Thick-skin-dominated orogens; from initial inversion to full accretion: an introduction

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Abstract: Fifty per cent of orogens have a thick-skin character, and have evolved from passive margin and intra-cratonic rift systems. One group of thick-skin provinces can be found at both pro- and retro-wedges of orogens associated with advancing subduction zones, that is, orogenic wedges whose advance vectors oppose the mantle flow. A second group can be found at pro-wedges of orogens associated with retreating subduction zones, that is, orogens whose advance vectors have the same direction as mantle flow. A third group is formed in intra-plate settings where mechanical strengthening is produced by internal shortening. Thick-skin province development is controlled by driving factors such as individual plate movement rates, overall convergence rates, orogen movement sense with respect to mantle flow, and pro-wedge v. retro-wedge location. These driving factors are themselves constrained by numerous internal and external factors. This introductory chapter focuses primarily on least-deformed case areas in order to understand the role of different factors in controlling the evolution of thick-skin tectonic provinces from the initial inversion stage to full accretion stage.

Many thrust belts exhibit both thin- and thick-skin structural styles in different portions of the belt. Some thrust belts, such as the Andes, change style along-strike (Allmendinger et al. 1997; Baby et al. 2013; Carrera & Muñoz 2013; Iaffa et al. 2013; Moretti et al. 2013). Other thrust belts, such as the US Cordillera–Rocky Mountains, exhibit thin-skin structural styles in their interior and thick-skin style in their exterior (Hamilton 1988). A quick screening of the basic characteristics of orogens whose deformation styles are known (see Nemčok et al. 2005) shows that 50% of orogens have a thick-skin character. They have evolved from either passive margins or intra-cratonic rift systems. The most typical examples of the former (initial passive margin setting) are various Andean thick-skin provinces, the Tianshan and the Western Alps (Schmid et al. 1996, 1997; Escher & Beaumont 1997; Baby et al. 2013; Carrera & Muñoz 2013; Iaffa et al. 2013; Kober et al. 2013; Tesón et al. 2013). The most typical examples of the latter (initial intra-cratonic setting) include the High Atlas, Middle Atlas, central and eastern Pyrenees and the Palmyrides (Carola et al. 2013; Teixell & Babault 2013).

Each of these examples underwent a different portion of the complete deformation history of a thick-skin orogen, that is, from initial inversion to full accretion. An excellent introduction to this complete history can be found in Teixell & Babault (2013), who look at shortening across a number of orogens. They record a total shortening of 20–25% across the High Atlas, 25–30% across the Eastern Cordillera of Colombia and about 40% across the Pyrenees. While Teixell & Babault (2013) reconstructed the complex deformation sequence in the Pyrenees and understood that mountain building comes from the numerous large-scale footwall edge short-cut faults through the rift margin, a less deformed record of the Eastern Cordillera allows one to see that a slow Paleocene–Eocene deformation was followed by the faster Oligocene–Early Miocene deformation and then by an accelerated deformation that took place since Late Miocene (Mora et al. 2013; Silva et al. 2013). Work on the timing of these
pulses of deformation by Jimenez et al. (2013) reveals that:

1. the faster Oligocene–Early Miocene deformation was characterized by the coeval activity of the main inversion faults and footwall short-cut faults, with inversion faults accommodating displacement rates, which were larger than rates of short-cut faults; and,
2. the fastest Late Miocene–Recent deformation was characterized by coeval activity of the main inversion faults and footwall short-cut faults, with short-cut faults accommodating displacement rates, which were larger than rates on the inversion faults.

By comparing the High Atlas, Eastern Cordillera and Pyrenees, Teixell & Babault (2013) have concluded that initial inversion is characterized by:

1. distributed deformation of the rift fill;
2. low structural relief; and
3. long longitudinal rivers.

Meanwhile, full accretion is characterized by:

1. strong inversion, located mainly at rift margins;
2. large basement-involved faults;
3. higher relief; and
4. river reorganization to transverse system.

Further shortening, they argue, would lead to the development of a doubly vergent orogen.

