds along the study

transect CD (Fig. 4). Therefore, unconformity U2 should not mean any problem in interpreting the magnetostratigraphy results but is important for the characterization of the growth unit internal configuration (location in Figs. 3A and 4). The Aghajari succession just above unconformity U2 contains the first appearance of carbonate conglomerates sourced from the Guri limestone providing important information about the progressive dismantling of the growing Pazan anticline.

The thick Aghajari growth unit is in turn eroded by the base of the Bakhtyari Formation, unconformity U3, and therefore showing the characteristic depositional wedge pattern of growth strata (Fig. 3A). In the NE flank of the syncline, Aghajari beds dipping between 35° and 60° crop out beneath the unconformity at Bakhtyari base. Above U3 Bakhtyari conglomerate beds gently dip (<15°) and downlap towards the NE (Figs. 3 and 4). This indicates that the conglomeratic strata did not accommodate a considerable amount of deformation and were therefore late-growth in this locality as reported elsewhere in the NW Zagros (Homke et al., 2004), northern Dezful embayment (Najafi et al., 2018), and in the Fars (Mouthereau et al., 2007).

Figure 3

Figure 4

Stratigraphy of the Dowlatabad Pre-growth and Growth Successions

The Miocene to Pleistocene siliciclastic foreland succession deposited coevally with the progressive growth of the Zagros Fold Belt constitutes the upper-most part of the Zagros sedimentary cover (James and Wynd, 1965; Kashfi, 1980; Koop and Stoneley, 1982; Khadivi et al., 2010; Ruh et al., 2014; Vergés et al., 2019; Etemad-Saeed et al., 2020). In this study, the Zagros foreland succession was examined across the Dowlatabad growth syncline in the Frontal Fars arc for the first time (Fig. 5). The Miocene to Pleistocene stratigraphy constituted by the Gachsaran,

Mishan (including the Guri Mb.), Aghajari and Bakhtyari formations that is deposited on top of the late Oligocene–early Miocene Asmari carbonate platform.

The Gachsaran Fm. is composed of about 500 m mudrocks, halite, gypsum, anhydrite, and thin-bedded carbonates in the SW flank of the Dowlatabad syncline. However, it grades to the detrital Razak Fm. towards the Interior Fars (James and Wynd, 1965; Motiei, 1993). In the Dowlatabad syncline, the Mishan Fm. overlays the Gachsaran evaporites and consists of 1060 m green and grey mudrocks intercalated with well-bedded yellowish and grey fossiliferous carbonates. The Mishan Fm. is composed of ~300 m of very thick-bedded to massive reefal limestones of the Guri Mb. at its base (Figs. 5A and B). The Mishan Fm. is deposited in a shallow open marine environment in the Frontal Fars arc (Rahmani et al., 2010; Pirouz et al., 2015).

Above the marine Mishan Fm., the Aghajari Fm. was deposited into a siliciclastic shoreline and meandering fluvial system across the Zagros foreland basin (James and Wynd, 1965; Najafi et al., 2018; Pirouz, 2018). The Aghajari Fm. shows a ~2000-m-thick coarsening upward detrital succession and consists of light-grey to reddish medium-thick sheet-like sandstone beds and red mudrocks. The sandstone beds present abundant sedimentary paleocurrent indicators including flute casts, cross-stratifications and ripple marks (Fig. 5C). The first appearance of conglomerate and pebbly sandstone layers in the Aghajari Fm. is observed around point 2900 m on the stratigraphic log, above the U2 unconformity (Fig. 5A). The conglomerates are monomictic and composed of medium–well sorted and poorly–medium rounded carbonate pebbles in the Dowlatabad syncline (Fig. 5D).

The Bakhtyari Fm. is deposited into a braided and high-energy fluvial system (James and Wynd, 1965; Najafi et al., 2018; Pirouz, 2018). The Bakhtyari Formation represents the uppermost late- to post-growth unit in the Dowlatabad syncline and is composed of more than 300 m of

coarsening-upward, clast-supported and imbricated polymictic conglomerates, which are intercalated with mudstones and cross-stratified red sandstone channels (Fig. 5E).

Figure 5

Provenance of the Aghajari-Bakhtyari Deposits Filling the Dowlatabad Syncline

Paleocurrent indicators, including cross-stratifications, flute casts and clast imbrications, were measured in 13 sites across the Dowlatabad syncline (Fig. 5). Measured paleocurrents within the Aghajari show great dispersion but large numbers are directed towards the ESE and SE with few NNW and NNE directions. A major change is seen in the few measures collected in the Bakhtyari conglomerates displaying S and SSE directions. The ESE to SE directions seem to indicate well-developed Aghajari longitudinal fluvial systems flowing subparallel to foreland and to growing anticlines. This fluvial system is pushed towards the SW by the arrival of S to SSE Bakhtyari transverse alluvial system as discussed in Vergés (2007) and Pirouz et al. (2011), among others. The conglomerates of the Aghajari Fm. are dominated by sub-angular limestone clasts, mainly sourced from the Guri Mb., pointing towards a local source area from adjacent and nearby growing anticlines (Fig. 5D). However, on top, the conglomerates of the Bakhtyari Fm. document the input of well-rounded metamorphic, igneous and radiolarian chert clasts derived from distal sources in the Zagros orogenic core (Fig. 5E). These observations point to an up section change in sediment provenance, from proximal carbonate sources supplied by outstanding anticlines in the Frontal Fars during the deposition of the Aghajari Fm. to distal source areas from the Zagros hinterland during the deposition of the Bakhtyari Fm. Provenance results from the Azanak growth syncline in northern Dezful embayment show substantial differences in both compositions and ages with respect to the study area (Etemad-Saeed et al., 2020), indicating the complexity of the drainage system in the Mesopotamian foreland basin.

DATING OF THE DOWLATABAD SECTION

In order to define the absolute ages of foreland succession in the frontal Fars, a new magnetostratigraphic section was sampled across the Dowlatabad growth syncline. The magnetostratigraphic ages constrained from 208 drilled sites were combined with ages provided by 20 samples of calcareous nannoplankton and 7 $^{87}\text{Sr}/^{86}\text{Sr}$ isotopes ages (Figs. 3A, 4 and 5A for locations of samples). Integration of all these results on the Gachsaran, Mishan, Aghajari and Bakhtyari formations provided a precise chronostratigraphic framework. The analysed succession along CD section is characterized by maximum thickness recording continuous deposition and is not apparently affected by any of the unconformities described in the syncline.

Magnetostratigraphy

The Dowlatabad (DW) magnetostratigraphic section is ~2-km-thick, its base is located at 28.076883° N, 52.327367° E and its top at 28.118167° N, 52.338550° E (Figs. 3A, 4 and 5A). The DW section begins at the top of the marls of the Mishan Fm., includes the entire shallow marine and fluvial Aghajari Fm. and ends up within the coarse fluvial deposits of the Bakhtyari Fm. 208 sites were drilled along the ~2200 m stratigraphic section with an average sampling resolution of 10 m/site. Two samples per site were obtained with a water-cooled electrical power drill with diamond core bits cooled by water. The samples were oriented in situ by a magnetic compass fixed to an orienting device with a clinometer. Sampling focused on fine-grained materials such as, mudstones, siltstones and fine-grained sandstones, as they are more efficient in recording the Earth's magnetic field than coarse-grained materials.

Paleomagnetic analyses were carried out in the Paleomagnetic Laboratory of Barcelona (CCiTUB-Geo3Bcn, CSIC). They consist of stepwise thermal demagnetization and subsequent measurement of the natural remanent magnetization (NRM) of 1 sample per site. Thermal

304 demagnetizers MMTD80 (Magnetic Measurements Ltd.) and TSD-1 (Schönstedt) and a
 305 superconducting rock magnetometer SRM755R (2G Enterprises) were used for this purpose.
 306 Additionally, undesired generation of new magnetic minerals during heating was controlled by
 307 measuring the bulk magnetic susceptibility after each demagnetization step with a KLY-2
 308 susceptibility bridge (Geofyzika Brno). Up to 16 demagnetization steps were applied to the samples
 309 progressively tightening the temperature increments from 100 °C to 50, 40 and 30 °C up to 670 °C.
 310 Demagnetization data are represented in vector-end point orthogonal diagrams (Zijderveld, 1967)
 311 (Fig. 6A). In most of the samples, a viscous component parallel to the recent magnetic field is
 312 observed, which is usually removed below 200 °C and which was not considered for further
 313 analyses. Above this temperature, the characteristic components usually kept stable up to 670 °C,
 314 pointing to hematite as the main remanence carrier. Characteristic components (ChComp) for each
 315 sample were obtained by means of linear regression (Fig. 6A). ChComp's were classified as class 1
 316 and 2 depending on yielding errors (class 1: below 10; class 2: below 20, Table S1). Finally, class 3
 317 was assigned to those components in which polarity could be deduced but the components were
 318 either obtained with only two points or yielded higher errors. Class 3 directions were not
 319 considered as reliable when building the local magnetostratigraphy, but this was not a big issue as
 320 only 8 class 3 sites were obtained, representing less than 4% of the total directions. Mean directions
 321 for both normal and reversed polarity sets (Fig. 6B) suggest no vertical axis rotations in Frontal
 322 Fars, in agreement with previous magnetotectonic studies in this area (Aubourg et al., 2008). Best
 323 grouping of the directions after tectonic correction point to a primary origin of the magnetization.
 324 Moreover, these directions passed the reversal test with a B classification (γ_c : 8.0 and γ_o : 4.1). The
 325 obtained directions were then computed to calculate the virtual geomagnetic pole paleolatitude
 326 (VGP) at the site level (Fig. 7, Table S1). Positive VGPs are interpreted as normal polarity and
 327 negative VGPs latitudes as reverse polarities and they are represented accordingly when building
 328 the local magnetostratigraphic section (Fig. 7, black for normal magnetozones and white for the

reversed ones). A total of 14 magnetozones were defined by at least two consecutive sites of the same polarity (Fig. 7). The obtained local magnetostratigraphy has been correlated to the Global Polarity Time Scale (Gradstein and Ogg, 2012) (Fig. 7) in order to get the absolute depositional ages. It can be observed that there is a very good match between the local magnetostratigraphy and the GPTS, which favours an independent correlation. Nevertheless, it is also well supported by the chronological constraints yielded by the Sr and calcareous nannofossil samples from the Mishan Fm. and Guri Mb (Fig. 8), as it will be discussed in the following section.

The magnetostratigraphic succession sampled along CD section, in the SW flank of the Dowlatabad syncline, is the thickest in the syncline and thus records the most continuous deposition. In addition, the section is away from the unconformities described in the syncline (Fig. 4). However, the beginning of growth above U1 is recorded by a shorter reverse magnetozones in the local magnetostratigraphy (R2) compared with the associated chron in the GPTS. This implies either a strong decrease in the sedimentation rate (from 74 to 24 cm/ky, Fig. 8) or even a short stop in sedimentation not longer than few tens of thousands of years since reversed polarities are maintained after the growth (Figs. 7). The intra-growth strata U2 does affect only few meters of Aghajari deposits and evolves to a parallel contact with underlying Aghajari beds along the study transect CD (Fig. 4). The U3 at the base of late-growth Bakhtyari conglomerate occurs around the boundary between N7 and overlying R7 of local polarity corresponding to C1 and C2 chrons in the GPTS, respectively. However, no effect of this unconformity is observed in the reversal pattern, since the relative length of the magnetozones is preserved, implying that no significant time is missed in the sampled section (Fig. 7). In general, the longer chron duration and lower sedimentation rate reduces the ability of magnetostratigraphic data in detecting short time hiatuses, as observed in the Southern Pyrenees (Garcés et al., 2020) and in the Fars arc of Zagros (Ruh et al., 2014). Despite unconformities, the good match of this magnetostratigraphic log with the GPTS is a strong indication of a correct correlation as discussed in Garcés and Beamud (2020).

