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Nederlandstalige samenvatting (Dutch Summary)

Appendix
In this thesis, two different areas in southeast Asia are focussed upon: South Peninsular Thailand and South(east) India, which are separated by the Indian Ocean and the intervening Andaman (back-arc) Basin at the Thai margin. These two study areas were chosen in order to improve our understanding of the evolution of the passive margins of this Oceanic Basin and the convergence and further post-collisional evolution of India/(Eur)Asia. It was an additional goal of this work to evaluate the interplay of lateral extrusions or escape tectonics in the framework of wide scale continental convergence on one hand and the extensional settings of back-arc (and pull-apart) basins on the other. Both geodynamic settings are associated with the India-Asia convergence zone in the Mesozoic and the associated effects of this convergence on the exhumation of various basement blocks along this zone.

Thailand, which is a part of Mainland Southeast Asia, can geographically be divided in a northern continental part and a southern peninsular part. The northern continental part of Thailand is bordered by Laos to the north and northeast, by Myanmar to the north and northwest and by Cambodia to the east. The peninsular part of Thailand stretches out to the south and is bordered by the Gulf of Thailand to the east, the Andaman Sea to the west, and by Malaysia to the south. Peninsular Thailand can be further divided into an Upper northern part and a Lower southern part. The latter forms the main area of interest for this thesis.

Peninsular Thailand is dissected by several major fault zones, of which two are important for our study area: the Ranong Fault (RF) and Khlong Marui Fault (KMF) zone. Besides the activity and exhumation along of these fault zone, the Cenozoic tectonic history of Peninsular Thailand was characterised by the opening of the Andaman Basin. The igneous history of this area however dates back to the Mesozoic. The associated granites are part of the well-known Southeast Asia Tin belt, which has been the focus of a host of research concerning their petrogenesis and ore mineralisations, amongst others. For example, Northern Thailand and Upper Peninsular Thailand have been extensively dated by means of the apatite (and titanite) fission track methods in order to unravel their thermochronologic history (e.g. Upton, 1999). However, there appears to be a prominent lack of such research regarding Lower Peninsular Thailand. We aimed to fill this gap by analysing a set of granitic samples along a more or less N-S transect in Lower Peninsular Thailand. The samples were dated by means of the apatite fission-track (AFT) method and when possible their thermal histories were modelled and reconstructed by applying of the HeFTy-software, which reconstructs t-T (time-temperature)-paths based on the AFT-ages, mean track length and the length distribution of apatite
fission tracks. Furthermore, this analysis was supplemented by petrographic and chemical analysis (ICP-OES) of these granitoids. Unfortunately, due to the young AFT-ages of the samples, combined with relatively low U-concentrations in these studied apatite separates, they had low and mostly insignificant amounts of confined tracks for length analysis. Hence, they were not ideal for thermochronologic history modelling.

Additionally and as a comparison to the more active Thai margin of the Indian Ocean-Andaman Basin system, a second study area, i.e. the Tamil Nadu State in Southeast India was investigated with a similar methodology.

India is bordered in the north by the Himalaya orogen and by the Indian Ocean to the east, west and south. Similar to the situation in Upper Peninsular Thailand, AFT-studies have already been performed in northern and especially in western India (e.g. Biswal and Seward, 2003 and Gunnell et al., 2003). Moreover, plenty of research has focussed on the origin of granite and granitoid protoliths in South India (e.g. Santosh et al., 2003), whilst little is known about their denudation history. Therefore this thesis also zooms in on the southern passive eastern Indian continental margin, more specifically the rocks from Tamil Nadu State. This particular area was chosen because even though the studied E-W section was located on the opposite side of the Indian Ocean with respect to the samples from Lower Peninsular Thailand, these samples have known an entirely different tectonic history. For example, these samples were expected to have been influenced by the break-up of Gondwana and the associated opening of the Indian Ocean. Contrary to the Thailand samples, the Indian samples were not expected to have been influenced by the opening of the Andaman Sea. Hence it was expected that the Tamil Nadu samples would unravel a more ancient part of the thermal history of this particular geodynamic association, whilst the Thai samples would shed more light in its more recent evolution.