This Special Publication contains contributions that were presented at the meeting ‘Thick-Skin-Dominated Orogenes: From Initial Inversion to Full Accretion’ hosted by ECOPETROL-ICP (Instituto Colombiano del Petróleo) at Barichara, Colombia, during 24–27 January 2011. The key problems that are addressed in this volume include:

- the role of rheologies of deforming layers in thick-skin orogenic deformation;
- the role of existence v. non-existence of potential detachment horizons in thick-skin orogenic deformation;
- the role of basement buttresses in thick-skin orogenic deformation;
- the role of crustal (lithospheric) thickness variations in thick-skin orogenic deformation;
- the role of inherited strength contrasts in thick-skin orogenic deformation;
- the role of pre-existing anisotropy in thick-skin orogenic deformation;
- the role of syn-tectonic erosion and deposition in thick-skin orogenic deformation.

This introductory chapter addresses the dynamic settings of thick-skin provinces, their dynamics, choice of the optimum case areas for a study of the entire development from initial inversion to full accretion, and factors controlling the orogenic deformation in thick-skin provinces.

**Dynamic setting categories of thick-skin provinces**

Thick-skin provinces can be further divided, according to their tectonic/orogenic setting, into:

1. provinces located in a pro-wedge setting (Fig. 1), for example, the Ouachitas, the Apennines, the Balkans, the Carpathians, the Southern and Western Alps (Arbenz 1989; Shumaker 1992; Doglioni 1993a, b; Kley & Eisbacher 1999; De Donatis et al. 2001; Nemčok et al. 2006; Stuart et al. 2011);
2. provinces located in a retro-wedge setting (Fig. 1), for example, the Rocky Mountain basement uplifts, the Coastal Cordillera of Venezuela, the Central Cordillera of Colombia, the Eastern Cordillera of Colombia and Peru, the Sub-Andes of Ecuador, the Salta and the Sierras Pampeanas regions of the Argentinian Andes (Eva et al. 1989; Oldow et al. 1989; Lowell 1995; Allmendinger et al. 1997; Baby et al. 2013; Jimenez et al. 2013; Mora et al. 2013; Moreno et al. 2013; Silva et al. 2013; Tesón et al. 2013); and
3. provinces having an intra-plate position, for example, the North Sea region, the High Atlas and, perhaps, the Sierra de Perija and Merida Andes (Chigné et al. 1996; Teixell et al. 2003; Zanella & Coward 2003; Bayona et al. 2013).

Orogenics can be subdivided, following Doglioni (1993a, b), into those whose advance vectors oppose the mantle flow and those whose advance vectors have the same direction as mantle flow (Fig. 2a, b).

A good example of the former, that is an orogen at an advancing subduction zone (Fig. 2a, right side, 2b, lower panel, Fig. 3, upper panel), is the Western Alps (Fig. 4a). The best results on the involvement of subducting crust in their structure come from the Swiss segment of the Alps thanks to a unique combination of the European Geotraverse (EGT), the National Research Program on the Deep Structure of Switzerland (NFP 20) and probably the best geological and supplementary geophysical data on any collisional orogen worldwide (e.g. Schmid et al. 1996). These data indicate that the whole upper European crust is involved in accretion to the advancing orogen. Updoming of the most external basement block is interpreted as a crustal-scale ramp fold related to a detachment at the interface between lower and upper European crust. This detachment is interpreted to be
kinematically linked to the lower crustal wedge of the over-riding orogen. Complete detachment near the top of the highly reflective European lower crust indicates a strength minimum situated immediately above the lower crust. The unchanged thickness of the European lower crust traced all the way from the Alpine foreland to a location beneath the orogen argues for its relatively high strength. As a general rule, the amount of strain increases from the most external basement mas- sifs, where strain is partitioned into separate shear zones, to the more internal units, where strain is penetrative (Escher & Beaumont 1997). This early deformation took place under greenschist-facies conditions in the external nappes and under amphi bolite facies in the more internal nappes. Most of the cover of the European upper crust was detached during formation of the basement nappes. Cover nappes in the Alps utilized weak horizons and were generated by brittle deformation mechanisms operating simultaneously with ductile mechanisms in the basement fold nappes (Escher et al. 1993; Epard & Escher 1996). The temperature threshold for the ductile deformation in the basement was assumed to be approximately 300 °C, following studies of Handy (1989, 1990), and interpreted as occurring below the depth of this isotherm. Ductile basement nappes are fold structures formed by basal simple shear and superimposed pure shear (Handy et al. 1993; Schmid et al. 1997).

