The role of folding in the development of the Mexican fold-and-thrust belt

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ABSTRACT

The Mexican fold-and-thrust belt in central Mexico has overall characteristics that fit the critical tectonic wedge model. It is thin-skinned, forward propagating, tapers toward the toe (the east), and displays an overall decrease in deformation toward the toe. The internal structures and heterogeneity of deformation are not typical of fold-and-thrust belts, however, due to the presence of two large carbonate platforms, flanked by more thinly bedded basinal carbonates. Kilometer-scale thrusts dominate deformation in the platform carbonates (a more brittle behavior), and mesoscopic buckle folds and associated cleavage dominate deformation in the basinal carbonates (a more ductile behavior). Total shortening across the belt, including both platforms and basins, is ~55%–65%, with higher values in the basins than in the platforms and a concentration of deformation near the platform borders. The dominant mechanism of folding in the basinal rocks is buckling, with thin chert horizons behaving as single layers and limestone and shaly limestone interbeds buckling as multilayers, with a dominant chevron style. A significant shear component of the deformation is indicated by monoclinic fold symmetry, with a consistent sense of vergence of top toward the foreland. We estimated strain and strain history from mesoscopic analysis of fold geometry and internal strain distribution at several locations across the basin and used this information used to assess the overall kinematics and progress-

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INTRODUCTION

Fold-and-thrust belts are one of the most common tectonic features worldwide, and their basic characteristics, tectonic style, and kinematics have been established by study of natural examples (e.g., Armstrong, 1968; Price and Fermor, 1985; Philippe et al., 1996; McQuarrie, 2004) and by numerical and physical modeling (e.g., Huiqi et al., 1992; Dixon, 2004; Stockmal et al., 2007; Simpson, 2009). Excellent descriptions of the structural characteristics of fold-and-thrust belts were provided by Dahlstrom (1969) and Boyer and Elliott (1982). Key features are the overall forward propagation of deformation, piggyback style, the presence of a basal décollement, and decreasing intensity of deformation toward the foreland. Most fold-and-thrust belts are wedge shaped, and since the mid-1980s, the critical taper theory of their development (Davis et al., 1983; Dahlen et al., 1984; Dahlen, 1990) has become generally accepted and is supported by field observations and by numerical and physical models. Although increasingly sophisticated numerical codes allow many features of fold-and-thrust belts to be simulated and the effects of key controlling parameters to be assessed, they do not fully explain the variations of deformation styles along a cross section at multiple scales of observation. The results of numerical simulations, in conjunction with theory, do, however, provide an excellent guide when studying natural examples, and study of natural examples in turn provides information that allows models and theory to be refined. When combined iteratively, all these approaches lead to a better understanding of deformational processes within fold-and-thrust belts.

The Mexican fold-and-thrust belt, like virtually all fold-and-thrust belts, is wedge shaped and thin skinned. It has been the subject of a number of regional studies (Campa-Uranga, 1983; Padilla y Sánchez, 1986; Eguiluz et al., 2000). We present here the results of a structural analysis along a cross section of the Mexican fold-and-thrust belt in central Mexico, for which the regional characteristics have been previously documented in classical papers by Suter (1987) and Carrillo-Martínez et al. (2001). In the present work, we attempt to relate local deformation on the mesoscopic scale to the overall deformation of the fold-and-thrust belt. In an earlier work (Fitz-Díaz et al., 2011), we focused our attention on the large-scale deformation and an interpretation of the structures in terms of critical taper wedge theory. In that paper, we also compared and contrasted the geometrical, lithological, and tectonic features of the Mexican fold-and-thrust belt with those of the classic cross section of the southern Canadian Rocky Mountains.

We present here new detailed qualitative and quantitative information from a study of small-scale buckle folds developed in carbonates in two basins traversed by the studied cross section. The geometrical analysis of these folds allows us to locally estimate the strain and its variation along the cross section. Additionally, a careful qualitative analysis of fold-related veins, axial planar cleavage, and fold geometry provides clues to the kinematics of folding and more broadly the evolution of the belt.

We use data on fold geometry and strain, based on observations at key outcrops, to relate mesoscopic-scale folding to deformation at the scale of the basins in a kinematic model. This, together with information on deformation in the platforms, allows the effects of lithological variations within the wedge as a whole to be assessed.

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REGIONAL FRAMEWORK

Subduction along the western margin of North America and associated accretion of terranes swept from reaches of the Pacific Ocean are considered to be responsible for the formation of the Cordilleran orogen (Coney et al., 1980; Coney and Evenchick, 1994; Armstrong, 1974; DeCelles, 2009). One of the consequences of such accretory or collisional processes is the shortening of sedimentary rocks in the foreland, giving rise to the formation of fold-and-thrust belts, a prime example of which is the Rocky Mountains fold-and-thrust belt in western North America. Although the Rocky Mountains fold-and-thrust belt is relatively narrow and has a thin-skinned deformation style, with a width between 100 and 200 km. In the southwest United States, it widens to more than 1000 km and is divided into two tectonic elements: the Sevier and Laramide fold-and-thrust belts (Fig. 1; Armstrong, 1968, 1974). The Sevier fold-and-thrust belt is thin skinned, and its front is to the west of the Colorado Plateau (Armstrong, 1968; Yonkee, 1992). The Laramide province extends eastward of the Sevier fold-and-thrust belt into the continental interior from central Montana to northern Mexico. It is characterized by basement-involved (thick-skinned) structures (Allmendinger et al., 1982; Brewer et al., 1982; Schmidt et al., 1993).

It is not clear how the Sevier and Laramide tectonic fronts continue to the south. Nevertheless, in a broad area in northern Mexico, low-angle thrusts and related folds are cut by large-scale high-angle reverse faults that involve blocks of basement (e.g., San Marcos and La Babia faults; Chávez-Cabello et al., 2007).

Further south, the Mexican fold-and-thrust belt is expressed topographically in the Sierra Madre Oriental. It is thin skinned and 100–250 km wide (Campa-Uranga, 1983; Eguiluz et al., 2000), extending from northern Mexico, where it is oriented NW–SE and its northern boundary is parallel with the Mojave-Sonora megashear trace (Fig. 1; Anderson and Schmidt, 1983; Anderson and Silver, 2005), to central and southern Mexico. At its southern end, this belt narrows, changes orientation to N–S, and finally bifurcates, with one branch that continues along the Gulf of Mexico coast and another that runs through the Morelos Guerrero platform (Fig. 1).

This study is focused on a well-exposed cross section of the Mexican fold-and-thrust belt in central Mexico (Fig. 1). The chosen transect is oriented ENE–WSW, and in only 150 km, we can see a full cross section of the belt, with a relief of ~3000 m. In this area, the Mexican fold-and-thrust belt shows a dominant thin-skinned deformation style, with an overall wedge shape tapering toward the east.

**Stratigraphy**

The stratigraphy of the area was first described in the 1940s (Inlay, 1944) and has been subsequently refined in a number of papers (Segerstrom, 1961; Suter 1980, 1984, 1990; Carrillo-Martínez, 1989; Wilson and Ward, 1993; Carrillo-Martínez et al., 2001, Dávila-Alcocer et al., 2009). The rocks are mostly Cretaceous carbonates, with lateral facies changes associated with different paleogeographical elements across the region. These elements are, from east to west: the Tampico-Misantla Basin, the Valles–San Luis Potosí Platform, the Zimapán Basin, and the El Doctor Platform (Carrillo-Martínez, 1997; Suter, 1987, 1990). To the west, the Cretaceous rocks are in fault contact with a suite of rocks here grouped as the Tolimán sequences (Figs. 2 and 3). The paleogeographic elements and the stratigraphy of the area play a major role in the characteristics of the deformation, as we will demonstrate in the following sections.