This thesis consists of eight chapters, which can be roughly divided in a general explanation of the apatite fission track method followed by a literature overview of the study area and the results of the present study. More precisely, chapter 1 highlights the principles of the apatite fission-track method, whilst in chapter 2 the application of this method as a geochronological tool is explained. Chapter 3 outlines the geological and tectonic history of India. In chapter 4 the tectonic and ingenuous history of Thailand and the Andaman region are covered. The analysed samples (and their background) are presented in chapter 5, which also gives an overview of the applied preparation and analytical techniques. The actual results are given in chapter 6 and these are further discussed in more detail and interpreted in chapter 7. The final chapter entails a review of the most important findings. These chapters are followed by a Dutch summary and an appendix.
Chapter 1: The apatite fission track dating method

This chapter deals with the basic principles of the FT dating method as and is largely based on the work by Wagner and Van den haute (1992).

1.1: Fission tracks: definition, structure and origin

In essence, a fission track (or FT) is a narrow trail of damage in a dielectric solid (e.g. a crystal lattice), caused by a process known as nuclear fission. Fission is a specific type of radioactive decay, a process in which nuclides of an unstable mother isotope disintegrate into (more) stable daughters. This disintegration is accompanied by the emission of nuclear particles. Some well-known and more common examples of radioactive decay are α-decay, which is characterised by the emission of an α-particle, i.e. a ²³⁵U-nucleus and β-decay, which is characterised by the emission of an electron. Where fission is concerned, the nucleus of the unstable mother isotope splits into (two) daughter fragments. In general, the reaction is binary (meaning that the mother nuclide disintegrates into two daughter nuclides) and asymmetric (meaning the two daughter nuclides have an unequal mass and atomic number). Moreover, these daughter fragments are often unstable and decay even further by means of e.g. β-emission. Actually, there are two kinds of fission reactions: the spontaneous fission and the induced fission. In the former, fission occurs spontaneously due to the natural instability of the mother nuclide. In the latter, the fission is created by bombardment with particles (e.g. neutrons, protons) or through irradiation with γ-rays. Spontaneous fission is restricted to the heavy nuclides (atomic number Z ≥ 90 and atomic mass A ≥ 230) or to isotopes of the elements from the actinide series.

Each fission reaction is accompanied by the emission of several neutrons and the release of a large amount of energy (usually around 210 MeV). A major part of the energy (approximately 170 MeV) occurs as kinetic energy of the fission fragments, resulting in a high velocity of the fission fragments. As the particles have a strong positive charge, the Coulomb forces are repelling and the two fission fragments will travel through the (dielectric) medium in opposite direction, leaving behind a linear damage trail, or a fission track. The medium in which tracks are formed is called a detector (e.g. minerals, glasses, plastic, ...). There are several formation models, of which the coulomb or ion explosion spike model (Fleischer et al., 1975) is the most widely cited theory (Tagami and O’ Sullivan, 2005). Figure 1.1 shows the three stages of track formation according to the ion explosion spike model. During the first stage (figure 1.1a) a strong positively charged particle strips electrons away from lattice atoms, leaving an array of positively charged lattice atoms behind on its trajectory. This stage is known as the ionisation of lattice atoms. During the second stage (figure 1.1b) the newly
formed ions repel each other due to the Coulomb interaction, creating vacancies and interstitials. During the third and final stage (figure 1.1c) elastic relaxation of the stressed regions results in a more homogenous distribution of local stresses. On the other hand, this will also cause a long-ranged strain of the surrounding undamaged lattice.

Other track formation models, such as the thermal spike model (e.g. Toulemonde, 2000) and the gap or Orsay model (Dartyge et al., 1981) or a combination of several models are however also in use.

