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“The greatest mistake you can make is to be continually fearing you will make one.”

Elbert Hubbard
Information on the kinematics, magnitude, and timing of vertical movements can be obtained in regional studies by means of basin analysis and structural approaches. An overview of the methods used in this Thesis is given in this chapter. Some of these methods are used in several chapters and are here presented as they are used in this manuscript.

### 2.1 Reflection seismsics

Since Mohorovičić’s discoveries at the beginning of the last century, seismic analysis has deeply contributed to increase the bulk of geological knowledge of both deep and near surface features [Prodehl et al., 2013]. Refractions and reflections of seismic waves occur at geologic interfaces with contrasting acoustic impedance. The pertinent processing of the seismic signal allows close-to-real reproductions of the subsurface structure. The raw seismic images of the lines used in this Thesis are shown in Appendix A.

#### 2.1.1 Analysis and interpretation

Seismic reflection analysis is used to map and interpret the subsurface geology (normally ~4 – 5 km, up to ~10 km for deep reflection), both in academia and especially in industry. Even though the large-scale high resolution 3D surveys that hydrocarbon companies use are normally lacking in academia, very relevant scientific advancements have been accomplished thanks to reflection seismic data.

At its most basic, seismic interpretation consists of the distinction, tracing, and correlation of relevant reflections. These reflections are then used to obtain a geological interpretation. Seismic interpretation and attribute analysis (see below), in combination with other methods, allow the construction of geological models for the areas of interest. However, the often low quality of the processed images, a consequence of poor processing, or the presence of highly reflective bodies, such as salts or volcanics, and the inherited low vertical and horizontal resolution of the seismsics, leads to a relevant degree of uncertainty in the seismic interpretation. This uncertainty can be minimized by carefully studying the seismic signal, by recognizing seismic facies that allow for jump-correlation, or through depth-conversion and balancing techniques with which the geometrical validity of a section is checked. Seismic data is, therefore, a strong, valid data if the interpretation of the seismic attributes is correct.

#### 2.1.2 Seismic facies

All the seismic lines used in this Thesis were provided as either .tiff or .png images and converted to SegY format for interpretation. The SegY transformer tool from the GeoSuite AllWorks® software suite, provided by Dr. Fabrizio Pepe, from Palermo University, was used to perform the operation. The interpretation of the SegY files enhances seismic facies analyses. Time-horizon attributes, such as coherence, frequency, dip/azimuth, and curvature of the reflections allows for the definition of distinctive packages as a function of their seismic qualities. Proper definition of seismic facies attest to the seismic unit definition, even when lateral unit disruptions occur. Thus, a more solid interpretation can be made thanks to the discrimination of the seismic facies.
2.1.3 Time-to-depth conversion

Geological interpretation of seismic sections in time is often misused for geometrical and structural analysis. This type of analysis carries the inherit risk of assuming a constant velocity model in which even simple geology can produce false highs, or obscure true ones [Etris et al., 2001]. On the other hand, depth-converted seismic sections eliminate the time-travel effects, reducing the uncertainties and recreating the real structure. In order to avoid time-related artifacts on the seismic lines presented in this Thesis the acoustic wave travel time is converted to actual depth on the basis of the acoustic velocity of the media. The seismic lines were converted to true-depth using the GeoSuite AllWorks® software and the velocity values for each unit (shown in Figs. 3.3 and 5.2). Depth-conversion permits the proper evaluation of the observed geometries and the production of depth and isochore maps, both of which are key to a real assessment of the tectono-sedimentary evolution of an area.

2.1.4 Seismic-to-well tie

Seismic-to-well matching is a common technique to link the different units of interest to the seismic signal, therefore assigning geologic value to the reflections. In most cases, this so-called “tie” is made at the well site by measuring some physical parameters, such as density log, sonic log, or VSP, and knowing other parameters used during the signal processing process (i.e., wavelet).