Figure 6

Figure 7

$^{87}\text{Sr}/^{86}\text{Sr}$ Isotopic Dating

Strontium Isotope Stratigraphy is a dating method based on the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio in the composition of marine authigenic sediments that record the chemical composition of the sea water at the time of mineral precipitation (Elderfield, 1986). The ratio of $^{87}\text{Sr}/^{86}\text{Sr}$ in sea water has changed significantly through geological times in response to the input of varying proportions of Sr from continental crust and upper mantle sources. These variations constitute a global standard that has been widely used for dating of carbonate sediments. Rapid monotonous increase in the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio from 0.7082 to 0.7092 during the Neogene, makes this period very promising for Sr dating as age uncertainties are less than 1 Ma (McKenzie et al., 1988; Hodell et al., 1991; McArthur et al., 2001). In this study, 7 samples were taken from limestones and marly limestones of the topmost part of the Gachsaran Fm., Guri Mb. and Mishan Fm. (Fig. 8). Following the quality criteria for Sr dating (Elderfield, 1986), samples were processed at the “Laboratorio de Geocronología y Geoquímica Isotópica of the Universidad Complutense de Madrid” in Spain. The obtained results are summarized in Table S2. The numerical ages were determined by plotting the obtained $^{87}\text{Sr}/^{86}\text{Sr}$ values on the $^{87}\text{Sr}/^{86}\text{Sr}$ LOWESS (Locally Weighted Scatterplot Smoother) calibrated curve (McArthur et al., 2001) (Fig. 8 and Table S2). Five samples of the Guri Mb. at the base of Mishan Fm. (SR81, SR82, SR83, SR85, SR76) cluster between 12.2 and 11 Ma (Fig. 8), suggesting an age around 11.5 Ma for the deposition of this unit in the front of Fars arc. This age is in agreement with reported ages between 11.5 ± 0.5 and 11.1 ± 0.1 Ma for the Mishan Fm. samples from Coastal Fars next to the modern coast of the Persian Gulf (Pirouz et al., 2015). However, two samples from topmost part of the Gachsaran Fm. (SR80 and SR93) show an age around 15 Ma.

377 **Calcareous Nannofossils**

378 In this study, 20 samples were collected to obtain the calcareous nannofossils content in the
379 Mishan Fm. marls and first meters of the Aghajari Fm. (Figs. 3A, 4, 5A and 8, for sample
380 locations). All the samples were prepared from unprocessed material as smear slide using standard
381 procedure (Bown, 1998). The smear slides were analysed using a Zeiss Axioplan petrographic
382 microscope at different magnification (x 750-x 1250). 15 of the 20 samples provided positive
383 results (Table S3). The studied samples were, in general, of poor quality due to the poor
384 preservation and low abundance of the autochthonous assemblages and the presence of abundant
385 reworked species (more than the 50% of the assemblage, in most of the samples). The lowermost
386 samples (NF77–NF84) contained almost no nannofossils, and the few observed ones were usually
387 poor preserved, most of them being reworked dissolution resistant species. The calcareous
388 nannofossil abundance and preservation improves up-section. Samples NF87, NF88 and NF92
389 were rich in moderately to well-preserved nannofossils, but most of them were reworked species.
390 Finally, from sample NF94 onwards both abundance and preservation got worse again. Table S3
391 summarizes the obtained results for Neogene species. The standard schemes of Martini (1971) and
392 Okada and Bukry (1980) have been adopted for this study. Cretaceous and Paleogene reworked
393 species are not included in order to simplify the presentation of results. It is worth noticing the
394 reduced amount in both number of species and number of specimens of the *Discoaster* genera in
395 the studied assemblages, pointing to a deposition in a shallow restricted sea.

396 Regarding biostratigraphy, we have obtained disparate results along section. Due to scarcity and
397 low preservation of calcareous nannofossils in the lower part of the section, it was difficult to
398 assign them to a biozone. In sample NF77 autochthonous calcareous nannofossil assemblages are
399 absent whereas samples NF81–NF84 showed only some Miocene species with a wide stratigraphic
400 range such as *Calcidiscus leptoporus*, *Reticulophenestra haqii*, *Unbilicospaera jafari* and

Umbilicosphaera rotula. In Table S3 the lowermost zone that can contain the observed assemblage is indicated. Samples NF81–NF84 are, thus, related to zone NN1–NN2, although they could belong to a higher zone. Samples NF85 and NF86 belonged to Zone NN7 as they recorded the presence of *Calcidiscus macintyre* and *Reticulofenestra pseudoumbilicus* (Table S3). The preservation of the assemblages, however, is still poor and thus they could pertain to a younger zone. Samples from the middle part of the section were the best quality ones. The presence of *Discoaster quinquerramus*, *Discoaster berggrenii* and *Discoaster bergeni* in samples from NF87 to NF92 indicated that this stratigraphic interval belonged to the lower part of Zone NN11a (Zone CN9a). There were no evidences suggesting a younger age, as species from genera *Amaurolithus* (CN9b) or *Ceratolithus* (CN10b, NN12) were not identified. However, its absence may obey to the general poor preservation state of the nannofossils assemblages. Actually, species from genera *Catinaster* (NN8–NN10) or *Triquetrorhabdulus* (Paleogene–NN12) were neither identified. The uppermost samples (NF94–NF97) were, again, of bad quality hindering any Zone assignation.

Integrated Dating Results

In the Dowlatabad section, the oldest sampled part of the Gachsaran Fm. shows Sr isotopic ages of ~15.5 Ma. In the Guri Mb. at the base of the Mishan Fm., despite some uncertainties, most of the samples yield Sr isotopic ages around 11.5 Ma, between 12.5 and 11.1 Ma.

On top of the Mishan Fm., samples NF87, NF89, NF90 and NF92 show the only calcareous nannofossil assemblages that provide an accurate zone determination. This assemblage is assigned to zone NN11a, with an age ranging from ~8 to 5.5 Ma (Fig. 8). The moderate to poorly preserved calcareous nannofossil assemblages identified in samples NF85 and NF86, in the middle part of Mishan Fm. have been assigned to zone NN7, ranging from 11.8 to 10.8 Ma (Fig. 8). This would mean that between sample NF86 and NF87, less than 100 m apart, there would be a time gap of about 2.8–3.8 Myr. In the Dowlatabad section there is no evidences of a sudden change in the

sedimentation rate or the presence of a hiatus between both samples that could explain this long-time gap in less than 100 m of sedimentary succession. Therefore, samples NF85 and NF86 are more likely to belong to younger zones, NN9 and NN10 respectively, and the absence of NN9 and NN10 biomarkers is therefore most likely due to the poor preservation of calcareous nannofossil assemblages of these samples.

The contact between Mishan and Aghajari formations is located close to the base of the magnetostratigraphic section and has an age slightly younger than 4.9 Ma within the C3 chron in the GPTS. Further up, the contact between Aghajari and Bakhtyari formations coincides with the boundary between C1 and C2 chrons, with an age of 1.8 Ma (Fig. 8).

The timing of the onset of deformation in the Dowlatabad syncline is marked by the beginning of growth strata after the deposition of ~230 m of Aghajari Fm. at paleomagnetic site DW226. The correlation with the GPTS shows an age of 4.65 Ma for the contact between pre-growth and growth strata (Fig. 8). The growth succession has a maximum thickness of 2000 m in the syncline and was deposited during 2.85 Myr. The base of the gently folded late-growth Bakhtyari Fm., from paleomagnetic site DW131 upward to DW150, has an age of 1.8 to 1.5 Ma (Fig. 8). Coevally, the reorganization in the drainage network took place from longitudinal to transverse arrangements, evidenced by changes both in clast composition with the input of external clasts, and paleocurrent directions in the Bakhtyari conglomerates (Fig. 8). Therefore, the amplification of the NW Pazan anticline in the Frontal Fars lasted for 2.85 Myr, from 4.65 Ma to 1.8 Ma, which yields a 1.4 mm/yr rate of fold amplification.

The sedimentation rates for the Aghajari Fm. were deduced from the proposed correlation of the local magnetostratigraphic section to the GPTS (Fig. 8). The base of the section is characterized by a change in sedimentation rates from 46 to 74 cm/kyr. The base of growth strata is marked by a significant short period during which there was either a decrease in sedimentation rates to only 26

cm/kyr or a short break in sedimentation above the U1 unconformity. This was followed by a subsequent increase to 79 cm/kyr. Sedimentation rate keeps high, ranging between 62 and 75 cm/kyr, during most of the growth development. The end of growth occurred with sedimentation rates around 62 cm/kyr and they slightly increased to 68 cm/kyr close to the contact with the Bakhtyari Fm. The estimation of the sedimentation rate for the Bakhtyari Fm. is problematic since there are few tie points within the formation. About 200 m of the Bakhtyari sediments were deposited from 1.8 to 1.5 Ma in the Dowlatabad syncline, which yields a sediment accumulation rate of 68 cm/kyr. These comparatively high sedimentary accumulation rates resemble those determined in the northern Dezful embayment where there is a significant accumulation of syntectonic deposits (Najafi et al., 2018; Lashgari et al., 2020).

Although much less constrained, from top Guri Mb. at ~11.5 Ma to base of the Aghajari Fm. at ~4.9 Ma, the marls of the Mishan Fm. would have been deposited at a rate of 12 cm/kyr. Using these 12 cm/kyr deposition rates of the Mishan Fm., the age of the top of Gachsaran Fm. can be assumed as ~14 Ma. This age would be in agreement with the results of the SR80 and SR93 isotopic samples (Fig. 8).

Figure 8

NUMERICAL MODELLING

Integrated dating of the Miocene sequences and related progressive and angular unconformities of the Dowlatabad syncline give insight into the timing of local folding and the switch from marine to non-marine sedimentary systems. A comparison of the obtained data to previous studies of folding ages across the Zagros Fold Belt (Homke et al., 2004; Khadivi et al., 2010; Ruh et al., 2014; Pirouz et al., 2017; Vergés et al., 2019; Lashgari et al., 2020) provides a better understanding of the temporal evolution of deformation propagation at a large-scale.

However, the effects of basement faulting on the cover folding propagation is difficult to discriminate from dating growth folding in the cover sequence alone. Therefore, a couple of numerical experiments were designed to better understand the large-scale structural evolution of a basement-involved fold-and-thrust belt and the deformation front propagation of its cover sequence in particular.