Numerical models made for orogens at advancing subduction zones have promoted an understanding of the basic mechanical controls on crustal deformation, assuming that orogenesis involved phases of oceanic subduction and continental collision, together with a transitional period between these phases (Beaumont et al. 1996; Ellis & Beaumont 1999; Fig. 1). Total subduction involves the asymmetric normal convergence between two lithospheric plates, in which the subducting lithosphere passes completely beneath the overriding lithosphere (Ellis & Beaumont 1999). Partial subduction and continental collision have been postulated to be variations of this process, in which the slab pull of the subducting lithosphere is insufficient for total subduction, and consequently only part of the lithosphere subducts and the rest becomes part of the orogen. The subduction/collision transition is understood as the time when the first basement nappes are derived from the subducting crust at great depths (Ellis & Beaumont 1999). These depths, based on examples of basement nappes from the Alps, are at about 70 km during early stages of collision (Escher & Beaumont 1997).

![Fig. 1. Plain-strain finite-element model of subduction–collision transition (Beaumont et al. 1999). Retro-lithosphere is fixed. Proto-lithosphere moves as indicated by white arrows. Phase 1: negative buoyancy of the subduction load drives the entire pro-lithosphere subduction. Phase 2: subduction load starts to decrease (increasing buoyancy) either with entrance of continental margin into the pro-lithosphere subduction or with slab break-off. Progressively more pro-lithosphere detaches and doubly vergent deformation begins. Phase 3: pro-lithosphere detaches entirely to form pro-wedge. Continued retro-transport creates a retro-wedge. The main differences among individual phases are controlled by increased buoyancy of proto-lithosphere (i.e. decreased downward flexing under slab pull and/or other forces).](image-url)
Fig. 2. (a) Orogens, foredeeps and foreland basins differentiated on the basis of the related subduction; namely systems with orogen advance vector the same as that of mantle flow (west-dipping, left-hand-side diagram) and systems with orogen advance vector opposing that of mantle flow (east or NE-dipping, right-hand-side diagram; Doglioni 1993b). Foredeeps and thick-skinned tectonic regions develop at three sides of these two systems: (1) at the front of pro-wedge of the west-dipping subduction system; (2) at the front of pro-wedge of the east or NE-dipping subduction system; and (3) at the front of its retro-wedge. Orogen advance vector is indicated by small arrow and mantle flow is shown by large arrow. Left side shows retreating subduction zone; right side shows advancing subduction zone. (b) Main differences between orogens whose advance vector is the same as that of mantle flow (west-dipping) and orogens whose advance vector opposes that of mantle flow (east or NE-dipping) shown in more detail, using the same geographic orientations as those in (a) (Doglioni 1993b). Orogens with advance vector the same as mantle flow are characterized by low structural and morphological elevation. The rocks occurring at the surface indicate relatively shallow burial. The tangent to the anticlines of the accretionary wedge descends orogen-wards. The depocentre of the deep foredeep basin is within the accretionary wedge. Orogens with advance vector opposite to mantle flow are characterized by low structural and morphological elevation. The rocks occurring at the surface indicate relatively deep burial. The tangent to the anticlines of the accretionary wedge ascends orogen-wards. The depocentre of the shallow foredeep basin is in front of the accretionary wedge. The curves on the right side represent a possible elevation history of a reference point, which was
Good examples of the latter, that is, orogens formed at a retreating subduction zone (Fig. 2a, left side, 2b, upper panel, Fig. 3, lower panel), are the West Carpathians (Fig. 4b) and the Eastern Balkans. Unlike the basement involvement known from the Western Alps, the thick-skinned tectonics of these regions developed during the latest stages of thin-skinned tectonics (Nemčík et al. 2006; Stuart et al. 2011). Furthermore, the depth (70 km) of basement nappe formation described above for the Western Alps is much greater than the depth range of thick-skinned tectonics determined from balanced cross sections through the Carpathian and Balkan accretionary wedges (Fig. 4b). This is only several tens of kilometres, indicating different controls on orogens formed at retreating subduction zones – compared with those formed at advancing subduction zones.