One singular well that was available for analysis is presented in Chapter 3. This well only contained a (partial) sonic log -and no other borehole input– thus making the construction of the synthetic seismogram highly speculative. Therefore, the mean interval velocities were obtained from the sonic log (slowness velocities). This possibility is considered to be the most reliable geologically if one considers that Dix equations are only valid for homogeneous isotropic low-steepness layers [Dix, 1955] and the non-uniqueness attached to the construction of velocity models [Al-Chalabi, 1997a, b; Reilly, 1993]. Using standard velocity values extracted from the sonic log is a broad approach but it might be especially valid for the sequence of interest (i.e., the upper part of the section), as variations in density or in compaction by overburden should be minor. The mean average interval velocities were obtained assuming constant velocities in each depth interval, and they are defined taking into account not only the slowness velocities but also the lithology-defined standard velocities, calculated as lithology percentages per unit, with special consideration for those intervals with high calcareous content (higher velocity values) [Bourbie et al., 1992]. Even though this estimation is essentially rough, the obtained values are quite consistent with the values found in [Aydemir and Ateş, 2006a, b].

2.2 Analysis of tectonic motions

2.2.1 Subsidence curves - Downgoing vertical motions

Vertical motions, specially subsidence, can be quantified by means of “backstripping”. The backstripping technique is used to estimate the depth that the basement would reach in the absence of sediment and water loading [?]. To produce the backstripped subsidence curves, information on palæobathymetry and absolute sea level
fluctuations is needed [e.g. Wood, 1981] and the measured thicknesses should be de-compact [Steckler and Watts, 1978]. The depth of deposition is usually achieved with detailed micropalæontological analysis and stratigraphic markers [e.g. Mitchum et al., 1977] or using the lithostratigraphic characteristics of a section or log. The possible range of water depths is determined using constraints from geological criteria [Wood, 1981], the type of fauna present in the section, and/or local or global sea level curves [e.g. Haq et al., 1987]. Decompaction during backstripping allows the sedimentary units to expand to their original depositional thickness [Steckler and Watts, 1978]. For all the subsidence curves shown in this Thesis, with the exception of those taken from Flecker et al. [1998], mechanical compaction is assumed and the approach by Steckler and Watts [1978] is performed, using standard mean, maximum and minimum porosity-depth functions as in Bond and Kominz [1984a]; Bessis [1986a]; Mavko [1981] (see Table 2.1), and average decompaction values [Sclater and Christie, 1980].

### Table 2.1: Standard porosity-depth relationships

<table>
<thead>
<tr>
<th>Lithology</th>
<th>( \varphi_1 )</th>
<th>( c_1 )</th>
<th>( \varphi_0 )</th>
<th>( c_0 )</th>
<th>( z_p )</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Maximum porosity curve</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sand</td>
<td>0.29</td>
<td>0.216</td>
<td>0.4</td>
<td>0.51</td>
<td>1.0</td>
</tr>
<tr>
<td>Silt</td>
<td>0.42</td>
<td>0.375</td>
<td>0.6</td>
<td>1.0</td>
<td>0.5</td>
</tr>
<tr>
<td>Shale</td>
<td>0.5</td>
<td>0.475</td>
<td>0.7</td>
<td>1.1</td>
<td>0.5</td>
</tr>
<tr>
<td>Carbonate</td>
<td>0.52</td>
<td>0.442</td>
<td>0.78</td>
<td>1.33</td>
<td>0.5</td>
</tr>
<tr>
<td>Halite</td>
<td>0.0</td>
<td>0.1</td>
<td>0.0</td>
<td>0.1</td>
<td>0.0</td>
</tr>
<tr>
<td>Anhydrite</td>
<td>0.0</td>
<td>0.1</td>
<td>0.0</td>
<td>0.1</td>
<td>0.0</td>
</tr>
<tr>
<td><strong>Minimum porosity curve</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sand</td>
<td>0.2</td>
<td>0.48</td>
<td>0.2</td>
<td>0.48</td>
<td>0.0</td>
</tr>
<tr>
<td>Silt</td>
<td>0.25</td>
<td>0.325</td>
<td>0.25</td>
<td>0.325</td>
<td>0.0</td>
</tr>
<tr>
<td>Shale</td>
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<td>0.47</td>
<td>0.53</td>
<td>1.05</td>
<td>0.5</td>
</tr>
<tr>
<td>Carbonate</td>
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<td>0.58</td>
<td>0.2</td>
<td>0.58</td>
<td>0.0</td>
</tr>
<tr>
<td>Halite</td>
<td>0.0</td>
<td>0.1</td>
<td>0.0</td>
<td>0.1</td>
<td>0.0</td>
</tr>
<tr>
<td>Anhydrite</td>
<td>0.0</td>
<td>0.1</td>
<td>0.0</td>
<td>0.1</td>
<td>0.0</td>
</tr>
</tbody>
</table>