Methods and Model Setup

The numerical model is based on a thermo-mechanical fully-staggered finite difference code with a visco-elasto-plastic/brittle rheology. For a more detailed description of the governing equations and their numerical implementation, we refer to Ruh (2017) and Ruh et al. (2018). The applied Eulerian box size measures 600 x 60 km in width and height, respectively, with a resolution of 1201 x 121 nodal points and 16 Lagrangian markers per node cell (Fig. 9). From bottom to top it consists of 25 km of diabase basement, 2 km of low-viscous detachment (equivalent to the Ediacaran–Early Cambrian Hormuz salt) between $100 \text{ km} < x < 550 \text{ km}$, and 10 km of upper sedimentary crust with two weak interlayers (equivalent to the pre-folding Phanerozoic sedimentary sequence of the Zagros Fold Belt). The top 23 km exhibit a so-called sticky-air layer of low density and viscosity to ensure zero stresses along the evolving rock/air interface, mimicking a free surface (Crameri et al., 2012). Rheological parameters and density of all marker types are given in Table S4. The initial temperature distribution describes a linear geotherm within the rock sequence with 0 °C at the rock/air interface and 600 °C at the base of the model. Velocity boundary conditions resemble those from sandbox experiments. At the base and the left side, a velocity of 1 cm/yr is applied, whereas the right side represents a rigid backstop with zero velocities in either direction (Fig. 9). The rock/air interface undergoes linear diffusion with a diffusivity coefficient of $10^{-6} \text{ m}^2/\text{s}$ to mimic erosion and sedimentation during folding (Ruh et al., 2018). All models in this study ran for 15 Myr to accommodate shortening equal to 25% of their original length.

In the Fars arc of the Zagros, basement reverse faulting is well-documented for the High Zagros Fault, Surmeh Fault, and Mountain Front Fault (Figs. 1B and 2) dominantly based on seismicity (Jackson, 1980; Berberian, 1995; Talebian and Jackson, 2004; Karasözen et al., 2019). These studies showed that fault planes of basement faults are relatively steep (40–50°), pointing towards reactivated inherited normal faults. To investigate deformation propagation within the upper crustal sequence and to furthermore test the effects of inherited basement faults (see Table S4 for rheological parameters) on deformation propagation and overall structural evolution, two experiments are presented here (Fig. 9): i) Model 1 lacks of predefined inherited basement faults and ii) Model 2 exhibits two inherited listric faults (initially at $x = 400$ km and $x = 550$ km) with an angle of 45° at the top of the basement shallowing downwards, geometrically comparable to the interpreted basement thrusts in the Fars (Fig. 2). Although the listric basement thrusts in the Zagros are relatively shallow (30–45°; Fig. 2), they were interpreted as reactivated listric normal faults (Jackson, 1980).

Figure 9

Modelling Results

Figure 10 illustrates the temporal evolution of numerical experiments without (Fig. 10A–C) and with pre-existing listric faults within the basement (Fig. 10D–F). In the case of no inherited basement faults, marker composition (upper panels) indicates significant basement thrusting close to the backstop wall (<120 km) after 5 Myr of shortening (Fig. 10A). Strain-rate plots infer diffuse rather than localized brittle failure of the basement up to 500 km away from the backstop wall. At this stage, the cover sequence is folded at $x \approx 150$ –250 km (~100 km in front of the most active basement thrust). Developing surface basins (synclines) between anticlines are filled by syn-orogenic sediments (Fig. 10; yellow colour). After 10 Myr of shortening, a sequence of faults, each with ~1–2 km offset, developed within the basement up to ~270 km away from the backstop wall (Fig. 10B). The uppermost cover sequence shows typical fault-propagation folding with thrusts in

the anticlinal hinges. Folding is more intense above thrust basement ($x = 100\text{--}300$ km) and at the salt pinch-out at $x \approx 460$ km. Although basement faults may exhibit offsets of several kilometres, they do not crosscut the salt detachment, which allows the cover sequence to migrate forward along it. Strain rates indicate that after 15 Myr of shortening, active deformation takes place in a thin-skinned manner in front of the salt pinch-out and its accompanying basement thrust (Fig. 10C).

Inherited lower-strength basement faults strongly affect the structural and mechanical evolution of numerical fold-and-thrust belts presented here. After 5 Myr of shortening, a significant amount of shortening is accommodated by the basement fault closer to the backstop at $x \approx 350$ km (Fig. 10D). A large-wavelength anticline forms above the reactivated basement fault. The second invariant of the strain-rate tensor furthermore indicates that also the frontal basement fault beneath the salt pinch-out slips early during the shortening process. The downward dispersion of strain rate localization of reactivated basement faults at $y \approx 45$ km indicates the brittle ductile transition. After 10 Myr, both basement faults exhibit offsets of $\sim 3\text{--}5$ km and the upper crustal sequence is shortened above the salt detachment by multi-layer fault-propagation folding (Fig. 10E). After 15 Myr of shortening, major out-of-sequence basement faulting occurs close to the backstop ($x \approx 50\text{--}150$ km), affecting the style of deformation in the upper crustal sequence (Fig. 10F). Activity along the frontal basement fault leads to thin-skinned shortening of sequences above intermediate weak layers pushed by the hangingwall ($x \approx 450\text{--}500$ km).

Summarizing, if there are no pre-existing weak faults within the basement, the upper crustal sequence shows a roughly constant wavelength of folding/thrusting, forming a gently dipping slope (Fig. 10C). A major forward-verging thrust fault roots into where the salt pinches out at the wedge toe. The basement is faulted regularly beneath the weak salt layer by mainly forward-verging thrusts. Reactivation of two inherited listric faults in the basement leads to more intense strain localization along them during shortening (Fig. 10F). Areas around the inherited faults experience less deformation than in the experiment without inherited faults. Reactivation of those basement

faults strongly affects deformation of the upper crustal sequence, where long-wavelength folds develop above the uplifted basement structures. Basement deformation is also restricted along the backstop and the inherited faults in case of two pre-existing faults (Fig. 10F). The overlying sedimentary sequence is folded with long-wavelength anticlines above the reactivated faults.

Figure 10

DISCUSSION

This section is organized around three major topics: a) Early Miocene–Pleistocene propagation of folding and migration of sedimentary belts across the Fars arc; b) Deformation propagation and effects of basement-involvement inferred from numerical modelling; and c) Timing of folding in the Zagros and its implications for the Arabia-Eurasia collision.

Early Miocene–Pleistocene Propagation of Folding and Migration of Sedimentary Belts Across the Fars Arc

The results of our study provide high-resolution age constraints on a well-preserved and continuous foreland succession as well as the timing of onset and cessation of folding in the Frontal Fars arc (Fig. 8). These results are then integrated with published magnetostratigraphic sections across the Fars arc in SE Zagros Mountains to establish a new Miocene–Pleistocene chronostratigraphic panel along a NE-SW trend, from the footwall of the Main Zagros Thrust (MZT) to the coast of the Persian Gulf (Fig. 11A). The most striking fact from the revised Miocene–Pleistocene chronostratigraphy for the Fars region is that the sedimentary units grouped in formations are diachronous. The transition between marine marls of the Mishan Fm. and the Aghajari continental sediments is dated to 18.8 Ma in the Dehmoord syncline ahead of the Main Zagros Thrust (Pirouz et al., 2017), whereas its age is 16.6 Ma in the Qalat syncline (Khadivi et al.,

2010) and 6 Ma in the Ghol Ghol syncline in Interior Fars (Ruh et al., 2014). Finally, it gets as young as 4.9 Ma in the Dowlatabad growth syncline in the frontal part of the onshore SE Zagros (this study). The contact between the Aghajari Fm. and the conglomerates of the Bakhtyari Fm. occurs at 16.5 Ma in the footwall of the Main Zagros Thrust (Pirouz et al., 2017), whereas it is 14.8 Ma in the Qalat syncline (Khadivi et al., 2010), and 3.2 Ma in the Ghol Ghol syncline in Interior Fars (Ruh et al., 2014), and eventually 1.8 Ma in the Frontal Fars (this study). The lateral facies changes between non-marine Aghajari and shallow marine Mishan formations was already interpreted in the pioneering work of James and Wynd (1965) in the Fars.

We consider the age variation of both Mishan–Aghajari and Aghajari–Bakhtyari contacts across the Zagros to calculate the rate of migration of the sedimentary belts (progradation) toward the foreland (Fig. 11). The Aghajari and Bakhtyari sedimentary belts have been migrating with rates of 18.9 mm/yr and 17.9 mm/yr, respectively, across the SE Zagros Mountains. The same calculation for the NW Zagros shows that the rate of the Aghajari sedimentary belt migration of 19 mm/yr is faster than the rate of the Bakhtyari sedimentary belt migration (11.4 mm/yr). The progressive migration (progradation) of Mishan, Aghajari and Bakhtyari formations towards the foreland is explained by the tectonic propagation in the same direction as in the Lurestan (Pirouz, 2018; Vergés et al., 2019).

Using magnetostratigraphic data, Vergés et al. (2019) calculated Aghajari and Bakhtyari sedimentary belts migration rates in Lurestan of NW Zagros to be 36 mm/yr and 18 mm/yr, respectively. They also show that the folding front propagation toward the foreland occurred at a mean rate of 30 mm/yr. In another study in SE Zagros, Pirouz et al. (2015) proposed between 17.8 and 50 mm/yr for the progradation of the Mishan facies based on Sr dating of carbonates.

The combination of our determined age for the growth strata in the Frontal Fars with published ages across SE Zagros reveals that folding front propagation was non-steady with highly variable

rates (between 8 and 34 mm/yr) during the evolution of the Zagros Fold Belt (Fig. 11). Deformation started around 18 Ma in the Dehmoord syncline ahead of the Main Zagros Thrust (Pirouz et al., 2017). The deformation front propagated very fast, 34 mm/yr, reaching the Qalat syncline at 14 Ma (Khadivi et al., 2010) (Fig. 11A). It was then migrating slowly, 8 mm/yr, across the Interior Fars and reached the Ghol Ghol syncline at 3.8 Ma (Ruh et al., 2014). Almost at the same time (4.6 Ma), folding started in the Dowlatabad syncline in the Frontal Fars arc (this study). Despite these changes, the average rate of folding front propagation from the suture zone to the current deformation front has been around 20 mm/yr from Burdigalian to recent. A similar calculation made in the NW Zagros shows between 7 and 55 mm/yr, with an average rate of 15 mm/yr for the propagation of deformation front, from Aquitanian to Present (Fig. 11B). Although our compilation suggests mean rates of folding front propagation in both SE and NW Zagros (20 and 15 mm/yr, respectively), it also shows that the folding front propagation rates have been significantly variable throughout the lifetime of the Zagros Fold Belt.

The migration of sedimentary belts in a foreland basin is related among other factors to the ratio between the available basin space, because lithospheric bending, and the amount of basin filling sediments: a high ratio produces sedimentary aggradation whereas a low ratio generates progradation of sedimentary sequences. The progradation of sedimentary belts and the propagation of deformation should not follow a similar path as the latter is highly conditioned by factors such as the existence of detachment levels, amount of syntectonic sedimentation, load distribution in the hinterland, and others (e.g., Graveleau et al., 2012). In the Fars arc, however, sedimentary progradation and folding propagation towards the SW occurred with great synchrony as the Aghajari and Bakhtyari calculated migration rates are 18.9 and 17.9 mm/yr and the propagation of folding is 20 mm/yr (Fig. 11).

The rates of syntectonic sediment accumulation vary across the Zagros foreland (Fig. 11A). The sedimentation rate of the Aghajari growth deposits varies between 22 and 28 cm/kyr from the Main

Zagros Thrust towards the Interior Fars (Khadivi et al., 2010; Ruh et al., 2014; Pirouz et al., 2017), reaching the highest rate of 63 cm/kyr in the frontal portion of the Fars arc (Fig. 11A).