The difference between orogens at retreating subduction zones and those at advancing zones is also highlighted by direct comparison of the relatively large convergence magnitude in the Alps

Fig. 3. Conceptual interpretation of the two end-member convergence scenarios (Waschbusch & Beaumont 1996).
(a) Scenario without subduction zone retreat requires space for crustal accretion to be created by retro-thrusting.
(b) Scenario with subduction zone retreat manages to create space for crustal accretion without a need for retro-thrusting.

Fig. 2. (Continued) originally located in the foredeep, during development of the respective orogen. The upper curve shows a reference point affected by the migration of three main tectonic events, A, B and C. The foredeep subsidence, A, is driven by roll-back of the subduction hinge and complicated by smaller-scale uplift events associated with individual thrust sheet accretion pulses. The subsidence can be as high as 1.6 mm a\(^{-1}\). It is followed by an uplift, B, controlled by mantle wedging at the top of the subduction hinge, shear and transition from shortening to extension. Segment C of the curves denotes the back-arc basin subsidence. The lower curve shows a reference point affected by migration of the two main tectonic events, D and E. The subsidence, D, is controlled by either thrust loading affecting the foredeep or localized subsidence controlled by out-of-sequence thrusts. The characteristic subsidence is about 0.3 mm a\(^{-1}\). The uplift, E, is controlled by folding.
(400–450 km) with that of the Carpathians (75–223 km; Platt et al. 1989; Roca et al. 1995; Schmid et al. 1996; Nemčok et al. 1999). The difference is also indicated by the fact that the entire upper crust of the more proximal parts of the European plate was accreted to the Alpine orogenic wedge, while involvement of the basement of the European plate in the West Carpathian accretionary wedge is minor (Roure et al. 1993; Roca et al. 1995; Schmid et al. 1996; Nemčok et al. 2000). Orogens at retreating subduction zones also have different thermal regimes. Geodynamic models of orogens at advancing and retreating subduction zones, based on systematic comparison of thrust-belts worldwide and on numerical simulations, indicate that the retreating situation is characterized by more heat being available (Dolgioni 1992; Royden 1993; Waschbusch & Beaumont 1996). This is
thought to be because the retreating subducting slab creates a gap that is commonly assumed to be filled by upwelling of hot asthenosphere.

When we consider the differences between the two types of orogens and divide each into pro- and retro-wedge cases, one can estimate the energy budgets involved in their development and the intensity of their deformation. Accordingly, the thick-skinned regions of pro-wedges (Fig. 2a) group into ‘forceful scenarios’, located in orogens at advancing subduction zones (Fig. 2a, right side), such as the Ouachitas and the Southern and Western Alps (Fig. 4a), and ‘weak scenarios’, located in orogens at retreating subduction zones (weak orogen; Fig. 2a, left side), such as the Apennines, the Balkans and the Carpathians (Fig. 4b). Interestingly, the thick-skinned regions of retro-wedges cannot be divided into ‘forceful’ and ‘weak scenarios’, because all of those known to us are located on the retro-side of the orogens developed at advancing subduction zones (forceful orogen; Fig. 2a). This, perhaps, indicates that the weak orogen does not have sufficient energy to develop any major thick-skinned provinces on its retro side. Furthermore, as shown by numerical models (Waschbusch & Beaumont 1996), plate settings with weak orogens and retreating subducting plates manage to create space for crustal accretion without a need for retro-wedge development (Fig. 3, lower panel), while the settings with forceful orogens and without subducting plate retreat require space for crustal accretion to be created by retro-wedge development (Fig. 3, upper panel).