In Figure 2, a characteristic stratigraphic column is presented for each area. Each column includes the regional stratigraphic nomenclature, graphical representation of lithology for the different units, age, a rough estimate of thickness and the shortening events observed. On the eastern side of the transect, the Tampico-Misanta Basin is characterized by a sedimentary record that starts in the Triassic(?)-Jurassic with continental sandstones and conglomerates (Huizachal and Cahuasas Formations) unconformably deposited on a basement of Grenville-age granulate gneiss. Above these units, a sea-level rise caused a marine transgression (Ochoa-Camarillo, 1996) and resulted in the deposition of calcareites and carbonaceous shale (Santiago Formation) with a strong influence of continental organic matter. Reactivation of normal faults resulted in horsts and graben structures, which gave rise to an uneven distribution of deep- and shallow-water facies (Ochoa-Camarillo, 1996). These rocks in turn were covered in the Early Cretaceous by thinly bedded black mudstone with abundant chert bands.
Figure 2. Chart synthesizing stratigraphic and lithological variations of units along the studied cross section, with characteristic columns for each major paleogeographic element. Each column displays the characteristic lithology of the different stratigraphic units, with their formational names, age constraints, and estimated thicknesses (t—thickness). Also indicated are the units that show field evidence for having been affected by pre-D1, D1, or D2 deformations. This stratigraphical information is a summary from a number of works (Imlay, 1944; Segerstrom, 1961; Suter, 1980, 1984, 1990; Carrillo-Martínez, 1989; Carrillo-Martínez et al., 2001; Hernández-Jaüregui, 1997; Dávila-Alcocer et al., 2009) and new age constraints from detrital zircon dating. Thickness estimates come from borehole records and field observations. The gray dashed lines correlate units deposited between the Berriasian and the upper limit of the Cenomanian. Notice the abrupt facies changes in the different paleogeographic elements for this time range. SLP—San Luis Potosí.
Figure 3. (A) Simplified geologic map of the study area in central Mexico, showing the location of cross-section A–A'. The location of this area in the North American Cordillera is shown in Figure 1. (B) Admissible and partly schematic cross section of the Mexican fold-and-thrust belt (cross-section A–A') showing the variation of deformational style within the wedge. (C) Restored cross-section A–A' showing the distributions of carbonate facies prior to deformation. Figure parts are modified from Fitz-Díaz et al. (2011). SLP—San Luis Potosí.
and facies (e.g., platform and basinal carbon-ates) are indicated using the same color. Figure 3C is a reconstruction in cross section of the facies distribution prior to deformation.

Column A of Figure 2 represents the Tolimán sequences described in detail by Dávila-Alcocer et al. (2009). In this area, the oldest exposed rocks are lower-greenschist-facies phyllites and schists of the El Chilar Complex, which exhibit block in matrix mélangé structures and phacoidal cleavage. This unit is unconformably overlain by the San Juan de la Rosa Formation (sensu Dávila-Alcocer et al., 2009), a volcaniclastic unit of Upper Jurassic–Lower Cretaceous age that grades upward to the Peña Azul Formation, which consists of thinly bedded marl and shale. Further up the section, there are thinly bedded carbonates of the Upper Tamaulipas Formation covered unconformably by the Soyatal Formation. The lower parts of the section have been attributed by Chauve et al. (1985) to the “Cordilleran domain” of the Sierra Madre Oriental.

Although basement rocks are not exposed along the cross section, they have been reported in borehole records (Eguiluz et al., 2000) along the eastern edge of the Tampico-Misantla Basin (and are so interpreted in the cross section of Fig. 3B). Some blocks of the basement have been dragged to the surface along major thrusts and are exposed a few kilometers to the south of the area (Suter, 1990). We infer the presence of Jurassic red beds under the carbonates of the Zimapán Basin from borehole records to the south of the area (Eguiluz et al., 2000). It is also likely that Permian–Triassic volcaniclastic rocks underlie the rocks of the Zimapán Basin and El Doctor Platform, based on the field observations by Rosales-Lagarde et al. (2005) to the south of the studied area. According to these authors, these units were involved in the deformation.

Geometry of the Wedge

A tapered orogenic wedge model was originally proposed by Chapple (1978), and the first robust wedge theory was developed by Davis et al. (1983), who demonstrated the applicability of the approach to submarine fold-and-thrust wedges. Improvements and refinements to this model have been made since then (Dahlen et al., 1984; Dahlen, 1990). In this model, the cited authors assumed a Mohr-Coulomb behavior of a homogeneous material within the wedge, and a mechanically weaker material along the basal detachment. Dahlen et al. (1984) demonstrated that the geometry of a subaerial wedge (given by the critical taper angle, which is the sum of the topographic slope angle and the dip of the detachment) is characteristic for a particular combination of: strength of the rock within the wedge (given by the coefficient of internal friction); frictional resistance along the detachment; and pore fluid pressure ratio (the ratio of hydrostatic and lithostatic pressures) within the wedge and along the detachment.

On the regional scale, the Mexican fold-and-thrust belt displays a wedge shape typical of thin-skinned fold-and-thrust belts (Fig. 3A). This wedge tapers toward the east, in the same direction that the topography decreases and the detachment fault ascends to the surface. In a

Figure 4. (A) Calcarenite and shale of the Soyatal Formation in the footwall of the Higuerillas thrust, showing complex deformation. Note the tight folds, with an associated axial planar cleavage (S1), refolded and affected by later thrust faults. (B) Calcarenite, siltstone, and shale of the Méndez Formation in the foothills of the Mexican fold-and-thrust belt, showing tight folds and a pervasive axial planar cleavage (S1), which, in some places, is cut by low-angle faults. (C) Calcarenite, siltstone, and shale of the Velasco Formation in the front of the Mexican fold-and-thrust belt, affected by a low-angle thrust fault that accommodates meter-scale displacement. Notice the absence of cleavage in the zoomed image.
previous study, an average critical taper angle between 5° and 6.3° was calculated for the studied cross section (for details, see Fitz-Díaz et al., 2011). This is consistent with taper angles in cross sections of the Mexican fold-and-thrust belt to the north and to the south of the area (Eguíluz et al., 2000), which have borehole records and seismic profiles to provide control at depth. The limited stratigraphic range and thickness of strata of the rocks exposed across the full length of the cross section are consistent with a relatively thin wedge.

The lateral facies changes along the cross section (as seen on the stratigraphic columns of the different paleogeographical elements; see Fig. 2) can be expected to have locally modified the critical taper angle or, in other words, controlled whether some segments favored displacement over internal shortening. As an example, there is a marked difference in the amount of shortening experienced by the Valles–San Luis Potosí Platform and the adjacent Tampico-Misantla Basin (Fig. 3). This difference in behavior can be explained by the difference in internal strength of the units above the detachment and a difference in frictional resistance along the detachment, both of which favored slip over shortening in the Valles–San Luis Potosí Platform and shortening over slip in the Tampico-Misantla Basin. The massive dolomitized limestone of the Valles–San Luis Potosí Platform was stronger than the well-bedded carbonaceous mudstone interbedded with shale and chert of the Tampico-Misantla Basin, and the frictional resistance of evaporites under the Valles–San Luis Potosí Platform would have been less than the frictional resistance of the carbonaceous shale under the Tampico-Misantla Basin. A difference in taper angle—lower associated with the Valles–San Luis Potosí Platform and higher associated with the Tampico-Misantla Basin—is linked to the difference in the relative importance of basal slip versus internal shortening (Fitz-Díaz et al., 2011).

In the following sections, the characteristics of deformation will be described, both qualitatively and quantitatively, for each paleogeographical element. These data will allow us to test the predictions we can make by applying the orogenic wedge model.

DEFORMATION WITHIN THE WEDGE

The large-scale structures within the Mexican fold-and-thrust belt wedge are shown in the cross section of Figure 3B. This cross section is an admissible and partially schematic representation of structures at the kilometer scale. In the following paragraphs, details of the structures on a smaller scale are described, including geometry, kinematics, and mechanisms of deformation and their variations along the cross section and among the different stratigraphic units.