In order to correct the effect of compaction in the sedimentary column:

1. Every stratigraphic unit defined is divided in lithologic fractions of sand, shale, silt, carbonate, halite, and anhydrite. These lithologic fractions together with the standard values of density and surface porosity shown in Table 2.1, allow the definition of:

   a. Porosity as a function of depth,

   \[
   \varphi_i = \varphi_0 \cdot e^{-c_0 z}, \quad \text{if } 0 < z < z_p, \tag{2.1}
   \]

   \[
   \varphi_i = \varphi_1 \cdot e^{-c_1 z}, \quad \text{if } z_p < z < \infty, \tag{2.2}
   \]

   in which \( \varphi_i \) is the porosity of the layer \( i \) at the new depth \( z \) (in km). The \( \varphi \) and \( c \) are characteristic constants of initial porosity and depth for each lithologic fraction that define a piecewise function with two intervals: shallow (\( \varphi_0, c_0 \)) and deep (\( \varphi_1, c_1 \)). The \( z_p \) is a constant depth value that delimits the shallow and deep regions (check Table 2.1).
b. Density as a function of depth,

\[ \rho_i = (1 - \varphi_i)\rho_0, \]  
(2.3)

where \( \rho_0 \) is the initial density (in g/cm\(^3\)) and \( \varphi_i, \rho_i \) are the porosity and the density of the \( i^{th} \) later at the specific depth of each lithologic fraction.

Proceeding in this manner, one-by-one decompression of each package was subsequently produced along the minimum and the maximum porosity-depth curves.

2. The density of the entire sedimentary column, \( \rho_t \), is calculated by means of:

\[ \rho_t = \sum_i \left\{ \frac{(1 - \varphi_i)\rho_i}{S} \right\} y'_i, \]  
(2.4)

with \( y'_i \) as the thickness of the \( i^{th} \) layer and \( S \) as the total thickness of the column corrected for compaction.

3. Finally, the depth of the basement corrected for the sediment load is calculated assuming Airy isostasy (\( Y \)), via:

\[ Y = S \left( \frac{\rho_m - \rho_t}{\rho_m} \right), \]  
(2.5)

in which \( \rho_m \) is the density of the compensated material, considered to be 3.3 g/cm\(^3\).

Subsidence analysis plots depict the vertical motion history of a given stratigraphic horizon from the moment of its deposition at that location and with respect to a datum [e.g. Van Hinte, 1978]. If the picked horizon corresponds with the first deposits of a basin, then the subsidence plots yields the vertical motions that basins undergo during their formation and in the posterior stages of their development. Isostatic reactions to lithospheric processes such as thickness variations, loads, or thermal events are recorded in the subsidence history of a basin, which in turn allows suggestions concerning the tectonic driving mechanism [e.g. Xie and Heller, 2006] and/or determination of the basin-scale fault activity [Escalona and Mann, 2011]. With secondary deviations of the tectonic-setting, specific trends occur in relation to sea level and/or sediment or tectonic loading variations. After backstripping, the subsidence curve plots represent the idealized subsidence of a basin that would have existed if it were only filled with water [Xie and Heller, 2006].

Some of the more relevant uncertainties of the method are attached to the palæowater and palæodeposition depth as well as those attached to the age constraints of the chosen basement or that of the stratigraphic horizons. Studying thick stratigraphic successions deposited in relatively shallow water depths reduces water-depth uncertainties since the possible errors become small when compared with the magnitudes of the curves [Xie and Heller, 2006]. On the other hand, palæowater depth differences implicit in bathyal-abyssal deposits may bypass relevant signals [Dickinson et al., 1986]. Even so, the subsidence history technique has been widely
used for more than three decades now and is still crucial to determine the driving causes related to sedimentary basin development [e.g. Bertotti, 2001]. All the back-stripped subsidence curves presented in this Thesis were constructed with in-house software BMOD developed by Dr. Fred Beekman.