The third type of thick-skinned province, that is the intra-plate thick-skinned provinces, seems to develop in situations where one of the converging plates needs to be strengthened by internal shortening. Occasionally this shortening can be significant, as for example in the Higher Atlas, which underwent both full inversion and accretion of the syn-rift fill (see Teixell et al. 2003; Tesón 2009; Teixell & Babault 2013).

Choice of optimum case area for a study of the entire development from initial inversion to full accretion

All the above indicates that it is an energy balance that controls the presence or absence of thick-skinned domains along pro-wedges, retro-wedges and contracted intra-cratonic rift systems. Thick-skinned provinces in forceful pro-wedge scenarios are apparently characterized by an excessive energy budget to spend on orogenic development. As a consequence, the resultant deformation is so intense and complex that it is rather difficult to restore individual snap shots of the entire deformation history.

While the forceful scenario seems to develop a thick-skinned province under an excessive energy budget, and its individual deformation events are not always separable, the weak pro-wedge and forceful retro-wedge scenarios may develop thick-skinned provinces, which allow us to see the entire complex sequence of:

1. new fault propagation events;
2. reactivation events on pre-existing faults;
3. energy events consumed on opposing gravity forces; and
4. internal deformation events on the thick-skinned block.

This is because they involve smaller energy budgets associated with less intense deformation and individual deformation events are more likely to be separable. Examples of the weak pro-wedge, including the Carpathians (Fig. 4b) and the Balkans, indicate that thick-skinned tectonics occurs at later stages of the orogenic build-up, when the taper angle is high, the thermal regime is elevated and an extra mechanism for out-of-sequence shortening is required to increase the orogenic taper further (see Nemčok et al. 2006; Stuart et al. 2011). The Eastern Cordillera, an example of the forceful retro-wedge detached at a depth of about 30 km (Mora et al. 2008, 2010; Parra et al. 2009; Fig. 5), indicates that a new thick-skinned tectonic event occurs at stages when the orogenic foreland refuses to flex under the advancing orogen and the orogen needs to build up extra strength in order to advance any further (Hermeston & Nemčok 2013).

For the same reason, that is, the inability of the orogenic foreland to flex, the most extreme examples of vast areas affected by thick-skinned tectonic events come from narrow foreland plates trapped between two orogens moving towards each other. The two known examples come from the Apulian plate between the Dinarides and the Apennines (Bertotti 2000, pers. comm.) and the Kura Valley basin between the Greater and Lesser Caucasus systems (Nemčok et al. 2013). They both illustrate an energy balance interplay between factors controlling orogenic and foreland mechanical stratigraphies.

This is well illustrated in the inverted Broad Fourteens basin of the North Sea, where it can be seen that local development of either thick-skinned or thin-skinned deformation is controlled by the local salt thickness (see Nalpas et al. 1995; Brun & Nalpas 1996; Fig. 6). This inversion shows clearly that the amount of coupling between the orogen and foreland plate is another important factor controlling the overall energy balance. This is further demonstrated by the fact that even...
Fig. 5. Regional balanced cross section through the Eastern Cordillera, Colombia with earthquake hypocentres/foci (circles) projected (Ingeominas 2009; Tesón et al. 2013).

Fig. 6. Structural map and line-drawing interpretation of seismic sections showing variation in structural style in the Broad Fourteens basin as a consequence of development of the Zechstein salt horizon (modified from Nalpas et al. 1995).
orogens characterized by high convergence rates, such as the Rocky Mountains in Alberta, can lack basement uplifts or any kind of compressional intraplate deformation in their foreland if the foreland does not contain any major intra-plate discontinuities, or if it lacks mechanical coupling with the orogen (Ziegler 1989). As indicated by examples from the West Carpathians (Nemčok et al. 2001, 2006) and Eastern Cordillera (Hermeston & Nemčok 2013), the coupling can be enhanced either by a lack of any major potential detachment horizon or by the presence of significant regional buttresses.