On the western side of the study area, El Chilar Complex rocks exhibit a complex pattern of deformation (Figs. 2A, 3A, and 3B). This complexity is the result of at least three events of shortening, including: (1) an old anastomosing, block-in-matrix fabric, referred to here as pre-D1, which occurred under low-grade metamorphic conditions, given that tectonic chlorite and sericite developed on the foliation; (2) tight folding of the pre-D1 fabric, with a well-developed axial planar cleavage (S1); the folding and the cleavage are associated with D1; and (3) open folding of the D1 folds related to low-angle thrust faults; these structures are designated as D2. Pre-D1 structures seem to predate any deformation related to the Mexican fold-and-thrust belt because they do not affect the Cretaceous units that uncomfortably overlie the El Chilar Complex. D1 and D2 structures do, however, affect the Early to Late Cretaceous rocks lying on the El Chilar Complex. These structures are represented by tight folds (D1) refolded by more open folds (D2), and both fold episodes are associated with an axial planar cleavage (S1 and S2). S1 cleavage is folded by D2 to form a spaced S2 crenulation cleavage. Also associated with D2 are younger, low-angle thrusts faults, which cut the early folds.

D1 and D2 structures are developed in rocks across the Mexican fold-and-thrust belt, as indicated on the left-hand side of the column on the stratigraphic charts (Fig. 2). D1, as represented by folds and thrusts, seems to have been the main tectonic event that built up the Sierra Madre Oriental and, across the whole belt, accommodated considerable amounts of shortening (several hundred kilometers; Fig. 3B). Based on stratigraphic constraints, this event seems to have taken place from Turonian to Maastrichtian times in the studied region (see Fig. 2). D2, on the other hand, is characterized by gentle folds and meter-scale displacement on late thrust faults. It affected most of the Cretaceous strata in the area, and it is the only deformation that affected the Early Cenozoic clastic deposits in the foothills. Evidence for this last statement comes from: (1) the complex deformation in the Late Cretaceous clastic deposits (Soyatal Formation; Fig. 4A) on the western side of the cross section; and (2) the unconformity that separates Late Cretaceous clastic deposits (Méndez Formation; Fig. 4B), with D1 folds and a strong S1 axial planar cleavage, from Early Cenozoic clastic deposits (Velasco Formation; Fig. 4C), in which S1 is absent. Both D1 and D2 structures have a vergence toward the NE throughout the Mexican fold-and-thrust belt. While it is possible to consider D2 simply a later stage of D1, we consider it to be a distinct deformation event because the S2 structures cut S1 at a high angle.

The main focus of this study is on the D1 structures, since we consider that D1 contributed the most to the development of the Mexican fold-and-thrust belt. The small amount of shortening and weak development of folds during D2 make it possible to find D1 structures almost in their original position. A large amount of structural orientation data was collected along the cross section, and this is presented in the equal-area projections of Figure 5. All these data are consistent with a general direction of transport and shortening toward the NE along the entire cross section, with local deflections (on the order of ten degrees) of the structures in the different paleogeographic elements. The data also indicate that there is reactivation of axial planar cleavage associated with the different lithologies within the Zimapán and Tampico-Misantla Basins.

Along the section line (A–A’ in Fig. 3), the style and intensity of deformation vary depending on the position along the section and on lithology, which is characterized by the carbonates of the two platforms and the lateral facies changes from the platforms into the two basins (Figs. 2 and 3). In the following paragraphs, the characteristics of the deformation will be summarized for the platforms and the basins and the boundaries between them.

Deformation in the Platforms

The deformation in the platforms is dominantly brittle and thrust dominated and can only be considered penetrative at a kilometer scale (Figs. 6A, 6B, and 6C). The folds in the thick platform carbonates are very wide and are controlled by the geometry of the thrusts; they are mostly fault-bend folds (Suppe, 1983), although in relatively thin layers, fault-propagation folds (Suppe, 1983) can also be observed. Except for strong grain-scale deformation concentrated along thrust zones, in the interior of the thrust sheets the deformation is negligible at a metric scale and is accommodated by scarce veins and well-spaced stylolites. To represent these structures in the regional cross section, we took into account structural data and observations at the surface (e.g., Fig. 6A), borehole records from the surrounding areas, and the typical geometric relationships between thrusts and fault-related folds and the inferred wedge-shaped geometry of fold-and-thrust belts (Fitz-Díaz et al., 2011), as well as the basic rules for cross-section balancing.
Role of folding in the development of the Mexican fold-and-thrust belt

Figure 5. Equal-area lower-hemisphere projections of structural data, measured in different units, within the different paleogeographical elements (see Fig. 3 for reference). Note the deflection in orientation of the beta axis of folds and S1 in the Zimapán Basin (ZB) and Tampico-Misantla Basin (TMB).
Stratigraphic studies and structural observations in the area (Segerstrom, 1961; Hernández-Jáuregui, 1997) indicate that the original thickness of El Doctor Platform (between 800 and 1200 m) is triplicated by thrusting for at least half of its width. In a similar manner, the thickness of the Valles–San Luis Potosí Platform was partially duplicated, and relatively less shortened than El Doctor Platform. Estimates of shortening, determined by three different methods, are ~55% for El Doctor Platform and ~35% for the Valles–San Luis Potosí Platform (Fitz-Díaz et al., 2011). The latter is somewhat less than the value of 42% reported by Suter (1987) and Contreras and Suter (1990) for exactly the same cross section. The difference reflects the conservative way in which we constructed the cross section.

In both platforms, the average strike of the major thrusts is roughly 150°. The slickenlines measured on these surfaces indicate a direction of transport toward 050°–060° (Figs. 5A and 5B). With the exception of a few small-scale back thrusts, no major structures with reverse vergence were observed in the study area.

Deformation in the Basins

Deformation and associated shortening in the basins are far higher than in the platforms and were accommodated mostly by meter-scale buckle folds unrelated to faults. In the Zimapán Basin, the folds exhibit a rather ductile style of deformation, as evidenced by a considerable thickening and thinning of the layers in the hinge and in the limbs, respectively (Fig. 7), indicating penetrative deformation at the grain scale. In the Tampico-Misantla Basin, however, fold style indicates less ductile behavior, as evidenced by smaller thickness variations and the presence of small faults and fractures in association with folds (Fig. 8).

In the Zimapán Basin, most folds are tight to isoclinal, and their geometry is strongly controlled by the lithology (Fig. 7). In clastic carbonates (e.g., Soyatal and Trancas Formations), folds are relatively open with rounded hinges and with a strong axial planar cleavage. They are typically class 1C and occasionally class 2 of Ramsay (1967). Limestone layers interbedded with chert and shale (Tamaulipas Formation) typically show tight to isoclinal folds with angular hinges (chevron style), and less commonly are associated with a pressure-solution cleavage. Fold styles are fairly homogeneous within stratigraphic horizons across the basin. Most of the folds in the Zimapán Basin are asymmetrical with an axial planar cleavage striking NW-SE and dipping ~60° to the SW (Figs. 5E–5J). Nevertheless, along stratigraphic transitions, the axial...

Figure 6. Kilometer-scale view of structures in the study area. (A–B) Picture and sketch of the contact between El Doctor Platform and Zimapán Basin along the northern margin of the Moctezuma River. (C) Sketch of the panoramic view of the contact between Valles–San Luis Potosí (SLP) Platform and the Tampico-Misantla Basin along the Moctezuma River. (D) View of the northern side of Amajac River in the Cahuasas area, where the basal detachment of the Mexican fold-and-thrust belt is exposed. This detachment is located along the strongly sheared Santiago Formation, which is positioned between (relatively undeformed) Jurassic red beds and (strongly folded) Cretaceous basinal limestones. Gray—platform carbonates; white—basinal carbonates; dots—Late Cretaceous turbidites; dashes—Jurassic red beds.
planes of folds are subhorizontal, subparallel to the contact. Figure 7A is a schematic cross section to show structural style in the middle of the Zimapán Basin. It shows the striking variations among the different stratigraphic units, which maintain their stratigraphic position across the basin. There is no significant variation in trend of the fold axes among the different stratigraphic packages (Figs. 5E, 5G, and 5I); however, there are systematic differences in dip of the axial planar cleavage, which is generally greater in the younger units (Figs. 5F, 5H, and 5J).