![Figure 1.1](image)

**Figure 1.1**: The three stages of the ion explosion spike model: a) ionisation of lattice atoms; b) creating of vacancies and interstitials by means of a displacement of the positive ions due to the Coulomb repulsion; c) straining of the surrounding undamaged lattice due to elastic relaxation of the stressed regions (Wagner and Van den haute, 1992, modified after Fleischer et al., 1975).

Fission fragments lose their kinetic energy, and hence their velocity, due to interactions with the host lattice. The amount of energy loss of a fragment along its trajectory is referred to as the stopping power of the medium. Often, a distinction is made between two mechanisms: electronic stopping power and nuclear stopping power. In the former case, a particle loses energy by means of interaction with electrons, e.g. by stripping electrons from target atoms. In the second case, a particle loses energy by elastic collisions with lattice atoms (Chadderton, 2003; Tagami and O'Sullivan, 2005).

The distance travelled by the fission fragment is the range. The length of the FT equals the sum of the ranges of the two fission fragments. The precise length of a FT depends not only on the energy of the fission fragments, but also on the type of detector and can vary from less than one 1 µm to several mm. The width of a FT is mostly limited to a few nm, meaning FT tend to be submicroscopic. Hence, they are invisible under an optical microscope. Such tracks are called latent tracks. By etching latent tracks with an appropriate chemical etchant, these latent tracks can be revealed and observed (and counted) under an optical microscope (De Grave, 2003).

For example, latent FT in apatite have diameters between 5 to 10 nm (Paul and Fitzgerald, 1992). Observed lengths of etched induced tracks in apatite, on the other hand, are 16.3 µm on average (e.g. Gleadow et al., 1986). As etchant for apatite, a HNO₃-solution is chosen and depending on the concentration of this solution, apatite is etched for 10 to 80 seconds at room temperature (Fleischer and Price, 1964; Wagner and Van den haute, 1992).
Chapter 1: The apatite fission track dating method

In terrestrial minerals such as apatite (Ca\(_5\)(PO\(_4\))\(_3\)(OH,F,Cl)) natural (or spontaneous) fission tracks (AFT) are solely caused by the decay of U and Th isotopes. Of these heavy nuclides, only \(^{232}\)Th and \(^{238}\)U-isotopes occur in measurable concentrations as major constituents in natural substances (Wagner and Van den haute, 1992). Table 1.1 indicates that nearly all FT in natural solids are the result of the fission of \(^{238}\)U, because the other isotopes have a too low abundance \((^{234}\)U\)) or too long half-life \((^{232}\)Th\)) or both \((^{235}\)U\)) to produce significant amounts of FT in comparison to \(^{238}\)U. Note that for samples with unusually high Th/U-ratio’s, it is better to take the contribution of Th into account (Tagami and Nishimura, 1992).

<table>
<thead>
<tr>
<th>Isotope</th>
<th>Relative abundance (compared to (^{238})U)</th>
<th>Total half-life ((T_{1/2} \text{ in a}))</th>
<th>Spontaneous fission half-life ((T_{1/2} \text{ in a}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>(^{232})Th</td>
<td>4*</td>
<td>(1.40 \times 10^{10})</td>
<td>(1.0 \times 10^{11})</td>
</tr>
<tr>
<td>(^{234})U</td>
<td>(5.44 \times 10^{-5})</td>
<td>(2.64 \times 10^{2})</td>
<td>(1.5 \times 10^{16})</td>
</tr>
<tr>
<td>(^{235})U</td>
<td>(7.25 \times 10^{-3})</td>
<td>(7.04 \times 10^{8})</td>
<td>(1.0 \times 10^{19})</td>
</tr>
<tr>
<td>(^{238})U</td>
<td>1</td>
<td>(4.47 \times 10^{9})</td>
<td>(8.2 \times 10^{15})</td>
</tr>
</tbody>
</table>