Honouring all the above, avoiding forceful pro-wedge scenarios and focusing on weak pro-wedge or forceful retro-wedge scenarios should be the best strategy when attempting to understand the role of different factors in controlling thick-skinned tectonics (Fig. 7, Table 1). This volume attempts to follow this strategy by examining the factors involved in the energy balance behind inversion and the potential subsequent full accretion from thick-skinned regions worldwide (Baby et al. 2013; Carola et al. 2013; Carrera & Muñoz 2013; Iaffa et al. 2013; Kober et al. 2013; Moretti et al. 2013; Nemčok et al. 2013; Teixell & Babault 2013; Fig. 8a–d) and by considering the systematic data collection from the Northern Andes, an example of the forceful retro-wedge scenario (Bayona et al. 2013; Caballero et al. 2013a, b; Hermeston & Nemčok 2013; Jimenez et al. 2013; Mora et al. 2013; Moreno et al. 2013; Silva et al. 2013; Teixell & Babault 2013; Tesón et al. 2013; Fig. 8a). This allows an assessment to be made of all the factors controlling the energy balance.

The natural laboratory of the Eastern Cordillera, Colombia

The advantage of the Eastern Cordillera natural laboratory is the relatively small amount of total shortening and the different stages of inversion development along its strike from south to north (see Hermeston & Nemčok 2013). This allows one to see the details of the transition from initial inversion to full inversion. Serial sections through the crustal distribution of earthquake locations (Hermeston & Nemčok 2013) change along the strike. They indicate an ongoing strengthening in the internal Eastern Cordillera in its northern portion that changes to fault activity at both its flanks and no internal strengthening in its southern portion. This resembles lateral sand compaction in sandbox experiments before the appearance of the first, large-displacement faults (see Koyi 1995, 1997). By analogy, the internal deformation of the axial flat region of the Eastern Cordillera has served the same purpose since the Paleocene (see the timing in Mora et al. 2013; Hermeston & Nemčok 2013), before the onset of intensive deformation along the Eastern Cordillera flanks, which was rapid during the Oligocene–Early Miocene and accelerated during the Late Miocene–Recent (Mora et al. 2013; Silva et al. 2013).

A closer inspection of the southernmost portion of the Eastern Cordillera indicates that there is no axial flat zone present there. On the contrary, this portion of the orogenic belt resembles a doubly-sided orogen. Its transition to the rest of the Eastern Cordillera, which does contain the axial flat zone, lies inside ‘swath 7’ of Hermeston & Nemčok (2013; their Fig. 3). This Eastern Cordillera segment matches well with a segment of the Llanos foreland basin that does not have a record of flexing starting with deposition of the Carbonera Formation (42–16 Ma). Furthermore, this is also the only segment of the Llanos foreland basin that records initiation of the shortening associated with basement-involved faults.

Challenges and techniques of the Eastern Cordillera studies

With regard to movement rates of individual plates controlling the Eastern Cordillera development history, it is interesting to compare ‘The tectonic acceleration since Late Miocene’ noted by Silva et al. (2013) and ‘The latest Miocene peak shortening rates and out-of-sequence reactivation of the main inversion faults, coeval with slower-slip footwall edge short-cut faults’ documented by Mora et al. (2013) with the intra-Middle Miocene increase in Mid-Atlantic ridge sea-floor spreading rate (Cobbold 2011, pers. comm.). Even more interesting is the outcome of the general discussion at the Barichara conference, which concluded that the most impressive thick-skinned terrains along the entire strike of the Andes are not necessarily correlated with flat-slab segments of the subducting Nazca Plate, but can be frequently correlated with segments having a steeply dipping subducting slab.