At least five generations of fold-related veins can be recognized in the limestones of the Zimapán Basin. These folds show considerable amounts of flattening, evidenced by their tightened hinges. (C) Typical folds in mudstone layers interbedded with chert and thinner layers of shale of the Tamaulipas Formation. The folds in the limestone layers are asymmetrical and steeply inclined; they show angular hinges and strong variations in the thickness of the layers, with attenuated forelimbs and thickened hinges. (D) Gently inclined folds in shaly limestone layers interbedded with calcarenite, also showing thickened hinges. (E) Asymmetrical, almost upright fold, in shale layers of Trancas Formation. Note the pervasive axial planar cleavage associated with this fold. (F) Nearly recumbent fold in layers of calcarenite and volcanic graywacke near the transition between the Trancas Formation and a volcanioclastic sandstone unit.

Figure 7. (A) Schematic but realistic representation of a typical section in the middle of the Zimapán Basin, showing the dominant styles of folding observed in the different lithological units exposed within the basin. Truncation of the folds at the boundaries of the units is not implied; the folds are presumed to decay toward the boundaries or to deflect the boundaries, since the stratigraphic sequence is preserved across the basin. (B) Typical gently inclined folds observed in mudstone interbedded with calcareous shale, interpreted by stratigraphic position as the transition between the Tamaulipas and Soyatal Formations. These folds show considerable amounts of flattening, evidenced by their tightened hinges.
In the Tampico-Misantla Basin, folds vary in style from isoclinal (approaching the western boundary of the basin) to more open toward the east. Most are typical buckle folds of chevron style, except those near the Valles–San Luis Potosí Platform, which are more rounded and have thickened hinges and attenuated limbs. The average trend of the fold axes is ~340° for most of the units and on average is ~20° different in a clockwise sense from the average trend in the Zimapán Basin (Figs. 5K, 5M, and 5O). The axial planar cleavage is weak to incipient overall in the Tampico-Misantla Basin, and in general it dips more gently in the Early Cretaceous units as it approaches the detachment zone (Figs. 5L, 5N, and 5P). Asymmetry of the folds decreases to the surface, as sketched in Figure 8A and illustrated in the photographs.

Fold-related veins are far less abundant in the Tampico-Misantla Basin than in the Zimapán Basin. It is common to find three generations of them: veins parallel to bedding, en echelon arrays of veins that accommodated shear parallel to bedding, and extensional veins normal to 1 m.
Role of folding in the development of the Mexican fold-and-thrust belt

bedding positioned in the outer arc of folds. It is also common to find high concentrations of veins along thrust planes or along the detachment zones. These have a thickness on the order of a few centimeters, lengths of tens of centimeters, and commonly accommodate subhorizontal extension. Later tectonic veins, unrelated to folds or thrusts, are rather scarce.

To estimate the amount of shortening accommodated by folding in the basins, we applied a systematic analysis of fold geometry across the basins along the same stratigraphic horizon. We chose for this the Tamaulipas Formation because it is present in both basins, it is relatively thick (~600 m), it is mechanically competent, and, unlike the shale- or carbon-rich units (e.g., the Trancas, Tamán, Pimienta, and Santiago Formations), it did not develop a strong axial planar cleavage. In summary, we consider that it behaves mechanically as an (almost) ideal multilayer package in the context of buckling theory (sensu Hudleston and Treagus, 2010) and hence provides representative estimates of the overall shortening in these basins.

The geometrical analysis applied here involves several steps. First, we selected trains of well-exposed, near-upright to upright chevron folds. Folds at a total of 20 outcrops, 7 across the Zimapán Basin and 12 across the Tampico-Misantla Basin, were analyzed. For each train of folds, we measured the orientation of limbs and fold axes. For trains of folds that did not show evidence of flattening (as suggested by a change in thickness of the layers around the fold), we compared the presumed initial length (l_0) of layers prior to folding with the horizontal distance between the same end points after folding (l_f), as shown in Figure 9A. This provides a measurement of shortening strain accommodated by folding. For folds that showed evidence of flattening, we first estimated the amount of flattening by restoring the layers to constant thickness (as illustrated in Fig. 9B). Then, with image analysis, we removed the flattening strain, after which we calculated the shortening due to buckling alone by applying the same technique as for the folds that did not show flattening (Fig. 9A).

Finally, we added the shortening due to buckling and that due to flattening to calculate the total shortening (Figs. 9C and 9D).

The average shortening estimates (of at least five individual folds) for each analyzed train of folds are plotted against distance from the western edge of the cross section, for each of the two basins, in Figures 9C and 9D. The data in Figure 9C are for the western half of Zimapán Basin. On the left, note that the shortening due to buckling is fairly homogeneous across the basin, indicating that buckle folding reduced the horizontal length of the Tamaulipas Formation by ~45%. The central graph in Figure 9C indicates that shortening due to flattening is less homogeneous than that due to buckling, but it is more variable, as evidenced by the error bars. It averages ~20%. The graph on the right combines shortening due to buckling and flattening for each fold train. It indicates a total shortening strain due to folding of 60%–65%, which corresponds to 60 km across the whole basin. Figure 9D shows the corresponding data for the Tampico-Misantla Basin. Note that shortening due to buckling (Fig. 9D, left) is at a maximum close to the boundary of the basin with the Valles–San Luis Potosí Platform and decreases eastward toward the toe of the fold-and-thrust wedge, with an average of ~45%. The amount of shortening due to flattening shows a similar trend (Fig. 9D, center), with a maximum of 22% on the western edge of the basin, and essentially none on the eastern edge (front of the Mexican fold-and-thrust belt). The sum of the two components of shortening (Fig. 9D, right) is ~60% on the western side of the Tampico-Misantla Basin, decreasing to ~20% at the eastern edge (the front of the Mexican fold-and-thrust belt). This corresponds to a 35–40-km reduction in length. In addition to the shortening due to folding, there is some offset on minor thrusts and some amount of thickening of layers during the early stages of folding. Our estimates of shortening due to thrusting and bed thickening prior to folding in the two basins are 10% and 5%, respectively. These can be added to the shortening estimates due to folding to arrive at estimates of total shortening across the Zimapán and Tampico-Misantla Basins of 65% and 45%, respectively. In the case of the Tampico-Misantla Basin, shortening decreases from 65% in the west to 20% in the east (Fig. 3).

Values of shortening for each of the major paleogeographical elements, including the platforms and basins, encountered along the line of cross section are shown above the cross section in Figure 3B. In general, shortening increases modestly from east to west, and this observation is in agreement with the critical taper orogenic wedge model, which predicts that more deformation is accumulated at the rear of the wedge than at the front (Davis et al., 1983). However, because of lithological heterogeneity, the gradient is not smooth, with deviations that depend on lithology. The platforms exhibit less overall shortening than the basins, although the western platform (El Doctor Platform) accommodates more shortening than the eastern one (Valles–San Luis Potosí Platform). The two basins each accommodated more shortening than either of the platforms, but the western basin (Zimapán Basin) accommodated more shortening than the eastern one (Tampico-Misantla Basin).