### 2.2.2 Palinspastic cross-sections - Horizontal motions

Cross-section balancing is a technique used for the geometrical validation of a geological interpretation. Cross-section balancing is based on the assumption that the cross-sectional area of post-depositional sedimentary bodies is equal before and after deformation [e.g. Mitra and Namson, 1989; Dahlstrom, 1969; Elliott, 1983]. The related technique of cross-section restoration is used for the quantitative analysis of horizontal deformations. Valuable inputs on the amounts of extension/shortening and on the displacement along the main structures can be achieved by retro-deformation of the cross-section to the original pre-deformational state [e.g. Marshak, 1988]. Restorations that do not balance show incompatibilities, unrealistic structures, or changes in area that expose errors in seismic interpretation or incorrect restoration parameters [e.g. Schultz-Ela, 1992].

In this Thesis, the equal-area balancing technique was used for cross-section restoration. This method is relatively simple and effective and was performed manually. There is still no consensus on which of the multiple restoration procedures is best [see Hauge and Gray, 1996, for details]. Therefore, the mentioned manual equal-area restoration performed in this Thesis is considered as valid as any of the possible algorithms that could have been used in modern software, such as Midland Valley 2DMove™.

To perform this restoration, the post-deformational cross-sectional area of the faults hangingwalls is measured and preserved during the displacement of the rigid body along the fault planes with fixed geometries. The retro-deformation occurs layer-by-layer, in an opposite sense to the transport direction and with a specific shear angle, normally perpendicular or close to perpendicular to the fault plane. In addition, flat topography at the surface was assumed. The method implies several relevant assumptions, such as non-penetrative deformation, the absence of pressure-solution events, or a lack of bedding-plane slip. Other sources of error are those specifically related to restoration in extensional terrains. The most relevant here is the absence of volume changes caused by compaction (the most important element of the finite strain with the exception of brittle faulting in extensional regimes [Gibbs, 1983]), which could account for volume changes extending to 40% [see Wood, 1981]. Yet the analysis of palinspastic restorations have proven valuable in understanding the history of horizontal deformation undergone by any specific area [e.g. Bertotti et al., 1993].

### 2.3 Fieldwork

Fieldwork is an essential component in geological studies. Structural-geological fieldworks are very limited in the area and time frame studied by this Thesis, due to the nature and the relatively minor deformation of the Tertiary rocks in the area. Regional integration of these contributions into a larger tectonic framework or with
other quantitative techniques are even sparser. The structural fieldwork campaigns for this Thesis were located in the Miocene marginal basins in the south and southwest of the Central Anatolian Plateau (CAP) (see Chapter 4). Key outcrops and A-quality attitude data obtained in the field is shown in Appendix B. The sediment infill of these basins recorded their geohistory prior to and during the growth of the SCAP, and thus that of the development of the modern Taurus Mountains. This Thesis aims to achieve three main goals during fieldwork:

- Determination of the character and geometry of the contact between the Miocene deposits and the basement palæorelief as well as the post-/syn-/pre-tectonism in these rocks. The nature of this contact and the geometry of the sedimentary packages, especially in the form of sedimentary wedges, provide relevant information on the wavelength of the vertical movements and on the age of deformation. Theoretically, there are three geometrical end-members in the development of a sedimentary wedge (see Fig. 2.2). Onlap relationships are post-tectonic, in which they indicate a time gap and tilting between two periods of deposition. Syntectonic wedges show ongoing deformation during deposition, which allow the determination of the age of the deformation. Pre-tectonic deposition of the sedimentary sequence develops no wedge.

- Analysis of strain and stress in these basins. Structural analyses of the Miocene infill is key to determine the present strain state and estimate the palæostress situation during basin development, thus understanding the general tectonic regime. Data such as fault slip data provide information on palæostress directions. In this sense, the “conjugate fault sets” deserve special mention since they provide the best constraints on the determination of the palæostresses. In conjugate sets, the main stress axis is determined by the bisector of the low angle between two conjugated faults (see Fig. 2.1). Some of the pitfalls of the palæostress determination method include the fact that striations and other kinematic indicators only record one or two episodes of fault movement and that stress directions can substantially vary from strain directions.