The role of crustal (lithospheric) thickness variations on the Eastern Cordillera development was discussed at the Barichara meeting by Saylor, who noted a change at the Oligocene–Miocene boundary from the eastward propagating deformation front to deformation fronts propagating from both sides of the Eastern Cordillera. We may speculate that the cause may be the arrival of increased foreland plate buoyancy associated with the thicker crust of the foreland plate arriving at the convergence zone. However, following Hermeston & Nemčok (2013), Mora et al. (2013), Saylor (conference talk) and Silva et al. (2013), it appears that the
(a) Basal boundary conditions

(b) Internal properties

(c) Boundary conditions on thickened crust
Eastern Cordillera underwent pulses of deformation migration, characterized by:

(1) the eastward arrival of the deformation front into the region of the future Eastern Cordillera;
(2) the subsequent internal deformational strengthening of the Eastern Cordillera material; and
(3) the eastward movement of deformation after the buoyant foreland plate stops flexing, this being accompanied by
(a) the orogen deformation intensity increasing,
(b) development of the mountain building asymmetry with pronounced building of the eastern flank relief,
(c) propagation of the detachment fault beneath the frontal thin-skin thrust belt, and
(d) development of new basement-involved thrust blocks in the proximal Llanos basin.

Two techniques have been used to investigate the timing of these deformation events. The first, a low-temperature thermo-chronology method, is

Table 1. Factors controlling the orogenic deformation

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<th>Factor group</th>
<th>Main factor</th>
<th>Factor</th>
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<td><strong>Main engine</strong></td>
<td>Overall convergence rate</td>
<td>Climate</td>
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<td>Overall plate movement rate</td>
<td>Mean annual precipitation</td>
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<td>Orogen movement sense with respect to mantle flow</td>
<td>Mean annual temperature</td>
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<td>Pro-wedge v. retro-wedge location of the thick-skin region</td>
<td>Relief</td>
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<td><strong>Internal factors</strong></td>
<td>Rheologies of deforming layers</td>
<td>Elevation</td>
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<td>Existence v. non-existence of potential detachment horizons</td>
<td>Uplift rate</td>
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<td>Occurrence of intrusions</td>
<td>Rock resistance to erosion</td>
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<td>Presence of basement buttresses</td>
<td>Elastic thickness of the foreland plate</td>
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<td>Crustal (lithospheric) thickness variations</td>
<td>Orogen taper</td>
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<td>Thermal regime</td>
<td>Orogen advance rate</td>
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<td>Inherited strength contrasts</td>
<td>Effectiveness of denudation</td>
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<td>Pre-existing anisotropy</td>
<td>Effectiveness of sediment transport into the basin</td>
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<td><strong>External factors</strong></td>
<td>Syn-tectonic erosion</td>
<td>Effectiveness of the sediment distribution system inside the foreland basin</td>
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<td></td>
<td>Syn-tectonic deposition</td>
<td>Gravity resistance against further shortening owing to increasing potential energy of the orogen</td>
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Fig. 7. External and internal factors controlling orogenic deformation (Ellis & Beaumont 1999). (a) Velocity conditions contain pro-mantle lithosphere converging and subducting at uniform velocity $v_p$ and retro-mantle converging at uniform velocity $v_R$. Mantle subduction point S moves with velocity $v_S$. Its positive v. negative values represent subduction zone advance v. retreat. (b) Internal model properties include rheologies of deforming layers, detachment horizons, intrusions and inherited strength contrasts. (c) External factors include erosion and deposition modifying the load distribution. Thickened crust grows against increasing gravity resistance because of its increasing potential energy. Loads and thickened layers are flexurally compensated by paired broken elastic beams below the model layer.
Fig. 8. Locations of study areas of Special Publication contributions on (a) South American, (b) Asian, (c) African and (d) European continents. Study areas include: 2, Kober et al. (2013); 3, Moretti et al. (2013); 4, Baby et al. (2013); 5, Carrera & Muñoz (2013); 6, Iaffa et al. (2013); 7, Carola et al. (2013); 8, Teixell & Babault (2013); 9, Nemčok et al. (2013); 10, Jimenez et al. (2013); 11, Moreno et al. (2013); 12, Tesón et al. (2013); 13, Bayona et al. (2013); 14, Caballero et al. (2013a); 15, Caballero et al. (2013b); 16, Mora et al. (2013); 17, Silva et al. (2013); 18, Hermeston & Nemčok (2013).
well suited to portions of the Eastern Cordillera where syn-tectonic strata are absent. The second, interpretation of syn-tectonic strata from outcrops and reflection seismic images, is useful in portions of the Eastern Cordillera with syn-tectonic strata and adjacent flexural basins. Interestingly a discrepancy in the timing determined using the two methods can sometimes occur. Because of this, the low-temperature thermo-chronology methods used in the Eastern Cordillera were typically combined with cross section balancing. The grid of balanced cross sections (see Mora et al. 2013; Tesón et al. 2013) includes several trans-Cordillera profiles. Long regional profiles have allowed a comparison to be made of the timing of deformation events derived from thermo-chronology in blocks without syn-tectonic strata and the timing derived from analysis of syn-tectonic strata. Blocks without syn-tectonic strata occur in the central portion of the Eastern Cordillera.
Syn-tectonic strata are preserved on both sides of the Cordillera and in adjacent proximal portions of the flexural basins. They were studied either in long regional balanced cross-sections or in a large grid of short, balanced cross-sections cutting through the frontal structures on both sides of the Eastern Cordillera (see Caballero et al. 2013a, b; Jimenez et al. 2013; Mora et al. 2013; Moreno et al. 2013; Tesón et al. 2013).