As described in section 2, shortening in the Fars arc accounts for about 65 km and 21% during the last ~16 Myr, giving rates of ~4 mm/yr. This almost two-times shortening amount is, however, comparable to the 20% shortening obtained for the Lurestan arc during 22 Myr (Casciello et al., 2009; Vergés et al., 2011b). The higher shortening rate in Fars arc could be achieved by a greater number of thrusts reactivating inherited faults. In both areas, shortening rates are significantly lower than folding front propagation rates of ~15 mm/yr in the Lurestan arc and ~20 mm/yr in the Fars arc. There are two likely causes for this fast propagation of Zagros folding towards the foreland. The reactivation of pre-existing faults in the Arabian basement due to the transmission of far field stresses as was suggested as a main reason (Mouthereau et al., 2007; Farahpour and Hessami, 2012; Pirouz et al., 2015). An alternative or combined explanation would be the presence of Ediacaran–Early Cambrian Hormuz salt-bearing evaporites at the basement-cover boundary (Fig. 2). Numerical and analogue experiments support that the strength of the basal detachment affects the relation between propagation and shortening rates (Cotton and Koyi, 2000; Ruh et al., 2012, 2013).

Figure 11

Effects of Inherited Basement Faults and Basal Detachment Level on Deformation Propagation Inferred from Numerical Modeling

In this study, numerical models are used to explore the propagation of folding across fold belts in an attempt to draw conclusions about the potential roles of a basal detachment level and inherited basement faults, applicable to the vast amount of data from the Zagros Fold Belt compiled in this work. Figure 12 illustrates rock composition after 15 Myr of shortening (upper panels) and

641 averaged horizontal shortening rate of the cover sequence (lower panels) for the entire time span of
642 the respective experiments (0–15 Myr). In the following, we discuss similarities and discrepancies
643 of the two presented experiments with and without deep-seated inherited basement faults and
644 discuss numerical results with respect to observations from the Zagros.

645 Both experiments show a very low surface taper angle ($\sim 1^\circ$), in agreement with calculations in the
646 Lurestan and Fars arcs that are typical of fold belts detached above an efficient evaporitic horizon
647 at depth as proposed by McQuarrie (2004), Ford, (2004), and discussed in Vergés et al. (2011a).
648 Furthermore, both experiments show in-sequence and fast forelandward propagation of folding
649 (lateral migration of shortening) within the cover sequence above the efficient basal detachment
650 level equivalent to the Hormuz salt (Fig. 12). This fast propagation forms a wave of folding
651 deforming strips of previously undeformed foreland that can account for the almost synchronous
652 growth onset of the Ghol Ghol (Ruh et al., 2014) and Dowlatabad (this study) synclines in the Fars
653 arc at 3.8 and 4.6 Ma, respectively (Fig. 11A).

654 Besides the aforementioned common points, there are particular discrepancies between the
655 experiment with and without inherited basement faults. The temporospatial evolution of shortening
656 rates in the experiment without inherited basement faults shows a generally constant rate of folding
657 propagation that slows down during the last ~ 5 Myr (Fig. 12A). A feature of this model is that the
658 basement-cover boundary, i.e. the weak basal detachment, maintains a smooth topography despite
659 having a basement affected by the same amount of shortening as the cover sequence. Another
660 outcome is that folds have a limited duration of growth restricted to a few million years. This is in
661 agreement with field data where the entire folding process was determined by dating the late- or
662 post-growth strata units, like in the Changleh, Jarik and Dowlatabad synclines showing active
663 growth thru 5, 5.8 and 2.85 Myr, respectively (Homke et al., 2004; Lashgari et al., 2020; this
664 study).

In contrast, the experiment with inherited basement faults results in an interrupted pattern of shortening propagation over time (Fig. 12B). Despite the general trend of forward propagation of deformation, zones of very low horizontal shortening (<1 mm/yr) occur above the hangingwall of the two inherited basement faults (Fig. 12B) that underwent relatively early inversion (Figs. 10D–F). These large basement-involved monoclines act as buttressing zones separating cover folding domains where folds remain active throughout the duration of the experiment. The early activity of basement thrusts at depth can explain the relatively early (~ 7.6 Ma) growth of the Changleh anticline actually located above the flexure of the Mountain Frontal Fault in the Lurestan arc (Homke et al., 2004; Emami et al., 2010). Such early growth with respect to the general deformation propagation is also inferred in the southern Fars region by Mouthereau et al. (2006). The longer duration of these basement blind thrusts, specifically their late activity, can be interpreted as partially out-of-sequence since cover folds above the thrust hangingwalls are passively transported. These sequential relationships are observed in the Lurestan front where the forelimbs of the large Anaran and Siah Kuh anticlines gravitationally collapsed along the hangingwall of the Mountain Front Fault (e.g. Emami et al., 2010; Casini et al., 2018). Therefore, although the regional forelandward sequence of deformation is clearly represented in the model, the pre-existence of basement faults results in a more intricate deformation sequence including local out-of-sequence folding.

As a summary, the general sequence in both experiments occurs in-sequence, advancing toward the foreland and thus comparable with magnetostratigraphic ages of folding onset younging in the same direction (Figs. 11 and 12). In the experiment without inherited faults, the propagation of the onset of folding is continuous and anticlines grow for a limited period of time of few million years. In the second experiment, propagation of folding onset shows a more discontinuous pattern where inherited faults concentrate shortening in the basement and thus a more complex scenario, in which local out-of-sequence folding may occur as a result of basement inheritance. However, the present

data set on ages of onset and duration of folding for the Zagros Simply Folded Belt, although abundant and strategically distributed, is probably not sufficient to determine potential local out-of-sequence folding activity that should further be analysed in more detail with multidisciplinary studies. It is worth noting that the in-sequence folding propagation established in this study does not indicate a significant disruption in the propagation wave of folding at a large-scale. This undisrupted pattern indicates that the inverted basement thrusts were not significantly affecting the forward migration of folding. This could be related to a more complex thrusting geometry at depth favored by the distribution of the effective basement-cover interface detachment level that could transfer shortening from deeper thrusting in the hinterland to Hormuz salt or equivalent far away in the foreland. In this scenario, the basement thrusting observed in the Zagros Simply Fold Belt would uplift already folded cover sequences stepwise as illustrated in Lurestan (e.g., Saura et al., 2015) and Dezful (e.g., Etemad-Saeed et al., 2020).

Figure 12

Timing of Folding Across the Zagros and its Implications for the Arabia-Eurasia Collision

The combination of our results in the Dowlatabad growth syncline with eight published magnetostratigraphic sections across the Zagros Fold Belt, indicates that the onset of cover folding is everywhere younger than 22 Ma in both Lurestan and Fars arcs (Fig. 13). The oldest growth ages were determined to be Aquitanian–Burdigalian in the Shalamzar and Dehmoord synclines, ahead of the Main Zagros Thrust (Pirouz et al., 2017) (Fig. 13). However, the onset of Arabia-Eurasia suturing is slightly older and is marked by the first arrival of clastics from the Sanandaj-Sirjan to the Zagros foreland basin as discussed in Mouthereau et al. (2012) and Gholami Zadeh et al. (2017). These clastics are found in red beds of the Razak Fm. in Iran and Suwaise Unit in Iraqi Kurdistan, with a late Oligocene age (Chattian, ~26–25 Ma) based on recently published detrital zircon and magnetostratigraphic data (Pirouz et al., 2017; Barber et al., 2018; Koshnaw et al.,

2018). The slow exhumation of the Sanandaj-Sirjan zone along the hangingwall of the Main Zagros Thrust started at ~24 Ma (Chattian), and accelerated between 20 and 14 Ma (Burdigalian–Langhian), as determined by thermochronological data in Kermanshah and Lurestan in NW Zagros (Barber et al., 2018). The first foreland sediments with Zagros provenance arrived to the Dezful embayment during late Oligocene–early Miocene times (~25–18 Ma) (Etemad-Saeed et al., 2020). However, these risings of tectono-topographic reliefs in the Zagros can be linked to either the onset of continental collision (soft-collision) or to a more advanced stage of orogenic evolution (hard-collision) (Ballato et al., 2011; Mouthereau et al., 2012).

The Arabia-Eurasia collision, therefore, incepted in Chattian (late Oligocene), and the Zagros folding initiated in Aquitanian–Burdigalian (early Miocene). The folding then propagated forelandward through the Interior Fars, Lurestan, Izeh Zone, and north Dezful embayment during the Langhian and Serravallian (middle Miocene), as defined in Afrineh, Chaman-Goli, Jarik and Qalat growth synclines (Khadivi et al., 2010; Vergés et al., 2019; Lashgari et al., 2020). According to these ages, the propagation of folding was occurring similarly along the strike of the Zagros fold belt from Lurestan to Fars arcs during middle and late Miocene times. The good correlation along the strike of the Zagros Fold Belt of onset of folding ages indicates that the present geometry of salients and reentrants bounding the Lurestan arc, the Dezful embayment and the Fars arc does not seem to influence the propagation of the folding wave (Fig. 13). The width of the folded domain was about 160 km at the end of Miocene whereas the maximum present-day width is 230 km in the Lurestan arc and 280 km in the Fars arc. This difference in the width of the Simply Folded Belt, apart from other factors such as the spatial variation of Hormuz salt at depth (Bahroudi and Koyi, 2003), could be influenced by the inherited segmentation of the Arabian margin as proposed in Navabpour et al. (2014). The partly coeval growth of the Ghol Ghol and Dowlatabad synclines in the Frontal Fars arc indicates that a number of close anticlines grew in a period of few million years from early Pliocene (Zanclean) to early Pleistocene (Gelasian) (Fig. 13). The numerical

simulations of the Fars folding by Yamato et al. (2011) and Fernandez and Kaus (2014) also show the coeval growth of several anticlines within a limited range on top of Hormuz salt-bearing basal detachment level as highlighted by our numerical modelling results.

This Neogene folding postdates two earlier pre-collision tectonic events in SE Zagros, i.e., a) the Hormuz salt diapirism from Palaeozoic to Recent (Jahani et al., 2007; Perotti et al., 2016; Stewart, 2018) and, b) the Late Cretaceous reactivation of N-S trending inherited basement faults from the Late Cretaceous (Burberry et al., 2011; Saura et al., 2011; Soleimany et al., 2011; Farahzadi et al., 2019). Both events developed broad and low amplitude antiforms in the sedimentary cover, which are still preserved in the Persian Gulf to the SW of the Zagros deformation front (Fig. 13). To the NE of this front, both diapiric domes and N-S trending folds are overprinted by the NW-SE trending Zagros Fold Belt structures as those presented in this study.

Figure 13

CONCLUSIONS

The ages of the syntectonic foreland deposits of the Frontal Fars domain have been determined in the Dowlatabad growth syncline by the combination of calcareous nannofossil analysis, $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic dating and magnetostratigraphy for the marine to non-marine mixed evaporite-carbonate-siliciclastic successions of Mishan, Aghajari and Bakhtyari formations. These ages have been integrated with provenance and paleocurrent analysis, as well as within the framework of a detailed structural study of the growth syncline and limiting anticlines. Following we summarize the main results of this study.

- The deposition of marine marls of the Mishan Fm. started around 11.5 Ma in the Dowlatabad growth syncline. The transition to a non-marine basin condition occurred at 4.9 Ma by the progradation of thick fluvial deposits of the Aghajari Fm. with a fast accumulation rate of 63

cm/kyr. Monomictic conglomerate beds, above the intra-Aghajari U2 unconformity, composed of Guri Mb. carbonate clasts were sourced from nearby growing anticlines and deposited between 2.8 and 1.95 Ma. At the uppermost part of the foreland succession, the Bakhtyari polymictic conglomerates derived from Zagros hinterland, unconformably cover the Frontal Fars synclines from 1.8 to <1.5 Ma, with an accumulation rate of 68 cm/kyr.