- Recognition of the possible structures accommodating the vertical motions, both upward and downward motions, undergone by the basins during Miocene and posterior times. Km-scale vertical displacements should have been accommodated by regional-scale structures. The appearance, character, and orientation of these structures provide relevant insights into the possible causes behind the vertical motions. Extensional south-dipping faults in the marginal basins of the southern margin of the Central Anatolian Plateau would indicate “passive” adjustment to uplift, possibly as a response to lithospheric detachment and isostatic rebound or thermal effects such as those linked to asthenospheric upwelling. In contrast, north-dipping shortening structures in the area would indicate “active” deformation, leading to crustal/lithospheric thickening and, therefore, surface uplift. The main challenge is then to recognize these large scale structures.

The combined study of the mentioned features and that of the tilting or plunging of structures will lead to a detailed tectonic evolution, focusing in vertical motions.
2.4 Thermo-mechanical finite element modelling of accretion

A compilation of available geologic data, integrated with the data presented in this Thesis, are used as constraints for 2D thermo-mechanically coupled numerical finite element models (FEM) of a transect from Central Anatolia to the Levantine Basin. The numerical models are used to understand the uplift mechanisms in the southern margin of the Central Anatolian Plateau (CAP), and its association with the accretionary complex of the central Cyprus arc (see Chapter 6). Concretely, the models are used to test a hypothesis in which south Turkey is the forearc high of the central Cyprus subduction system, which drives the surface uplift undergone by this area.

The thermo-mechanical FEM used in this Thesis were initially used to model orogenic wedges [Willett et al., 1993; Pope and Willett, 1998a], and then adapted to model subduction wedges, i.e., to account for the accretion of a thin sedimentary layer, flexural isostasy, and frictional stick-slip boundary conditions [Fuller, 1996; Fuller et al., 2006a]. Later, it was modified to include variable sedimentation rates,
strain softening (and healing), and sediment-tracking capabilities [Cassola and Willett, 2013a, b, in prep.]. A general description of the model, extracted exclusively from the above mentioned publications, is given below. The reader is referred to those contributions for more details.

### 2.4.1 Model formulation

The FEM simulates subduction assigning kinematically defined motions to the subducting slab and mantle, and allows the overriding crust to deform dynamically in response to the boundary conditions on its edges. For this, two domains are defined, a mechanical and a thermal domain.

The mechanical domain, which comprises the overriding crust and the pile of incoming accreting sediments, extends from the surface until the base of the overriding crust on the landward side until the top of the subducting crust on the seaward side of the model. In the mechanical domain, the model responds to a quasi-static creeping flow with a plastic yield criterion [Zienkiewicz and Godbole, 1974]. The flow equation of a Newtonian viscous material with laminar flow, when described in an Eulerian reference frame, can be expressed with a simplified Stokes equation:

\[-\frac{\partial \bar{\sigma}}{\partial x_j} + \mu \frac{\partial}{\partial x_i} \left( \frac{\partial \nu_i}{\partial x_j} + \frac{\partial \nu_j}{\partial x_i} \right) + \delta g_j = 0,\]

where \( \sigma \) is the mean stress (pressure), \( \mu \) is the viscosity, \( \nu \) is the velocity, \( \delta \) is the density, \( g_j \) is the acceleration of gravity in the \( x_j \) direction, and \( i, j = 1, 2 \). The mean stress, \( \bar{\sigma} \), is expressed as:

\[\bar{\sigma} = -\frac{\sigma_{kk}}{2}.\]

where the summation is implied over repeated indices.