Thus, both basins accommodated more shortening than the adjacent platforms (Fig. 3C). Although the deformation is accentuated toward the boundaries of the Zimapán Basin, the fold analysis suggests that the internal deformation of the basinal limestone is very homogeneous across the basin (Fig. 9C). In the Tampico-Misantla Basin, by contrast, we can see a gradient of deformation accommodated by folding in the Tamaulipas Formation that decreases toward the foreland.

Deformation along the Platform-Basin Boundaries

The boundaries between the basins and platforms along cross-section A–A′ in Figure 3 are typically observed as sharp tectonic contacts, in the form of thrusts. Considerable vertical displacements are accommodated along these thrusts. The stratigraphic offsets are largest to the west, on the order of a few kilometers (e.g., the Higuerillas and El Doctor thrusts; Fig. 3C), and decrease toward the east, the front of the Mexican fold-and-thrust belt, where they accommodate displacements of about a kilometer to a few hundred meters (e.g., Enramaditas thrust; Fig. 3C). The thrusts have a constant strike of NW–SE and dip toward the SW. Their dip is shallower toward the front of the fold-and-thrust belt, and the lineation on the thrust planes indicates a consistent direction of transport toward the NE. Most of the thrust zones exhibit, at least in part, a brittle fabric. Toward the west, deformation associated with the Higuerillas, El Doctor, and El Volantín thrusts exhibits a dominant cataclastic fabric, which locally, in fine-grain horizons, grades into a calc-mylonite. The cataclastic fabric is commonly cut by later reverse faults. Toward the east (e.g., Jiliapan, Puerto de Piedra, and Enramaditas), a coarser-grained brecciated fabric is more commonly associated with the thrusts, although finer cataclasites also developed in shale-rich horizons.

Thrust-related veins are thicker and more abundant along the Higuerillas and El Doctor thrust zones and on the western side of El Doctor Platform. Such veins are thinner and less abundant along the easternmost thrusts. This might suggest that less fluid circulated along the eastern thrusts during deformation, associated with the smaller displacements on these thrusts, consistent with the predictions of Hubbert and Rubey (1959).

AGE OF DEFORMATION

As shown in Figure 1 of Fitz-Díaz et al. (2011), there are two main orogenies in western North America involving Cretaceous-age rocks.
Figure 9. Illustration of two methods used to determine shortening in the basins, where the dominant structures are buckle folds, some of which experienced flattening. (A) Direct length of arc method. This was applied only in trains of upright chevron folds in layers that did not show significant variations in thickness or a pervasive axial planar cleavage. (B) Method for determining shortening due to flattening, which takes into account the following geometrical considerations: (1) Following Ramsay (1974), we take chevron folds to have linear limbs and the outer arc of the hinge to be defined by a circle centered at the vertex of the inner arc, with radius equal to the thickness \( t \) of the layer, which does not change during buckling. (2) If a homogeneous flattening is now imposed perpendicular to the axial plane of the fold, the circumscribed circle in the hinge transforms into an ellipse corresponding to the strain ellipse due to flattening. Most of the flattened chevron folds used in this analysis showed elliptical outer arcs in the hinges. Circumscribed ellipses in individual folds, on planes perpendicular to the fold axes, were retrodeformed to circles (using a computer graphics package) to find the flattening strain. Once flattening was removed, the method in A was used to find preflattening shortening. We also applied the method described by Ramsay (1974) to determine shortening in chevron folds, which gave similar results. (C) Averaged shortening for each train of folds analyzed in the Zimapán Basin plotted against distance from the western edge of the cross section, for buckling, flattening, and for the combination of buckling plus flattening. (D) As in C for the Tampico-Misantla Basin.
These are traditionally known as the Sevier and the Laramide. They have been well studied since the beginning of the twentieth century in the central part of the cordillera, and their ages have been relatively well documented in recent decades (Armstrong, 1968; Allmendinger et al., 1982; Brewer et al., 1982; Yonkee, 1992; Schmidt et al., 1993). However, the ways in which the effects of these two orogenies are reflected in the cordillera to the south into the Mexican fold-and-thrust belt are not yet well understood. This is partly because there is not a clear continuation of the fronts of the Sevier or Laramide fold-and-thrust belts to the south, as a result of a number of younger fault systems, and partly because the traces of the main Laramide thrusts are not continuous for long distances.

In northern Mexico, low-angle thrusts and related folds are cut by high-angle reverse faults (e.g., La Babia and San Marcos). The latter, according to Chávez-Cabello et al. (2007), were originally normal faults formed during the opening of the Gulf of Mexico, in the Middle Jurassic, and were apparently reactivated on multiple occasions. One of the most important reactivations occurred, according to the same authors, in the Paleocene–Eocene during the Laramide orogeny. This coexistence of thin- and thick-skinned structures in Cretaceous limestones makes it likely that both, Sevier and Laramide, structures are recorded in northern Mexico, as shown in the overlap area in Figure 1.

In the studied area in central Mexico, there are no large stratigraphic separations associated with any of the high-angle reverse faults related to D1. In the cross-section A–A′ of Figure 3, the Mexican fold-and-thrust belt has a wedge shape and is bounded below by a detachment zone, typical of thin-skinned fold-and-thrust belts (Fig. 3A). We did not observe high-angle faults cutting this detachment. Even the late shortening structures (D2) are represented by low-angle reverse faults and related folds (Fig. 10A). Based on these structural characteristics, we consider that the Mexican fold-and-thrust belt has more affinity in style with the Sevier fold-and-thrust belt than with the Laramide belt.

Stratigraphic relationships in the studied cross section, which can be consulted in Figure 2, suggest that syntectonic sedimentation (Soyatal Formation; Fig. 4A) on the western side of the cross section started in the Turonian (Hernández-Jáuregui, 1997). On the other hand, similar clastic deposits at the front of the Mexican fold-and-thrust belt (Méndez Formation; Fig. 4B) were strongly deformed during D1, as evidenced by a pervasive axial planar cleavage (S1), which is cut by younger thrust faults, presumably during D2. The Méndez Formation is unconformably covered by younger clastic units.

Figure 10. Kinematic models to explain the variation in folding styles observed in typical sections of the Zimapán (A) and Tampico-Misantla (B) basins. The left hand columns in A and B are from the columns in Figures 7A and 8A, respectively, with scaled thicknesses of the units (t—thickness). (A–B) Kinematic models involving a homogeneous pure shear followed by simple shear. (C–D) Kinematic models similar to those in A and B, but involving general shear with the components k and Γ applied to match the estimated strain in the different horizons of the two basins. On the right sides of C and D, we show the average strain for these sections, found by weighting k and Γ according to the thickness of each horizon. APD—axial plane dip. See text and Table 1 for details.
The angle of internal friction and fluid pressure base, and pore-fluid pressure (Davis et al., 1983). The deformation event that contributed most to the development of the Mexican fold-and-thrust belt in central Mexico, started in the west in the Turonian (ca. 90 Ma) and ended to the east in the Maastrichtian (ca. 65 Ma). Also, the second shortening event, D2, that affected these rocks occurred after the Maastrichtian.

In the Rocky Mountains of the United States, syntectonic basins in the Sevier fold-and-thrust belt have a range of ages between Cenomanian and Maastrichtian (Lawton and Trexler, 1991; Selting and Schmitt, 1999; DeCelles, 2004), while those associated with the Laramide structures are younger, with a range of ages between Maastrichtian and Eocene (Dickinson et al., 1988; DeCelles et al., 1991). The oldest syntectonic basins in the Sevier belt coincide with the age of syntectonic magmatism in the Sierra Nevada and Idaho Batholiths to the west (DeCelles, 2004; fig. 1 in Fitz-Díaz et al., 2011).

Comparisons of our data with the ages of the classic Sevier and Laramide orogenic belts indicate that D1 deformation in the Mexican fold-and-thrust belt seems to have occurred at least in part during the Sevier orogeny in the United States and Canada. Also, D2, the second episode of deformation in the Mexican fold-and-thrust belt, which accommodated minor shortening, seems to have been coincident in time with the Laramide orogeny in the United States and Canada. In this portion of the Mexican fold-and-thrust belt, the deformation style of Laramide-age structures (D2), however, does not correspond to the thick-skinned basement-involved style displayed by Laramide structures in the type location of the western United States.