- Folding of the Frontal Fars arc started at 4.65 Ma, as marked by the beginning of growth strata above unconformity U1. Several anticlines grew almost coevally during the early Pliocene. The major pulse of folding in the Dowlatabad growth syncline lasted for 2.85 Myr, ceasing at around 1.8 Ma, in early late Pleistocene. The NW Pazan anticline amplified with a rate of 1.4 mm/yr.

- Age variations of both Mishan–Aghajari and Aghajari–Bakhtyari contacts from the Zagros suture zone towards the deformation front mark the migration rate of foreland sedimentary belts (progradation) that is consistent with the propagation of folding obtained from dating of growth strata across the Lurestan and Fars arcs. The propagation of folding and the progradation of the depositional systems occurred by a mean rate of ~20 mm/yr in SE Zagros (Fars arc), and ~15 mm/yr in NW Zagros (Lurestan arc), during the early Miocene–Pleistocene. Propagation rates vary through time and space between 34 mm/yr in an early stage close to the High Zagros Fault and 8 mm/yr close to the Zagros front.

- High rates of folding propagation are about one order of magnitude larger than age equivalent shortening rates, ~4 mm/yr in Fars and 2 mm/yr in Lurestan, indicating the existence of an efficient detachment level at the base of the folded sedimentary cover enhancing the stress propagation towards the foreland.

- Numerical experiments of basement-involved fold-and-thrust belts, based on the Fars arc parameters, show in-sequence folding comparable to results from magnetostratigraphic ages of

folding across the Zagros Simply Folded Belt. In experiment without inherited faults, the onset of folding is roughly continuous and each anticline grows during few million years whereas the experiment with inherited basement faults displays a more complex scenario with discontinuous onset of folding, longer duration of anticlines growth and local out-of-sequence propagation of deformation.

ACKNOWLEDGEMENTS

This research was funded by the DARIUS Programme as well as the Group of Dynamics of the Lithosphere (GDL), Geosciences Barcelona, Geo3Bcn, CSIC, Spain. We thank the Paleomagnetic Laboratory CCI-TUB-Geo3Bcn, CSIC where the paleomagnetic analyses were conducted, and also the Laboratorio de Geocronología y Geoquímica Isotópica of the Universidad Complutense de Madrid, Spain, where the Sr samples were processed. In addition, the study of calcareous nanofossils was supported by the Research Group of the Basque University System IT-930-16. We are also grateful to the National Iranian Oil Company (NIOC) for their support during fieldwork. We thank DARIUS Programme and its sponsors, and projects Alpimed (PIE-CSIC-201530E082) and Subtetis (PIE-CSIC-201830E039) for additional support. We are grateful to Stefano Tavani and an anonymous reviewer as well as to the Associate Editor for their constructive reviews, which greatly helped to improve the quality of this paper. We dedicate this work to the memory of Mohammad Sepehr, a lover of life and geology, who died unexpectedly during our first reconnaissance trip to the study region.

REFERENCES

Agard, P., Omrani, J., Jolivet, L., and Mouthereau, F., 2005, Convergence history across Zagros (Iran): Constraints from collisional and earlier deformation: *International Journal of Earth Sciences*, v. 94, p. 401–419, doi:10.1007/s00531-005-0481-4.

809 Alavi, M., 2007, Structures of the Zagros fold-thrust belt in Iran: *American Journal of Science*, v.
810 307, p. 1064–1095, doi:10.2475/09.2007.02.

811 Aubourg, C., Smith, B., Bakhtari, H.R., Guya, N., and Eshraghi, a., 2008, Tertiary block rotations
812 in the Fars Arc (Zagros, Iran): *Geophysical Journal International*, v. 173, p. 659–673,
813 doi:10.1111/j.1365-246X.2008.03732.x.

814 Bahroudi, A., and Koyi, H.A., 2003, Effect of spatial distribution of Hormuz salt on deformation
815 style in the Zagros fold and thrust belt: an analogue modelling approach: *Journal of the*
816 *Geological Society*, v. 160, p. 719–733, doi:10.1144/0016-764902-135.

817 Ballato, P., Uba, C.E., Landgraf, A., Strecker, M.R., Sudo, M., Stockli, D.F., Friedrich, A., and
818 Tabatabaei, S.H., 2011, Arabia-Eurasia continental collision: Insights from late Tertiary
819 foreland-basin evolution in the Alborz Mountains, northern Iran: *Geological Society of*
820 *America Bulletin*, v. 123, p. 106–131, doi:10.1130/B30091.1.

821 Barber, D.E., Stockli, D.F., Horton, B.K., and Koshnaw, R.I., 2018, Cenozoic Exhumation and
822 Foreland Basin Evolution of the Zagros Orogen During the Arabia-Eurasia Collision, Western
823 Iran: *Tectonics*, v. 37, p. 4396–4420, doi:10.1029/2018TC005328.

824 Berberian, M., 1995, Master “blind” thrust faults hidden under the Zagros folds: active basement
825 tectonics and surface morphotectonics: *Tectonophysics*, v. 241, p. 193–224,
826 doi:10.1016/0040-1951(94)00185-C.

827 Bown, P.R., 1998, *Calcareous nannofossil biostratigraphy*: Chapman and Hall, Kluwer Academic,
828 London, 315 p.

829 Bretis, B., Bartl, N., and Grasemann, B., 2011, Lateral fold growth and linkage in the Zagros fold
830 and thrust belt (Kurdistan, NE Iraq): *Basin Research*, v. 23, p. 615–630, doi:10.1111/j.1365-
831 2117.2011.00506.x.

832 Burberry, C.M., Jackson, C.-L., and Cosgrove, J.W., 2011, Late Cretaceous to Recent deformation
833 related to inherited structures and subsequent compression within the Persian Gulf: a 2D
834 seismic case study: *Journal of the Geological Society*, v. 168, p. 485–498, doi:10.1144/0016-
835 76492010-022.

836 Casciello, E., Verges, J., Saura, E., Casini, G., Fernandez, N., Blanc, E., Homke, S., and Hunt,
837 D.W., 2009, Fold patterns and multilayer rheology of the Lurestan Province, Zagros Simply
838 Folded Belt (Iran): *Journal of the Geological Society*, v. 166, p. 947–959, doi:10.1144/0016-
839 76492008-138.

840 Casini, G., Romaine, I., Casciello, E., Saura, E., Vergés, J., Fernández, N., and Hunt, D.W., 2018,
841 Fracture characterization in sigmoidal folds: Insights from the Siah Kuh anticline, Zagros,
842 Iran: *AAPG Bulletin*, v. 102, p. 369–399, doi:10.1306/0503171615817076.

843 Collignon, M., Fernandez, N., and Kaus, B.J.P., 2015, Influence of surface processes and initial
844 topography on lateral fold growth and fold linkage mode: *Tectonics*, v. 34, p. 1622–1645,
845 doi:10.1002/2015TC003843.

846 Cotton, J.T., and Koyi, H.A., 2000, Modeling of thrust fronts above ductile and frictional
847 detachments: Application to structures in the Salt Range and Potwar Plateau, Pakistan:
848 *Geological Society of America Bulletin*, v. 112, p. 351–363, doi:10.1130/0016-
849 7606(2000)112<351:MOTFAD>2.0.CO;2.

850 Crameri, F., Schmeling, H., Golabek, G.J., Duretz, T., Orendt, R., Buitert, S.J.H., May, D.A., Kaus,
851 B.J.P., Gerya, T. V, and Tackley, P.J., 2012, A comparison of numerical surface topography
852 calculations in geodynamic modelling: an evaluation of the sticky air method: *Geophysical*
853 *Journal International*, v. 189, p. 38–54, doi:10.1111/j.1365-246X.2012.05388.x.

854 Dashti, F., Lucente, F.P., Motaghi, K., Bianchi, I., Najafi, M., Govoni, A., and Shabanian, E., 2019,

855 Crustal scale imaging of the Arabia-Central Iran collision boundary across the Zagros suture
856 zone, west of Iran: *Geophysical Research Letters*, v. 47, p. e2019GL085921,
857 doi:10.1029/2019GL085921.

858 Ehrenberg, S.N., Pickard, N.A.H., Laursen, G. V, Monibi, S., Mossadegh, Z.K., Svåná, T.A.,
859 Aqrawi, A.A.M., McArthur, J.M., and Thirlwall, M.F., 2007, Strontium Isotope Stratigraphy
860 of the Asmari Formation (Pligocene-Lower Miocene), SW Iran: *Journal of Petroleum*
861 *Geology*, v. 30, p. 107–128, doi:10.1111/j.1747-5457.2007.00107.x.

862 Elderfield, H., 1986, Strontium isotope stratigraphy: Palaeogeography, palaeoclimatology,
863 palaeoecology, v. 57, p. 71–90, doi:10.1016/0031-0182(86)90007-6.

864 Emami, H., 2008, Foreland propagation folding and structure of the Mountain Front Flexure in the
865 Pusht-e Kuh arc (NW Zagros, Iran) [Ph.D. thesis]: Universitat de Barcelona, 199 p.

866 Emami, H., Verges, J., Nalpas, T., Gillespie, P., Sharp, I., Karpuz, R., Blanc, E.P., and Goodarzi,
867 M.G.H., 2010, Structure of the Mountain Front Flexure along the Anaran anticline in the
868 Pusht-e Kuh Arc (NW Zagros, Iran): insights from sand box models: *Geological Society*,
869 London, Special Publications, v. 330, p. 155–178, doi:10.1144/SP330.9.

870 Etemad-Saeed, N., Najafi, M., and Vergés, J., 2020, Provenance evolution of Oligocene–Pliocene
871 foreland deposits in the Dezful embayment to constrain Central Zagros exhumation history:
872 *Journal of the Geological Society*, v. 177, p. 799–817, doi:10.1144/jgs2019-206.

873 Ezati-Asl, M., Faghih, A., Mukherjee, S., and Soleimany, B., 2019, Style and timing of salt
874 movement in the Persian Gulf basin, offshore Iran: Insights from halokinetic sequences
875 adjacent to the Tonb-e-Bozorg salt diapir: *Journal of Structural Geology*, v. 122, p. 116–132,
876 doi:10.1016/j.jsg.2019.02.002.

877 Falcon, N.L., 1961, Major Earth-Flexuring in the Zagros Mountains of South-West Iran: *Quarterly*

Journal of the Geological Society, v. 117, p. 367–376, doi:10.1144/gsjgs.117.1.0367.

Farahpour, M.M., and Hessami, K., 2012, Cretaceous sequence of deformation in the SE Zagros fold-thrust belt: Journal of the Geological Society, v. 169, p. 733–743, doi:10.1144/jgs2012-042.

Farahzadi, E., Alavi, S.A., Sherkati, S., and Ghassemi, M.R., 2019, Variation of subsidence in the Dezful Embayment, SW Iran: influence of reactivated basement structures: Arabian Journal of Geosciences, v. 12, p. 616, doi:10.1007/s12517-019-4758-5.

Fernandez, N., and Kaus, B.J.P., 2014, Fold interaction and wavelength selection in 3D models of multilayer detachment folding: Tectonophysics, v. 632, p. 199–217, doi:10.1016/j.tecto.2014.06.013.

Ford, M., 2004, Depositional wedge tops: interaction between low basal friction external orogenic wedges and flexural foreland basins: Basin Research, v. 16, p. 361–375, doi:10.1111/j.1365-2117.2004.00236.x.