The development history of thick-skin provinces starts with initial inversion, characterized by distributed deformation affecting broad regions and resulting in low structural relief. It ends with full accretion, characterized by the deformation localized along margins of the thick-skin province and resulting in a high relief. Initiation, termination and eventual shift of this deformation into a different regime is controlled by three groups of factors (Table 1), including the main engine (i.e. main convergence-driving), and internal and external factor groups. Our conceptual model for complete thick-skin province development and its control by the factors listed in Table 1 presents the conceptual framework for an understanding of thick-skin orogenic deformation.

The following 18 papers in this Special Publication aim at different aspects of this model, with special focus on the individual controlling factors. They are arranged into two main groups, both containing predominantly field-based case studies.

The first and smaller group of papers draws from the data gathered on thick-skin provinces located anywhere, except the Northern Andes of Colombia (Fig. 8a–d). Field-based studies determining the role of pre-existing anisotropy in thick-skin orogenic deformation are presented in papers by Baby et al. (2013), Carrera & Muñoz (2013), Iaffa et al. (2013) and Kober et al. (2013), while the role of the inherited strength contrasts is studied by Morretti et al. (2013). The effect of the existence v. non-existence of potential detachment horizons on a large scale is discussed by Carola et al. (2013), and on a smaller scale is discussed by Nemčok et al. (2013). The role of buttressing effects of basement blocks is documented by Carrera & Muñoz (2013) and Iaffa et al. (2013). A detailed description of the structural architecture development from initial inversion to full accretion stages is given by Teixell & Babault (2013).

The second and larger group of papers draws from the data gathered on the Northern Andes of Colombia (Fig. 8a). The role of pre-existing anisotropy in thick-skin orogenic deformation is presented in papers by Bayona et al. (2013), Silva et al. (2013) and Tesón et al. (2013). The role of crustal (lithospheric) thickness variations in thick-skin orogenic deformation is discussed by Hermeston & Nemčok (2013). The role of potential detachment horizons in thick-skin orogenic deformation is documented by Moreno et al. (2013), who also study the effect of thickness variations of the syn-tectonic sedimentary wedge. The detailed documentation of ‘detachment linkages’ in the serial balanced cross sections of Jimenez et al. (2013) studies the effect of the rheologies of deforming layers. External factors controlling orogenic deformation are discussed by Silva et al. (2013), Caballero et al. (2013a, b) and Jimenez et al. (2013), with the first group studying the role of syn-tectonic erosion and the remaining groups studying syn-tectonic deposition.

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