**ROLE OF PLATFORMS IN THE DEFORMATION OF THE BASINS**

In a Coulomb critically tapered wedge, the taper angle is controlled by the angle of internal friction of the wedge material, friction at the base, and pore-fluid pressure (Davis et al., 1983). The angle of internal friction and fluid pressure (expressed as the lithostatic/hydrostatic pressure ratio) exert important controls on the deformation of rocks within the wedge. On the other hand, friction and the fluid pressure along the underlying detachment also play a critical role in determining the critical taper angle, and determining whether or not the wedge is displaced or thickened through internal deformation.

The platformal and basinal rocks in the Mexican fold-and-thrust belt constitute two distinct facies, despite the fact that the dominant lithologies of both are carbonates. The platform carbonates are massive or thickly bedded and are mainly calcite and dolomite. On the other hand, the basinal carbonates (calcite rich) are typically thinly bedded, with interbeds of carbonaceous mudstone, chert, and thin (<1 cm) layers of shale.

Platforms and basins present striking differences in the style of deformation, which is thrust dominated in the platforms, and fold dominated in the basins. Reasons for such differences are explained in Fitz-Díaz et al. (2011).

Independent of the variations in the deformation style, shortening of the Mexican fold-and-thrust belt platform carbonates is significantly less (55% for the El Doctor Platform and 35% for the Valles–San Luis Potosí Platform) than the shortening accommodated by the basal strata (~70% close to the platform borders) on either side. This deformation could be controlled by changes in the angle of internal friction, which could have been enhanced in the platforms by the presence of dolostone along the borders of the platforms and diagenetic dolomite mixed with calcite in the interior of the platform, which is scarce in the basins.

This difference in the amount of shortening makes it easy to visualize the platforms as relatively rigid objects, representing local backstopss concentrating shortening in the adjacent basins. For this to occur, the displacement along the detachment zone underneath the platforms must be facilitated by a “weaker rock.” In the western side of the area underneath El Doctor Platform and the Zimapán Basin, the lack of seismic information or borehole records makes it difficult to predict the lithology along the decollement.

In the Valles–San Luis Potosí Platform, however, borehole records in the surrounding areas and the exposure of anhydrite toward the base of the platform (Figs. 3A and 3B) make it likely that the detachment lies within these weak rocks, facilitating displacement. This is not the case for the Tampico-Misantla Basin, where strongly sheared carbonaceous shale and limestone of the Santiago Formation are pervasively present in the detachment zone. The rocks of the Santiago Formation are mechanically stiffer than evaporates and therefore less efficient in accommodating displacement.

The combination of a competent massive dolomitized limestone within the Valles–San Luis Potosí Platform and a weak detachment makes it possible that the Valles–San Luis Potosí Platform accommodated small internal shortening and large displacement, locally acting as a “backstop,” which induced deformation in the Tampico-Misantla Basin. The deformation in the Tampico-Misantla Basin, as in the Zimapán Basin, was accommodated by internal buckling (Biot, 1965), folding unrelated to faulting, and that was the result of stress applied parallel to a multilayer that possessed anisotropy (alternating layers with different viscosity and stiffness). This phenomenon has been documented in nature and experiments (e.g., Cobbold et al., 1971).

Because of the mechanical contrast between platforms and basins, the platforms play a major role in establishing boundary conditions for the deformation of the basins. In the Zimapán Basin, there is an intense internal deformation of fairly constant magnitude (Fig. 9C) and a constant orientation of structures within the same stratigraphic horizon. These characteristics can be related to the fact that this basin is confined between two platforms. In the Tampico-Misantla Basin, however, there is a deformation gradient from west to east (Fig. 9D), corresponding with a (fixed) boundary to the west (the Valles–San Luis Potosí Platform) and a free boundary to the east (the toe of the Mexican fold-and-thrust belt).

**KINEMATICS OF DEFORMATION WITHIN THE BASINS**

**Representative Vertical Sections**

Two well-exposed sections in the middle of the Zimapán and Tampico-Misantla Basins (Figs. 7A and 8A) were chosen to analyze fold kinematics. In these two sections, we analyzed fold geometry to estimate strain at the different stratigraphic/structural levels. The estimated finite (total) strain is characterized by the axial ratio of the strain ellipse and orientation of the long axis of this ellipse, θ. In deciding how to estimate strain, we took into account qualitative field observations interpreted in light of folding theory. This allowed the selection of an appropriate method to estimate strain. For the quantitative analyses, we applied a number of different methods, including: the one described in Figure 9 (which is based on Ramsay, 1974), the Wellman method (Willman, 1962; Shad and Srivastava, 2006), the method of Aller et al. (2008), and estimating strain within the layers by analyzing distortion of calcispherulid shells with the Rf/phi method (Dunnett, 1969; Lisle, 1985). The results are presented in the second column of Table 1.

In each section, several different stratigraphic/structural horizons or levels could be distinguished, taking into account the dominant lithology and folding style (tightness, asymmetry, hinge shape, and presence/absence of pressure-solution cleavage). Each horizon was
assigned a thickness (with precision of ±100 m) on the basis of field observation, and a characteristic column was drawn for each basin.

Figure 7 illustrates the dominant folding styles within the different horizons in the Zimapán Basin. This section shows how deformation was accommodated in the upper units of the Zimapán Basin—the Tamaulipas and Trancas Formations (see Fig. 4C for reference)—and the transition between these units and between these units and those above and below. The two units remain in their original stratigraphic position, and their overall thickening seems to be due to folding. The Tamaulipas Formation consists of mudstone and wackestone interbedded with abundant chert bands and thin layers of shale, while the Trancas Formation is dominantly carbonate siltstone interbedded with calcarenite and volcanic graywacke. In the Tamaulipas Formation, the deformation is mostly accommodated by asymmetric, steeply inclined, tight to close folds, and pressure-solution cleavage is only observed in the shale layers. In contrast, deformation of the Trancas Formation is accommodated by close to open folds with rounded hinges and a pervasive pressure-solution cleavage. In the lithological transitions, the asymmetry of folds is accentuated, and the dip of the axial planes is reduced, sometimes to the point that it becomes subparallel with the contact. One way to explain these variations in axial plane dip is to relate them to ramps and flats of major thrust faults, in such a way that the low-angle axial planar folds develop associated with shear along flats, while the steeper axial planar folds develop adjacent to ramps. However, the amount of displacement required to produce the observed shortening should also produce significant stratigraphic offsets or anomalous thickness variations in the different units, and these are not observed.

Figure 8 illustrates the dominant folding styles within the different horizons in the Tampico-Misantla Basin (the thickness-scaled representation can be observed in Fig. 10B). It includes most of the stratigraphic units involved in the deformation within the wedge down to the detachment zone (see Fig. 4C). All the units preserve their stratigraphic positions without duplication, except for the Tamaulipas Formation in the upper part of the cross section, which is duplicated by thrusting. This section includes slightly deformed red beds (Cahauas Formation) under the detachment, strongly sheared and faulted carbonaceous shale (lower part of Santiago Formation) along the detachment, and recumbent folds in association with thrust faults developed in carbonaceous limestone with scarce chert bands (upper part of Santiago Formation); above this, there are gently to moderately inclined asymmetrical tight folds with sharp hinges in carbonaceous shale with abundant shear bands (Tamán and Pimienta Formations), upright chevron folds in layers of noncarbonaceous limestone with chert bands (Lower Tamaulipas Formation), and steep folds in thinly bedded limestone with chert bands (Upper Tamaulipas Formation). We can see that, in general, the axial planes of folds are steeper upward and become subparallel to the detachment zone in the lower portions of the section. This gradual variation in axial plane dip can be explained in the framework of the orogenic wedge model by small differences in the strength of the rocks in the detachment and within the wedge, and with smooth stratigraphic transitions.