Garcés, M., and Beamud, E., 2020, Geochronology: Magnetostratigraphic Dating., *in* Encyclopedia of Geology. Reference Module in Earth Systems and Environmental Sciences, Elsevier, doi:10.1016/B978-0-12-409548-9.12386-3.

Garcés, M., López-Blanco, M., Valero, L., Beamud, E., Muñoz, J.A., Oliva-Urcia, B., Vinyoles, A., Arbués, P., Cabello, P., and Cabrera, L., 2020, Paleogeographic and sedimentary evolution of the South Pyrenean foreland basin: Marine and Petroleum Geology, v. 113, p. 104105, doi:10.1016/j.marpetgeo.2019.104105.

Le Garzic, E., Vergés, J., Sapin, F., Saura, E., Meresse, F., and Ringenbach, J.C., 2019, Evolution of the NW Zagros Fold-and-Thrust Belt in Kurdistan Region of Iraq from balanced and restored crustal-scale sections and forward modeling: Journal of Structural Geology, v. 124, p.

51–69, doi:<https://doi.org/10.1016/j.jsg.2019.04.006>.

Gholami Zadeh, P. et al., 2017, Revised version of the Cenozoic Collision along the Zagros Orogen, Insights from Cr-spinel and Sandstone Modal Analyses: Scientific reports, v. 7, p. 10828, doi:[10.1038/s41598-017-11042-1](https://doi.org/10.1038/s41598-017-11042-1).

Gradstein, F.M., and Ogg, J.G., 2012, The chronostratigraphic scale, *in* The geologic time scale, Elsevier, p. 31–42.

Graveleau, F., Malavieille, J., and Dominguez, S., 2012, Experimental modelling of orogenic wedges: A review: Tectonophysics, v. 538–540, p. 1–66, doi:[10.1016/j.tecto.2012.01.027](https://doi.org/10.1016/j.tecto.2012.01.027).

Hessami, K., Koyi, H. a., Talbot, C.J., Tabasi, H., and Shabanian, E., 2001, Progressive unconformities within an evolving foreland fold-thrust belt, Zagros Mountains: Journal of the Geological Society, v. 158, p. 969–981, doi:[10.1144/0016-764901-007](https://doi.org/10.1144/0016-764901-007).

Hodell, D.A., Mueller, P.A., and Garrido, J.R., 1991, Variations in the strontium isotopic composition of seawater during the Neogene: Geology, v. 19, p. 24–27, doi:[10.1130/0091-7613\(1991\)019%3C0024:VITSIC%3E2.3.CO;2](https://doi.org/10.1130/0091-7613(1991)019%3C0024:VITSIC%3E2.3.CO;2).

Homke, S., Vergés, J., van der Beek, P., Fernández, M., Saura, E., Barbero, L., Badics, B., and Labrin, E., 2010, Insights in the exhumation history of the NW Zagros from bedrock and detrital apatite fission-track analysis: Evidence for a long-lived orogeny: Basin Research, v. 22, p. 659–680, doi:[10.1111/j.1365-2117.2009.00431.x](https://doi.org/10.1111/j.1365-2117.2009.00431.x).

Homke, S., Vergés, J., Garcés, M., Emami, H., and Karpuz, R., 2004, Magnetostratigraphy of Miocene-Pliocene Zagros foreland deposits in the front of the Push-e Kush Arc (Lurestan Province, Iran): Earth and Planetary Science Letters, v. 225, p. 397–410, doi:[10.1016/j.epsl.2004.07.002](https://doi.org/10.1016/j.epsl.2004.07.002).

- 923 Jackson, J., 1980, Reactivation of Basement faults and crustal shortening in orogenic belts: *Nature*,
924 v. 283, p. 343–346, doi:10.1038/283343a0.
- 925 Jackson, J., and Fitch, T., 1981, Basement faulting and the focal depths of the larger earthquakes in
926 the Zagros mountains (Iran): *Geophysical Journal International*, v. 64, p. 561–586,
927 doi:10.1111/j.1365-246X.1981.tb02685.x.
- 928 Jackson, J. a., Fitch, T.J., and McKenzie, D.P., 1981, Active thrusting and the evolution of the
929 Zagros fold belt: *Geological Society, London, Special Publications*, v. 9, p. 371–379,
930 doi:10.1144/GSL.SP.1981.009.01.34.
- 931 Jahani, S., Dominique, J.C., Lamotte, F. De, Letouzey, J., and Leturmy, P., 2007, The Salt Diapirs
932 of the Eastern Fars Province (Zagros, Iran): A Brief Outline of their Past and Present: In:
933 Lacombe, O., Lavé, J., Roure, F. and Vergés, J. “Thrust Belts and Foreland Basins. From Fold
934 Kinematics to Hydrocarbon System” *Frontiers in Earth Sciences*, Springer. Chapter 15, p.
935 289–308, doi:10.1007/978-3-540-69426-7_15.
- 936 James, G.A., and Wynd, J.G., 1965, Stratigraphic nomenclature of Iranian oil consortium
937 agreement area: *AAPG bulletin*, v. 49, p. 2182–2245, doi:10.1306/A663388A-16C0-11D7-
938 8645000102C1865D.
- 939 Jiménez-Munt, I., Fernández, M., Saura, E., Vergés, J., and Garcia-Castellanos, D., 2012, 3-D
940 lithospheric structure and regional/residual Bouguer anomalies in the Arabia-Eurasia collision
941 (Iran): *Geophysical Journal International*, v. 190, p. 1311–1324, doi:10.1111/j.1365-
942 246X.2012.05580.x.
- 943 Karasözen, E., Nissen, E., Bergman, E.A., and Ghods, A., 2019, Seismotectonics of the Zagros
944 (Iran) From Orogen-Wide, Calibrated Earthquake Relocations: *Journal of Geophysical*
945 *Research: Solid Earth*, v. 124, p. 9109–9129, doi:10.1029/2019JB017336.

946 Kashfi, M.S., 1980, Stratigraphy and environmental sedimentology of Lower Fars Group
 947 (Miocene), south-southwest Iran: AAPG Bulletin, v. 64, p. 2095–2107,
 948 doi:10.1306/2F919742-16CE-11D7-8645000102C1865D.

949 Khadivi, S., Mouthereau, F., Larrasoana, J.-C., Vergés, J., Lacombe, O., Khademi, E., Beamud, E.,
 950 Melinte-Dobrinescu, M., and Suc, J.-P., 2010, Magnetostratigraphy of synorogenic Miocene
 951 foreland sediments in the Fars arc of the Zagros Folded Belt (SW Iran): Basin Research, v. 22,
 952 p. 918–932, doi:10.1111/j.1365-2117.2009.00446.x.

953 Koop, W.J., and Stoneley, R., 1982, Subsidence history of the Middle East Zagros basin, Permian
 954 to recent: Philosophical Transactions of the Royal Society A: Mathematical, Physical and
 955 Engineering Sciences, v. 305, p. 149–168, doi:10.1098/rsta.1982.0031.

956 Koshnaw, R.I., Stockli, D.F., and Schlunegger, F., 2018, Timing of the Arabia-Eurasia continental
 957 collision—Evidence from detrital zircon U-Pb geochronology of the Red Bed Series strata of
 958 the northwest Zagros hinterland, Kurdistan region of Iraq: Geology, v. 47, p. 47–50,
 959 doi:10.1130/G45499.1.

960 Lashgari, A., Heyhat, M.R., Vergés, J., Beamud, E., Najafi, M., Khatib, M., and Karimnejad, H.R.,
 961 2020, Age of Synorogenic Deposits and Timing of Folding in Dezful Embayment, SW Zagros
 962 Fold Belt: Marine and Petroleum Geology, v. 113, p. 104148,
 963 doi:10.1016/j.marpetgeo.2019.104148.

964 Madanipour, S., Ehlers, T.A., Yassaghi, A., and Enkelmann, E., 2017, Accelerated middle Miocene
 965 exhumation of the Talesh Mountains constrained by U-Th/He thermochronometry: evidence
 966 for the Arabia-Eurasia collision in the NW Iranian Plateau: Tectonics, v. 36, p. 1538–1561,
 967 doi:10.1002/2016TC004291.

968 Martini, E., 1971, Standard Tertiary and Quaternary calcareous nannoplankton zonation, *in* Proc. II

969 Planktonic Conference, Roma 1970, Roma, Tecnoscienza, v. 2, p. 739–785.

970 McArthur, J.M., Howarth, R.J., and Bailey, T.R., 2001, Strontium isotope stratigraphy: LOWESS
 971 version 3: best fit to the marine Sr-isotope curve for 0–509 Ma and accompanying look-up
 972 table for deriving numerical age: *The Journal of Geology*, v. 109, p. 155–170,
 973 doi:10.1086/319243.

974 McKenzie, J.A., Hodell, D.A., Mueller, P.A., and Mueller, D.W., 1988, Application of strontium
 975 isotopes to late Miocene-early Pliocene stratigraphy: *Geology*, v. 16, p. 1022–1025,
 976 doi:10.1130/0091-7613(1988)016%3C1022:AOSITL%3E2.3.CO;2.

977 McQuarrie, N., 2004, Crustal scale geometry of the Zagros fold-thrust belt, Iran: *Journal of*
 978 *Structural Geology*, v. 26, p. 519–535, doi:10.1016/j.jsg.2003.08.009.

979 Motamedi, H., Sherkati, S., and Sepehr, M., 2012, Structural style variation and its impact on
 980 hydrocarbon traps in central Fars, southern Zagros folded belt, Iran: *Journal of Structural*
 981 *Geology*, v. 37, p. 124–133, doi:10.1016/j.jsg.2012.01.021.

982 Motiei, H., 1993, *Geology of Iran: Stratigraphy of Zagros*: Geol. Surv. of Iran, 536 p.

983 Mouthereau, F., Lacombe, O., and Meyer, B., 2006, The Zagros folded belt (Fars, Iran):
 984 Constraints from topography and critical wedge modelling: *Geophysical Journal International*,
 985 v. 165, p. 336–356, doi:10.1111/j.1365-246X.2006.02855.x.

986 Mouthereau, F., Lacombe, O., and Vergés, J., 2012, Building the Zagros collisional orogen:
 987 Timing, strain distribution and the dynamics of Arabia/Eurasia plate convergence:
 988 *Tectonophysics*, v. 532–535, p. 27–60, doi:10.1016/j.tecto.2012.01.022.