To solve velocity and pressure in equation 2.6, we need to account for mass conservation:

\[\frac{\bar{\sigma}}{\kappa} + \frac{\partial \nu_i}{\partial x_i} = 0,\]

where \( \kappa \) is used to reduce compressibility to near zero [Bathe, 1996]. A non-linear power-law viscosity equation for high temperature lasted deformation is used in the models to relate stress to the rate of deformation:

\[D_{ij} = A_\mu \sigma_{ij}^n \mu \exp \left( \frac{-Q}{RT} \right),\]

where the \( Q \) is the activation energy, \( R \) the molar gas constant, \( T \) the temperature, and \( A_\mu \) and \( n_\mu \) are constants dependent on the material. The rate of deformation, \( D_{ij} \), can be expressed as well as:

\[D_{ij} = \frac{1}{2} \left( \frac{\partial \nu_i}{\partial x_j} + \frac{\partial \nu_j}{\partial x_i} \right).\]

From equations 2.6 and 6.1 can be obtained that:

\[\mu_{nlv} = I_2' \left( \frac{1-n_\mu}{2n_\mu} \right) A_\mu \frac{1}{n_\mu} \exp \left( \frac{Q}{n_\mu RT} \right),\]
where the $I_2'$ is the second invariant of the deviatoric rate of the deformation tensor. A Coulomb failure criterion, in which the Coulomb yield stress is:

$$\sigma_y = c \cos(\varphi) + \frac{1}{3} J_1 \sin(\varphi),$$

is used to simulate the frictional-plastic behavior of the material, which undergoes plastic failure if:

$$\sqrt{I_2'} \geq \sigma'^y,$$

with $c$ being the cohesion and $\varphi$ the internal friction angle of the material, $J_1$ the first stress invariant and $I_2'$ is the second invariant of the deviatoric stress. Calculation of an effective viscosity for use in equation 2.6 that satisfies the Coulomb failure criteria is done when plastic failure occurs:

$$\mu_{eff} = \frac{c \cos(\varphi) + \frac{1}{3} J_1 \sin(\varphi)}{2 \sqrt{I_2'}}$$

An arbitrary Langrangian-Eulerian finite element formulation [Fullsack, 1995; Willett and Pope, 2004] is used to solve the unknown velocities and pressure in equations 2.6 and 2.7 in a Eulerian grid of quadrilateral elements.

The upper stress-free surface updates (at every time step) to take into account tectonic advection of the surface, erosion, and sedimentation. A passive Lagrangian mesh allows the tracking of integrated deformation.

### 2.4.2 Boundary conditions

The model reproduces a subduction system in an accretionary margin (Fig. 2.3). The simulation is governed by boundary conditions on velocity and frictional stick-slip along the domain limits. The plates converge by a horizontal velocity, $\nu_c$, which is set in the left side of the model, while the right side is fixed in space. The “S” point, being the intersection point where the subducting slab meets the continental Moho, divides a left domain with a frictional stick-slip boundary condition, and a right domain with a zero tangential velocity boundary condition.

This set of boundary conditions allows the calculation of the slip in the interface between the subducting and overriding plates. This slip depends on the shear stress at the plate interface, $\tau$, and the shear strength of the interface, $\tau_c$, defined as:

$$\tau_c = \sigma_n \tan(\varphi_b),$$

where $\sigma_n$ and $\varphi_b$ are the normal force and friction angles at the plate interface, respectively. No slip at the interface occurs when $\tau < \tau_c$, and the overlying material is then “locked”. Contrarily, if $\tau > \tau_c$, slip occurs. The amount of slip is found iteratively by applying $\tau = \tau_c$, i.e. the shear stress on the interface equals the shear strength. This boundary conditions entails the off-interface remaining at $\nu_c$ to drive the deformation within the model.
2.4.3 Relevant subroutines

Flexural isostasy

Elastic deformation is excluded from the mechanical model and elastic plate flexure is explicitly included for the entire model. The plates respond isostatically to crustal loads by vertical motions, in which the deflection of the plate is calculated analytically. That is, the weight of the crust is applied as a load on both plates, which behave as semi-infinite, or broken, elastic plates [Hetényi and Hetbenyi, 1946] coupled at the “S” point to ensure their contact. Thus, the flexural rigidity of the plates and the density of the mantle, crust, and water layer, control the flexural response in both plates [Fuller, 1996].