This kind of fold profile has been observed in analogue models focused on analyzing deformation in carbonate basins influenced by the presence of adjacent platforms (see fig. 2 in Dixon, 2004), a situation analogous to that in the Mexican fold-and-thrust belt.

### Kinematic Models

We analyzed the variation in fold style and finite strain in the two vertical sections represented by the structures illustrated in Figures 7 and 8 by considering two kinematic models. In doing so, we made these general simplifying assumptions: (1) Deformation is plane strain, and there is no volume change. (2) The deformation producing the asymmetrical folds is a combination of pure shear and simple shear, and the long axis of the bulk total strain ellipse (represented by angle $\theta'$) is parallel to the axial planes of the folds. (3) The trains of asymmetrical chevron folds in the Zimapán Basin and Tampico-Misantla Basin were formed in two stages: first, one of buckling that gave rise to asymmetrical chevron folds, followed by flattening that produced the observed changes in layer thickness and that helped develop asymmetry. (4) The deflection of the axial planes of the folds from the vertical is due to simple shear, which also contributed to the development of fold asymmetry and different limb lengths. (5) Discrete displacement along the contacts between the different stratigraphic units can be neglected. In the cross section, we indicate the places where thrust faults were observed in the field.

Assumptions 2 and 3 are supported by Dixon’s (2004) analogue models, in which he analyzes the effect of a stiff platform on the deformation of an adjacent, softer, well-bedded basinal unit. His physical models show that, in most cases, the basal layers develop symmetrical buckle folds that preserve layer thickness in the initial states of deformation and progressively become asymmetrical and inclined, with thickened fold hinges.

The analysis of the geometry of trains of chevron folds in the same horizon along the whole cross section of Figure 3C strongly supports assumption 3. Assumption 4 is also supported by Dixon’s analogue experiments and the analysis of strain associated with the trains of chevron folds in this study (Fig. 9). The folds analyzed in the Zimapán Basin and on the western side of the Tampico-Misantla Basin exceeded the locking angle for chevron folds (Ramsay, 1974) and accommodated a shortening of close to 45% (after removing flattening), while the folds toward the eastern side of the Tampico-Misantla Basin show a progressively more open interlimb angle and do not show

### Table 1: Comparative Chart of Axial Ratios and Orientations of the Strain Ellipse Calculated in the Kinematic Models Consisting of Pure Shear Followed by Simple Shear, General Shear, and Estimates from Fold Analysis

<table>
<thead>
<tr>
<th>Horizon number</th>
<th>Estimates from folds</th>
<th>Pure shear then simple shear, $k = 0.5$</th>
<th>Strain ellipse input for general shear model</th>
<th>Strain components in general shear</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$R$</td>
<td>$\theta'$</td>
<td>$k$</td>
<td>$\gamma$</td>
</tr>
<tr>
<td>1</td>
<td>10–13</td>
<td>25–32°</td>
<td>17.5</td>
<td>28.5°</td>
</tr>
<tr>
<td>2</td>
<td>7–12.5</td>
<td>40–63°</td>
<td>6.8</td>
<td>51.5°</td>
</tr>
<tr>
<td>3</td>
<td>9–14</td>
<td>29–38°</td>
<td>17.5</td>
<td>33.5°</td>
</tr>
<tr>
<td>4</td>
<td>6.5–9.5</td>
<td>57–73°</td>
<td>4.9</td>
<td>65°</td>
</tr>
<tr>
<td>5</td>
<td>8.5–14.5</td>
<td>20–27°</td>
<td>25.3</td>
<td>23.5°</td>
</tr>
<tr>
<td>6</td>
<td>5–6.4</td>
<td>68–72°</td>
<td>4.5</td>
<td>70°</td>
</tr>
<tr>
<td>7</td>
<td>4.2–5.5</td>
<td>82–85</td>
<td>4.1</td>
<td>83.5°</td>
</tr>
<tr>
<td>8</td>
<td>9–14</td>
<td>36–44°</td>
<td>9.8</td>
<td>40°</td>
</tr>
<tr>
<td>9</td>
<td>18–25</td>
<td>16–24°</td>
<td>34.3</td>
<td>20°</td>
</tr>
<tr>
<td>10</td>
<td>25–36</td>
<td>10–16°</td>
<td>7</td>
<td>13°</td>
</tr>
</tbody>
</table>

Note: $R = \text{axial ratio} = (1 + \epsilon_1)/(1 + \epsilon_2)$; $\theta' = \text{inclination of (1 + \epsilon_1)}$; $k = (1 + \epsilon_2)$, pure shear component of the deformation matrix; $\gamma = \tan \psi = \text{simple shear component}$; $\Gamma = \text{effective shear strain in general shear}$. 

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evidence of flattening, suggesting that they did not reach the locking angle. These variations in geometry, from much-evolved folds in the Zimapán Basin to less-deformed folds at the front of the fold-and-thrust belt, strongly support the idea of the basinal layers being initially deformed by buckling, giving rise to trains of symmetrical chevron folds, followed by flattening, which accentuated the asymmetry, in a deformation event that involved both pure shear and simple shear. The fact that pressure solution is locally important suggests that assumption 1 is not fully satisfied, and the fact that different units (e.g., the Tamaulipas and Trancas Formations in the Zimapán Basin) exhibit different pervasiveness of pressure-solution cleavage implies further that assumption 5 is not fully satisfied.

Pure Shear + Simple Shear Models

The different units in the representative sections in the two basins are scaled according to thickness in Figure 10, and the rock columns so produced provide the basis for two different kinematic models. In the first model (Figs. 10A and 10B for Zimapán Basin and Tampico-Misantla Basin, respectively), we assume that the different units (1–10 in Figs. 10A and 10B) experienced a homogeneous pure shear in a first stage of deformation, followed by different amounts of simple shear necessary to produce folds with the observed asymmetry, as expressed by variations in the axial plane dip of the folds in Figs. 7 and 8. Mathematically, this is expressed by the linear transformation matrices, \( D_p \) and \( D_s \), which combined are equivalent to a transformation \( D \):

\[
D = D_p D_s = \begin{pmatrix} 0 & \frac{\gamma}{k} & \frac{\gamma k}{k-1} \\ \frac{\gamma k}{k-1} & \frac{\gamma k}{k-1} & \frac{\gamma}{k} \\ \frac{1}{k} & \frac{1}{k} & \frac{1}{k} \end{pmatrix} = \begin{pmatrix} \frac{k}{1} & \frac{\Gamma}{1-k} \\ 0 & 1 \\ 0 & 1 \end{pmatrix},
\]

where the pure shear in the horizontal (x-direction) is given by \( k \), and the simple shear is given by \( \gamma \). We applied a pure shear \( k = 0.5 \) corresponding to a maximum shortening of 50% taking into account the 45% shortening that we estimated through the analysis of chevron folds (Fig. 9) and taking into account the fact that buckled layers experience a small amount of homogeneous shortening prior to fold growth at nearly constant limb length (e.g., Hudleston and Treagus, 2010).

Our reference direction (x-axis) is horizontal, since shortening is subparallel to this direction. By applying a bulk pure shear \( k = 0.5 \) to the layers, we can produce upright buckle folds with an interlimb angle of between 65° and 70°, as sketched in Figures 10A and 10B. The amounts of simple shear that must be added to this to produce folds with axial plane dips similar to the ones observed in the field (column 5 in Table 1, which is the median value of the range given in column 3) are given in the sixth column of Table 1. The sketches in Figures 10A and 10B illustrate how pure shear might produce folding by buckling in a first stage of deformation and how asymmetry and flattening of folds can be related to the subsequent simple shear.