989 Mouthereau, F., Tensi, J., Bellahsen, N., Lacombe, O., De Boisgrollier, T., and Kargar, S., 2007,
 990 Tertiary sequence of deformation in a thin-skinned/thick-skinned collision belt: The Zagros

- 991 Folded Belt (Fars, Iran): *Tectonics*, v. 26, doi:10.1029/2007TC002098.
- 992 Najafi, M., Vergés, J., Etemad-Saeed, N., and Karimnejad, H.R., 2018, Folding, thrusting and
 993 diapirism: Competing mechanisms for shaping the structure of the north Dezful Embayment,
 994 Zagros, Iran: *Basin Research*, v. 30, p. 1200–1229, doi:10.1111/bre.12300.
- 995 Najafi, M., Yassaghi, A., Bahroudi, A., Vergés, J., and Sherkati, S., 2014, Impact of the Late
 996 Triassic Dashtak intermediate detachment horizon on anticline geometry in the Central Frontal
 997 Fars, SE Zagros fold belt, Iran: *Marine and Petroleum Geology*, v. 54, p. 23–36,
 998 doi:10.1016/j.marpetgeo.2014.02.010.
- 999 Navabpour, P., Barrier, E., and McQuillan, H., 2014, Oblique oceanic opening and passive margin
 1000 irregularity, as inherited in the Zagros fold-and-thrust belt: *Terra Nova*, v. 26, p. 208–215,
 1001 doi:10.1111/ter.12088.
- 1002 Okada, H., 1980, Supplementary modification and introduction of code numbers to the low-latitude
 1003 coccolith biostratigraphy (Bukry, 1973; 1975): *Mar. Micropaleontol.*, v. 5, p. 321–325,
 1004 doi:10.1016/0377-8398(80)90016-X.
- 1005 Oveisi, B., Lavé, J., Van Der Beek, P., Carcaillet, J., Benedetti, L., and Aubourg, C., 2009, Thick-
 1006 and thin-skinned deformation rates in the central Zagros simple folded zone (Iran) indicated
 1007 by displacement of geomorphic surfaces: *Geophysical Journal International*, v. 176, p. 627–
 1008 654, doi:10.1111/j.1365-246X.2008.04002.x.
- 1009 Paul, A., Hatzfeld, D., Kaviani, A., Tatar, M., and Péquegnat, C., 2010, Seismic imaging of the
 1010 lithospheric structure of the Zagros mountain belt (Iran): *Geological Society, London, Special*
 1011 *Publications*, v. 330, p. 5–18, doi:10.1144/SP330.2.
- 1012 Perotti, C., Chiariotti, L., Bresciani, I., Cattaneo, L., and Toscani, G., 2016, Evolution and timing of
 1013 salt diapirism in the Iranian sector of the Persian Gulf: *Tectonophysics*, v. 679, p. 180–198,

doi:10.1016/j.tecto.2016.04.040.

Pirouz, M., 2018, Post-collisional deposits in the Zagros foreland basin: Implications for diachronous underthrusting: *International Journal of Earth Sciences*, v. 107, p. 1603–1621, doi:10.1007/s00531-017-1561-y.

Pirouz, M., Avouac, J.-P., Hassanzadeh, J., Kirschvink, J.L., and Bahroudi, A., 2017, Early Neogene foreland of the Zagros, implications for the initial closure of the Neo-Tethys and kinematics of crustal shortening: *Earth and Planetary Science Letters*, v. 477, p. 168–182, doi:10.1016/j.epsl.2017.07.046.

Pirouz, M., Simpson, G., Bahroudi, A., and Azhdari, A., 2011, Neogene sediments and modern depositional environments of the Zagros foreland basin system: *Geological Magazine*, v. 148, p. 838–853, doi:10.1017/S0016756811000392.

Pirouz, M., Simpson, G., and Chiaradia, M., 2015, Constraint on foreland basin migration in the Zagros mountain belt using Sr isotope stratigraphy: *Basin Research*, v. 27, p. 714–728, doi:10.1111/bre.12097.

Rahmani, Z., Vaziri-Moghaddam, H., and Taheri, A., 2010, Facies Distribution and Paleoecology of the Guri Member of the Mishan Formation in Lar area, Fars Province, SW Iran: *Iranian Journal of Science & Technology*, v. 34, p. 257–266.

Ricou, L.-E., 1971, Le metamorphisme au contact des peridotites de Neyriz (Zagros interne, Iran); developpement de skarns a pyroxene: *Bulletin de la Société Géologique de France*, v. 7, p. 146–155, doi:10.2113/gssgfbull.S7-XIII.1-2.146.

Ruh, J.B., 2017, Effect of fluid pressure distribution on the structural evolution of accretionary wedges: *Terra Nova*, v. 29, p. 202–210, doi:10.1111/ter.12263.

- 1036 Ruh, J.B., Gerya, T., and Burg, J.-P., 2013, High-resolution 3D numerical modeling of thrust
1037 wedges: Influence of décollement strength on transfer zones: *Geochemistry, Geophysics,*
1038 *Geosystems*, v. 14, p. 1131–1155, doi:10.1002/ggge.20085.
- 1039 Ruh, J.B., Hirt, A.M., Burg, J.-P., and Mohammadi, A., 2014, Forward propagation of the Zagros
1040 Simply Folded Belt constrained from magnetostratigraphy of growth strata: *Tectonics*, v. 33,
1041 p. 1534–1551, doi:10.1002/2013TC003465.
- 1042 Ruh, J.B., Kaus, B.J.P., and Burg, J.-P., 2012, Numerical investigation of deformation mechanics in
1043 fold-and-thrust belts: Influence of rheology of single and multiple décollements: *Tectonics*, v.
1044 31, doi:10.1029/2011TC003047.
- 1045 Ruh, J.B., Vergés, J., and Burg, J.-P., 2018, Shale-related minibasins atop a massive olistostrome in
1046 an active accretionary wedge setting: Two-dimensional numerical modeling applied to the
1047 Iranian Makran: *Geology*, v. 46, p. 791–794, doi:10.1130/G40316.1.
- 1048 Sadeghi, S., Storti, F., Yassaghi, A., Nestola, Y., and Cavoizzi, C., 2016, Experimental deformation
1049 partitioning in obliquely converging orogens with lateral variations of basal décollement
1050 rheology: Inferences for NW Zagros, Iran: *Tectonophysics*, v. 693, p. 223–238,
1051 doi:10.1016/j.tecto.2016.05.014.
- 1052 Saura, E. et al., 2011, Basin architecture and growth folding of the NW Zagros early foreland basin
1053 during the Late Cretaceous and early Tertiary: *Journal of the Geological Society*, v. 168, p.
1054 235–250, doi:10.1144/0016-76492010-092.
- 1055 Saura, E., Garcia-Castellanos, D., Casciello, E., Parravano, V., Urruela, A., and Vergés, J., 2015,
1056 Modeling the flexural evolution of the Amiran and Mesopotamian foreland basins of NW
1057 Zagros (Iran-Iraq): *Tectonics*, v. 34, p. 377–395, doi:10.1002/2014TC003660.
- 1058 Sepehr, M., and Cosgrove, J.W., 2004, Structural framework of the Zagros Fold-Thrust Belt, Iran:

1059 Marine and Petroleum Geology, v. 21, p. 829–843, doi:10.1016/j.marpetgeo.2003.07.006.

1060 Sherkati, S., Letouzey, J., and de Lamotte, D.F., 2006, Central Zagros fold-thrust belt (Iran): New
 1061 insights from seismic data, field observation, and sandbox modeling: *Tectonics*, v. 25, p. 1–27,
 1062 doi:10.1029/2004TC001766.

1063 Soleimany, B., Poblet, J., Bulnes, M., and Sàbat, F., 2011, Fold amplification history unravelled
 1064 from growth strata: the Dorood anticline, NW Persian Gulf: *Journal of the Geological Society*,
 1065 v. 168, p. 219–234, doi:10.1144/0016-76492010-085.

1066 Stewart, S.A., 2018, Hormuz salt distribution and influence on structural style in NE Saudi Arabia:
 1067 *Petroleum Geoscience*, v. 24, p. 143–158, doi:10.1144/petgeo2017-011.

1068 Talebian, M., and Jackson, J., 2004, A reappraisal of earthquake focal mechanisms and active
 1069 shortening in the Zagros mountains of Iran: *Geophysical Journal International*, v. 156, p. 506–
 1070 526, doi:10.1111/j.1365-246X.2004.02092.x.

1071 Tavakoli-Shirazi, S., Leturmy, P., de Lamotte, D., and Ringenbach, J.-C., 2013, Structural style and
 1072 kinematic evolution of the high Zagros, Iran: *Journal of Petroleum Geology*, v. 36, p. 309–
 1073 334, doi:10.1111/jpg.12559.

1074 Tavani, S., Camanni, G., Nappo, M., Snidero, M., Ascione, A., Valente, E., Gharabeigli, G.,
 1075 Morsalnejad, D., and Mazzoli, S., 2020, The Mountain Front Flexure in the Lurestan region of
 1076 the Zagros belt: Crustal architecture and role of structural inheritances: *Journal of Structural*
 1077 *Geology*, v. 135, p. 104022, doi:<https://doi.org/10.1016/j.jsg.2020.104022>.

1078 Teknik, V., and Ghods, A., 2017, Depth of magnetic basement in Iran based on fractal spectral
 1079 analysis of aeromagnetic data: *Geophysical Journal International*, v. 209, p. 1878–1891,
 1080 <http://dx.doi.org/10.1093/gji/ggx132>.

1081 Tunini, L., Jiménez-Munt, I., Fernández, M., Vergés, J., and Villaseñor, A., 2015, Lithospheric
1082 mantle heterogeneities below the Zagros Mountains and the Iranian Plateau: a petrological-
1083 geophysical study: *Geophysical Journal International*, v. 200, p. 596–614,
1084 doi:10.1093/gji/ggu418.

1085 Vergés, J., 2007, Drainage Responses to Oblique and Lateral Thrust Ramps: A Review (G. Nichols,
1086 C. Paola, & E. Williams, Eds.): *Sedimentary Processes, Environments and Basins*, v. 38, p.
1087 29–47, doi:10.1002/9781444304411.ch3.

1088 Vergés, J., Emami, H., Garcés, M., Beamud, E., Homke, S., and Skott, P., 2019, Zagros Foreland
1089 Fold Belt Timing Across Lurestan to Constrain Arabia–Iran Collision, *in* *Developments in*
1090 *Structural Geology and Tectonics*, Elsevier, v. 3, p. 29–52, doi:10.1016/B978-0-12-815048-
1091 1.00003-2.

1092 Vergés, J., Goodarzi, M.G.H.M., Emami, H., Karpuz, R., Efstathiou, J., and Gillespie, P., 2011a,
1093 Multiple detachment folding in Pusht-e Kuh arc, Zagros: Role of mechanical stratigraphy:
1094 Thrust fault-related folding: *AAPG Memoir*, v. Chapter 4, p. 69–94,
1095 doi:10.1306/13251333M942899.

1096 Vergés, J., Saura, E., Casciello, E., Fernández, M., Villaseñor, A., Jiménez-Munt, I., and García-
1097 Castellanos, D., 2011b, Crustal-scale cross-sections across the NW Zagros belt: implications
1098 for the Arabian margin reconstruction: *Geological Magazine*, v. 148, p. 739–761,
1099 doi:10.1017/S0016756811000331.

1100 Wrobel-Daveau, J.-C., 2011, From rifting to current collision, vertical movement and propagation
1101 of deformation in the Zagros belt, Iran: insights from section balancing and detrital low-
1102 temperature thermochronology [Ph.D. thesis]: Université de Cergy-Pontoise, 200 p.

1103 Yamato, P., Kaus, B.J.P., Mouthereau, F., and Castelltort, S., 2011, Dynamic constraints on the

1104 crustal-scale rheology of the Zagros fold belt, Iran: *Geology*, v. 39, p. 815–818,
1105 doi:10.1130/G32136.1.

1106 Zijderfeld, J.D.A., 1967, AC demagnetization of rocks: analysis of results, *in* *Developments in*
1107 *solid earth geophysics*, Elsevier, v. 3, p. 254–286, doi:10.1016/B978-1-4832-2894-5.50049-5.