Sedimentation

At the upper surface of all models, submarine sedimentation simulated. The model assumes that sediments are abundant and have the same material properties as the rest of the crust. Sedimentation takes place instantaneously from sources that are not within the model, first filling the accommodation space provided by the closed depressions, which eventually fills completely, which leads to overfilled basins and by-pass of sediments (“fill-to-spill”). Filling within the basins occurs from the lowest point by advection of points as a function of the inputted accumulation rate [Cassola and Willett, 2013a].

Thermal state

In the thermal domain, the temperature is computed using a finite element method that solves the advective-conductive heat transfer equation with radiogenic heat production:

\[ \rho c_p \frac{\partial T}{\partial t} = \frac{\partial}{\partial x} \left( k \frac{\partial T}{\partial x} \right) + \frac{\partial}{\partial z} \left( k \frac{\partial T}{\partial z} \right) - \rho c_p v_x \frac{\partial T}{\partial x} - \rho c_p v_z \frac{\partial T}{\partial z} + A, \quad (2.16) \]

in which \( \rho \) is the density, \( c_p \) the specific heat, \( T \) the temperature, \( t \) the time, \( k \) the thermal conductivity, and \( v_x \) and \( v_z \) are the horizontal and vertical components of the velocity, respectively. The velocities in the remaining regions are assigned kinematically to ensure the conservation of heat and mass within the model domain. Heat production is included in a layer of thickness \( d_a \), representing the region of the highest concentration of radiogenic elements in the crust. In the upper and lower surfaces, constant temperature boundary conditions are used, with the exception of the area where the subduction lithosphere reaches the base of the model. This area as well as the sides of the model have no heat flux. To simulate the thermal pre-collision situation in the subduction zone, the thermal model runs for a certain amount of time, \( t_{\text{runup}} \), with a steady initial velocity field.

2.4.4 Model design and parametrization

The overall design of the model aims at representing a standard continental accretionary margin (see Fig. 2.3). The specific parametrization and geometry of this margin is intended to simulate a 550 km-long transect in the central Cyprus subduction system, from farther than south of the Cyprus trench to the Central Anatolia interior basins. All the models presented in this Thesis start with a setting below
sea level, with a column of water of 1000 m. The thickness of the incoming pile of sediments at the trench, $d_a$, is 3 km, and the velocity of convergence, 35 mm/yr. The thickness of the accreted sediments is obtained iteratively; while the velocity of convergence remains constant, the thickness of the incoming sediment accreting in the upper plate, changes for every model, until the resulting cross-sectional area of the model matches that interpreted from geophysical data of the area. The final cross-sectional area varies as a function of the running time of the model, kept as 25 My for all the models, i.e., assumed to start in latest Oligocene. However, the models are scalable; thus comparable results are obtained for margins with half the accretionary flux that started to form in a period twice as long [Fuller, 1996]. No preexisting relief is considered for the vast majority of the model except for the retroward end, in order to set up the retroward bound of the basin, which would otherwise expand outside the limits of the model.

Cohesion, $c$, and internal friction angle, $\phi$, control the material strengths in the model. The friction angle $\phi_b$ manages the strength of the décollement. These parameters are uniform throughout the mechanical domain, which implies the neglect of the strength contrast among the different material that compounds the margin, i.e. strength variation between the unconsolidated sediments in the seaward side of the system and those present in the oldest retroward side. This design allows a simple parameterization [Fuller, 1996]. To ensure numerical stability, a cohesion value of 1000 Pa is used. This value is higher than expected for crustal material strength, yet it has being proven not affect the final results [Fuller et al., 2006b]. The values of the effective friction angles were reduced to correct the effects of pore fluid pressure [Hubbert and Rubey, 1959], which are not explicitly included. Friction angles are chosen as $\phi = 24^\circ$ and $\phi_b = 8^\circ$. Friction angles up to $30^\circ$ for both the model domain and plate interface imply pore fluid pressure ratios within the expected values for accretionary settings [e.g. Fuller, 1996, and the references therein]. Flexural rigidity is set to be $6.4 \times 10^{24}$ N·m. However, changes in up to two orders of magnitude have proven not to produce significant changes in the end result. Crustal, mantle, and seawater densities adopt standard values [Fowler, 1990].