Even though this simple model illustrates the overall effects of buckling and flattening during folding, the axial ratio of the finite strain ellipse associated with the folds in each horizon, as calculated from Equation 1 (column 4 in Table 1), is different from our estimated ranges (column 2 in Table 1). We, therefore, consider an alternative model.

General Shear Kinematic Model

There are two differences between the first model outlined here and the second model. One, we do not assume that the horizontal shortening experienced by all units in the rock column is the same, and, two, we substitute simple shear for the model in which simple shear precedes pure shear. We follow the method and nomenclature of Tikoff and Fossen (1993) in combining simple shear and pure shear into general shear, in which the two components occur simultaneously. The basic assumptions 1 through 5 above remain the same. The deformation matrix is now given by:

\[
D = \begin{pmatrix} k & \Gamma \\ 0 & 1/k \end{pmatrix}.
\]

In general shear, every finite strain ellipse (characterized by its axial ratio and orientation) is the result of a unique combination of \( k \) and \( \Gamma \). By using the average estimates of strain magnitude and orientation in the two characteristic cross sections, determined from the strain analysis using the folds, and using Equation 2 and the equations given in Appendix B of Ramsay and Huber (1983), we calculated \( k \) and \( \Gamma \) for each horizon for the two analyzed cross sections (1–10 in Fig. 10). The results are given in columns 9 and 10 of Table 1 and are represented diagrammatically in Figures 10C and 10D.

We see in Table 1 and Figures 10C and 10D that the pure shear and simple shear components vary among the 10 different horizons in the two analyzed cross sections. In the case of the Zimapán Basin, the variations in folding style and the amounts shortening (\( k \)) and shear (\( \Gamma \)) required to match the estimated finite strains in the different horizons are not systematic. More shortening and less shear are required to produce the steeply inclined asymmetrical folds, while less shortening and more shear are required to produce the nearly recumbent folds under a general shear regime (Table 1). The alternating distribution of practically upright and almost recumbent folds in the analyzed cross section (Figs. 10A and 10C) might be due partly to stratigraphic heterogeneity and partly due to the nature of the propagation of deformation within the basin. If deformation in the Zimapán Basin developed without internal thrusts, then the shortening should be almost independent of stratigraphic level, regardless of the relative contributions of pure shear and simple shear, with the whole package riding eastward on the basal detachment. If there are internal thrusts, and they developed in the classic piggyback, forward-propagating style, then we would expect to see less shortening at deeper levels, reflecting the progressive incorporation of deeper stratigraphic units into the wedge (see, for example, the detailed measurements of shortening by Lebel et al. [1996] in the foothills of the Canadian Rockies). The fact that we do not observe significant internal thrusts, which would repeat stratigraphy, and the fact that there is not a simple pattern of increase in shortening upward suggest either that there was some minor out-of-sequence thrusting to create irregular variations in the amount of shortening with depth (as illustrated in Fig. 11A), or that there is more complex deformation partitioning in the basinal strata that does not involve duplication of units by thrusting. It is not surprising that the shortening within individual units might be variable, although the fold data in Figure 9 suggest that shortening on average is fairly constant across the basin.

In the Tampico-Misantla Basin (Figs. 10B, 10D, and 11B), the situation is simpler. The pure shear (shortening) component of deformation increases toward the upper part of the section, in the younger units, while the simple shear component decreases. This systematic variation is commonly observed in fold-and-thrust belts that have experienced in-sequence forward propagation of thrusts (Elliott, 1976; Boyer and Elliott, 1982), where the strength difference between the detachment and the rocks within the wedge is not very large. This phenomenon has been documented in the foreland of the Canadian Rocky Mountains fold-and-thrust belts, where shortening is mostly accommodated by thrusting (Lebel et al., 1996), as well as in physical experiments (Dixon, 2004). In the Tampico-Misantla Basin, however, later out-of-sequence thrusts might have accentuated deformation in the upper part of the Tamaulipas Formation.
Deformation in the Mexican fold-and-thrust belt is thin skinned and occurred in two phases, D1 and D2, with D1 being far more important in producing the observed structures. D2 gave a modest penetrative overprint to D1 and is also manifest in kilometer-spaced small-displacement thrusts. The occurrence of syntectonic turbidites (Soyatal Formation) suggests that D1 on the western side started in the Turonian and finished in the east in the Maastrichtian. The absence of D1 structures in Paleogene clastic deposits to the front of the Mexican fold-and-thrust belt indicates that D2 can be as old as the Paleocene–Eocene.

The systematic study of D1 structures along a cross section in central Mexico allowed us to analyze the role of the stratigraphy in the deformation of the marine sedimentary cover within the fold-and-thrust belt wedge and along the detachment. The carbonate cover sequence is mechanically heterogeneous as a result of lateral facies variations corresponding to paired platform-basin depositional environments, resulting in more or less competent rock packages. This mechanical heterogeneity favored different deformation styles (dominated by kilometer-spaced thrusts in the platforms and by mesoscopic folds in the basins) with different amounts of overall shortening deformation being accommodated in the different segments. Vertically, the stratigraphy controlled variations in strength between the detachment and the rocks within the wedge, which in turn play an important role in the distribution of strain within the wedge.

The geometry of mesoscopic folds in the different stratigraphic units in the basins shows a consistent geometry. Based on observations and measurements of these folds, we estimated and averaged strain for distinct stratigraphic units within each basin and the transition zones between units. The overall pattern of strain, and strain magnitude and orientation within each unit/stratigraphic level, can be accounted for by models involving pure shear and simple shear. In the models, we assumed that the folds originated as symmetric upright chevron folds and that their current inclined orientation is a result of simple shear, with their axial planes parallel to the maximum bulk finite extension direction. In one model, we assumed that shortening was constant within each basin and the same at all stratigraphic levels. This does not give a good match to the strain magnitudes at the various levels. In a second model, we assumed general shear and computed the amounts of pure shear and simple shear required to produce the strains determined from the fold measurements. For the Zimapán Basin, this model, with some out-of-sequence thrusting or with some other form of strain partitioning, could explain the observed strain pattern. For the Tampico-Misantla Basin, this model, involving normal in-sequence thrusting, can well explain the observed strain variations.

Overall, we get a reasonable approximation of the average strain in the basins if we weight the contributions of pure and simple shear in the general shear model by the thickness of the different horizons.

We finally conclude that mesoscopic analysis of folds provides a good means of estimating strain in fold-dominated parts of fold-and-thrust belts and provides the basis for developing kinematic models of deformation in fold-and-thrust belts and the evolution of basins during deformation.

CONCLUSIONS

Deformation in the Mexican fold-and-thrust belt is thin skinned and occurred in two phases, D1 and D2, with D1 being far more important in producing the observed structures. D2 gave a modest penetrative overprint to D1 and is also manifest in kilometer-spaced small-displacement thrusts. The occurrence of syntectonic turbidites (Soyatal Formation) suggests that D1 on the western side started in the Turonian and finished in the east in the Maastrichtian. The absence of D1 structures in Paleogene clastic deposits to the front of the Mexican fold-and-thrust belt indicates that D2 can be as old as the Paleocene–Eocene.

The systematic study of D1 structures along a cross section in central Mexico allowed us to analyze the role of the stratigraphy in the deformation of the marine sedimentary cover within the fold-and-thrust belt wedge and along the detachment. The carbonate cover sequence is mechanically heterogeneous as a result of lateral facies variations corresponding to paired platform-basin depositional environments, resulting in more or less competent rock packages. This mechanical heterogeneity favored different deformation styles (dominated by kilometer-spaced thrusts in the platforms and by mesoscopic folds in the basins) with different amounts of overall shortening deformation being accommodated in the different segments. Vertically, the stratigraphy controlled variations in strength between the detachment and the rocks within the wedge, which in turn play an important role in the distribution of strain within the wedge.

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