1108 FIGURE CAPTIONS

1109 Figure 1. A) Location of the Zagros foreland Fold Belt on the NE margin of the Arabian plate, B)
1110 Map of the Zagros tectonic domains and major basement faults. The current study area is located in
1111 the Dowlatabad growth syncline in the front of the Fars arc. The age of folding initiation of
1112 previously studied growth synclines are indicated on the map, including Shalamzar, Dehmoord
1113 (Pirouz et al., 2017), Chaman-Goli, Afrineh (Emami, 2008; Vergés et al., 2019), Changleh
1114 (Homke et al., 2004), Jarik (Lashgari et al., 2020), Qalat (Khadivi et al., 2010), and Ghol Ghol
1115 (Ruh et al., 2014) synclines. These ages were determined based on the magnetostratigraphy of
1116 growth strata.

1117 Figure 2. Regional structural cross-section extending from the Main Zagros Thrust to Saudi Arabia,
1118 across the High Zagros, Fars arc and Persian Gulf. The onshore segment closely follows the cross-
1119 sections by McQuarrie (2004) and Mouthereau et al. (2007). The depth of Moho is based on
1120 geophysical studies and numerical modeling (Paul et al., 2010; Jiménez-Munt et al., 2012; Tunini et
1121 al., 2015; Dashti et al., 2019).

1122 Figure 3. A) Geological map of the Dowlatabad growth syncline neighbored by the Sefid anticline
1123 to the north and the NW Pazan anticline to the south (modified from 1:100,000 scale geological
1124 map series provided by NIOC). Location of paleomagnetic, strontium and calcareous nannofossil
1125 sampling sites are indicated on the map. B) CD geological cross-section constructed by projecting
1126 pre-, syn- and late-growth strata patterns across the Dowlatabad syncline and the NW Pazan
1127 anticline.

1128 Figure 4. A) Oblique satellite view and B) schematic evolution model of the Dowlatabad syncline
1129 with indication of the unconformities, growth strata units and sediment provenance. Unconformity
1130 U1 characterizes the base of growth strata, which define a gradual thinning of middle and upper
1131 Aghajari units from 1700 m to 500 m towards the NW, and the onset of folding in the Pazan
1132 anticline. The U2 is a local intra-growth angular unconformity at the base of first appearance of
1133 Guri Mb carbonate clasts derived from the dismantled crest of the Pazan anticline. The U3
1134 represents a syncline-scale unconformity at the base of the late-growth gently dipping Bakhtyari
1135 conglomerate layers mostly composed of volcanic and metamorphic clasts from hinterland.

1136 Figure 5. A) Stratigraphy of the Miocene–Pleistocene foreland deposits in the Dowlatabad growth
1137 syncline. B) Very thick-bedded to massive carbonates of the Guri Mb. C) Flute cast paleocurrent
1138 indicators in medium to coarse grained sandstones and mudrocks of the Aghajari Fm. D)
1139 Monomictic carbonate conglomerates and pebbly sandstone of the topmost part of the Aghajari
1140 Fm., with sub-angular clasts from the Guri Mb. E) Polymictic well-rounded clast-supported
1141 conglomerates of the Bakhtyari Fm., showing imbrication and containing sandstone channels.

1142 Figure 6. A) Demagnetization diagrams of representative samples for the Mishan, Aghajari and
1143 Bakhtyari formations showing normal (DW203-1A, DW130-1A) and reversed (DW211-1A,
1144 DW136-1A) polarities. In red, the characteristic component (ChComp) with indication of the
1145 selected range of temperatures used to calculate it. BTC: before tectonic correction; ATC: after
1146 tectonic correction. B) Equal-area stereographic projection of the characteristic components
1147 (quality 1 and 2) and the obtained mean directions before (BTC) and after (ATC) tectonic
1148 correction with the 95% confidence angle. N: number of directions; dec: declination; inc:
1149 inclination; k: precision parameter; α_{95} : confidence angle; rsum: module of the sum vector for the
1150 N unitary vector directions. It can be observed that directions are better group ATC and that they do
1151 not reveal any vertical axis rotation.

1152 Figure 7. Stratigraphic log of the studied section with the location of magnetostratigraphic samples.
1153 Measured paleocurrent directions and changes in clast composition are also indicated. In addition,
1154 declination, inclination and latitude of the virtual geomagnetic pole, local magnetostratigraphy and
1155 the proposed correlation to the Global Polarity Time Scale (GPTS) (Gradstein and Ogg, 2012) are
1156 presented respectively from left to right. Circles represent class 1 and 2 samples and white squares
1157 are used for class 3 samples. Half bars show single site magnetozones, not considered as reliable.

1158 Figure 8. Integration of magnetostratigraphy, Sr isotopic and calcareous nannofossil ages for about
1159 3.5 km of foreland sediments in the Dowlatabad section. Local magnetostratigraphy for the topmost
1160 Mishan, Aghajari and Bakhtyari formations is correlated to the GPTS2012 (Gradstein and Ogg,
1161 2012). Sr isotopic ages for the topmost Gachsaran and Mishan formations are plotted on the
1162 LOWESS curve of McArthur et al. (2001). The mean values, illustrated by small circles and error
1163 ranges, are shown by bars. The vertical bars show uncertainty of Sr isotopic ratios and horizontal
1164 bars indicate the age uncertainty for each sample. Calcareous nannofossils are assigned to their
1165 specific biozones. Using the age ties, rate of sediment accumulation is calculated for each
1166 depositional interval.

1167 Figure 9. Setup of the initial stage of numerical models. From bottom to top, each model consists of
1168 25 km of diabase basement, 2 km of low-viscous detachment (equivalent to Hormuz salt), and 10
1169 km of upper sedimentary crust with two weak interlayers. All model parameters are given in Table
1170 S1. (A) shows a thin-skinned model without inherited basement faults, while (B) represents a thick-
1171 skinned model with listric inherited basement faults.

1172 Figure 10. Temporal evolution of numerical experiments without (A–C) and with listric inherited
1173 basement faults (D–F). Upper panels: Composition of rock markers. See Fig. 9 for color code.
1174 Lower panels: Second invariant of the strain-rate tensor. Blue colors indicate fast deformation,
1175 white denotes slow deformation.

1176 Figure 11. A) Revised Miocene–Pleistocene foreland chronostratigraphy for the SE Zagros (Fars
1177 arc) constructed by synthesizing the results of this study with previously published
1178 magnetostratigraphic sections (Khadivi et al., 2010; Ruh et al., 2014; Pirouz et al., 2017).
1179 Sedimentation rates (cm/kyr) for the different sections in SE Zagros with indication of time of
1180 growth initiation are also represented below the chronostratigraphic chart. B) Foreland
1181 chronostratigraphy for the NW Zagros (Lurestan arc) based on the previously published
1182 magnetostratigraphic ages (Homke et al., 2004; Pirouz et al., 2017; Vergés et al., 2019). Beginning
1183 time of deformation according to the base of growth strata and propagation rates of deformation
1184 front (pr) are represented on the chronostratigraphic charts (A, B).

1185 Figure 12. Deformation propagation patterns obtained from numerical experiments. A) Experiment
1186 without pre-existing basement faults. B) Experiment with two listric pre-existing basement faults.
1187 Upper panels: Composition of rock markers after 15 Myr of shortening. See Fig. 9 for color code.
1188 Lower panels: Temporal evolution of horizontal shortening rate within the cover sequence.

1189 Figure 13. Map depicting the foreland propagation of deformation across the Zagros Fold Belt
1190 during early Miocene to Pleistocene (modified from Vergés et al., 2019). In addition, onset of
1191 folding in Persian Gulf is indicated for the Hormuz salt diapirs and the N-S Arabian folds. The map
1192 is defined based on well-dated growth synclines including Shalamzar, Dehmoord (Pirouz et al.,
1193 2017), Chaman-Goli, Afrineh (Emami, 2008; Vergés et al., 2019), Changuleh (Homke et al., 2004),
1194 Jarik (Lashgari et al., 2020) Qalat (Khadivi et al., 2010), Ghol Ghol (Ruh et al., 2014), and
1195 Dowlatabad (this study) synclines. The age of the folding onset is marked for each growth syncline.
1196 The out-of-sequence reactivation of the Mountain Front Fault is highlighted. Locations of two
1197 foreland chronostratigraphic panels of Fig. 11 are illustrated on the map.

Fig. 1

Fig. 2

Fig. 3

A) Geological map of the Dowlatabad growth syncline

B) Cross-section of the Dowlatabad syncline

Fig. 4

B

a) onset of Pazan anticline growth

S - N

b) onset of growth strata

c) Guri limestone erosion

d) late-growth strata

Fig. 5

Magnetostratigraphy (Fig. 7)

Fig. 6

Fig. 7

Fig. 8

Fig. 9

A) No inherited faults

B) Deep-seated listric inherited faults

Continental basement
Inherited basement faults

Hormuz salt horizon
Sedimentary cover sequence

Intermediate evaporite horizons
Syn-tectonic deposits

Fig. 10

Fig. 11

A) SE Zagros (Fars arc)

Persian Gulf Mountain Front Fault Frontal Fars arc Sumreh Fault Interior Fars arc High Zagros Fault High Zagros Main Zagros Thrust Sanandaj-Sirjan

SW NE

Deformation front Dowlatabad This study Ghol Ghol Ruh et al. (2014) Qalat Khadivi et al. (2010) Dehmoord Pirouz et al. (2017)

Ma 0 2 4 6 8 10 12 14 16 18 20 22

pr: 11 mm/yr 1.8 4.6 4.9 3.8 3.2 6 14 14.8 16.6 16.5 18.8 20 ~18

progradation aggradation progradation

Mishan Fm. Bakhtyari Fm. Aghajari Fm. Razak Fm.

Base of growth strata (onset of folding) Base of Bakhtyari Fm. pr: propagation rate of deformation front Base of Aghajari Fm. Base of Mishan Fm.

300 250 200 150 100 50 0 km

Bakhtyari Fm. Aghajari Fm. Mishan Fm. Razak Fm.

end of growth Frontal Fars Dowlatabad syncline This study

growth

High Zagros Dehmoord syncline Pirouz et al. (2017)

growth

Interior Fars Ghol Ghol syncline Ruh et al. (2013)

growth

Interior Fars Qalat syncline Khadivi et al. (2010)

0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20 21 22 23 Ma

C1 C2 C2A C3 C3A C3B C4 C4A C5 C5A C5AA C5AB C5A C C5A D C5B C5C C5D C5E C6 C6A C6AA C6B C6C

H. Pleistocene Pliocene Miocene

B) NW Zagros (Lurestan arc)

Dezful-Kirkuk Embayment

Mountain Front Fault

Lurestan arc

High Zagros Fault

Main Zagros Thrust

SW

NE

Deformation front

Changuleh
Homke et al. (2004)

Afrineh
Vergés et al. (2019)

Chaman-Goli
Vergés et al. (2019)

Shalamzar
Pirouz et al. (2017)

Ma

0

2

4

6

8

10

12

14

16

18

20

22

pr: 9 mm/yr

pr: 20 mm/yr

progradation

Aghajari Fm.

Bakhtyari Fm.

pr: 55 mm/yr

aggradation/
progradation

pr: 7 mm/yr

Gachsaran Fm.

Razak Fm.

250

200

150

100

50

0 km

C1

C2

C2A

C3

C3A

C3B

C4

C4A

C5

C5A

C5B

C5C

C5D

C5E

C5F

C5G

C5H

C5I

C5J

C5K

C5L

C5M

C5N

C5O

C5P

C5Q

C5R

C5S

C5T

C5U

C5V

C5W

C5X

C5Y

C5Z

C6

C6A

C6B

C6C

Fig. 12

A) No inherited basement faults

B) Deep-seated inherited listric faults

Fig. 13