**1.2 FT revelation**

1.2.1 The principle of chemical etching

The goal of FT revelation is to widen the FT so that they become visible under an optical microscope. Chemical etching is the most widely used visualization technique (Tagami and O’Sullivan, 2005). It is a destructive technique, which both affects the mineral surfaces and crystal defects (Wagner and Van den haute, 1992). Crystal defects such as FT are preferential sites for chemical attacks, meaning the tracks can be widened in this way. The track ends (representing low energy damage) often remain unetched, meaning the entire range of a fission fragment will not be revealed. Hence, the etchable length of a FT tends to be shorter than the actual length of the FT (figure 1.2). This difference is referred to as the range or length deficit (Fleischer et al., 1975).

Etching always starts at the surface of a detector, hence it reveals tracks that intersect the etched surface, these are called surface tracks (Wagner and Van den haute, 1992).
Figure 1.2: The gray line represents the energy loss of a fission fragment along its trajectory (dE/dx). The etchable range or the length over which the track can be etched $R_e$ depends upon the medium and on the specific registration threshold. The range deficit is the difference between the true range ($R_l$) and the etchable range ($R_e$), from Jonckheere (1995).

The geometry of an etched track depends on two quantities: (1) etching rate along the track (or $V_T$) and (2) the bulk etching rate of the etched detector surface ($V_G$). Both quantities depend strongly on etching conditions (such as duration and temperature) as well as on the type of detector. $V_T$ is generally a factor 10 higher than $V_G$ in crystals, giving the tracks a needle-like appearance. This is called the *etch channel* and is the main diagnostic feature of a FT in minerals. Narrow etch channels can open up into a funnel shaped opening at the intersection with the crystal surface, called the *etch pit*. The shape of an etch pit is determined by $V_G$, which in turn depends on the crystallographic structure, anisotropy and orientation of the etched material. For example, an etch pit on the basal plane (perpendicular to the c-axis) in apatite can develop a typical hexagonal symmetry (figure 1.3) (Wagner and Van den haute, 1992).

Figure 1.3: The hexagonal symmetry of etch pits on a basal plane in apatite (Wagner and Van den haute, 1992).
1.2.3: Etch efficiency and track densities

The *etch efficiency* refers to the fact that potentially not all of the tracks that intersect the crystal surface are effectively revealed by the etching process. This etch efficiency can vary from one crystal plane to another, as well as from one direction to another within an etched surface (Wagner and Van den haute, 1992). Fleischer and Price (1964), and Fleischer *et al.* (1975) have tried to quantify this process by means of the *etch efficiency factor* $\eta$ (which depends on the angle at which tracks intersect the surface). Jonckheere and Van den haute (1996) claim that these values are insufficient and introduced a new concept: the *track-counting efficiency* or $\eta q$. Where $\eta$ represents *etch efficiency* and $q$ represents *observer efficiency*. The track-counting efficiency is also defined as the ratio between the measured and true density of etchable fission tracks intersecting the detector surface (Jonckheere and Van den haute, 2002). Besides taking into account the influence of etching conditions, more subjective aspects linked to the observer are also considered. Since both quantities are inseparable and cannot be calculated based on existing models of track revelation and observation, they are determined empirically (Jonckheere and Van den haute, 1996; 2002).

1.2.4 Confined tracks

As mentioned in section 1.2.1, etching usually only reveals tracks that intersect the crystal surface. On occasion, tracks that are entirely located within the crystal (or detector) can also be etched. These tracks are referred to as *confined tracks* and they get etched through a surface track or a natural cleavage (figure 1.4a,b) (Bhandari *et al.*, 1971). Depending on how the etchant reaches them, the confined track are called TINT (*track in track*) or TINCLE (*track in cleavage*) (Lal *et al.*, 1969). These tracks have the advantage that they are etched over their entire length and consequently as latent tracks are temperature dependent, their length distributions can be used as a thermochronologic tool (see section 2.2.2).

According to De Grave (2003), in small samples or samples with low spontaneous track densities (thus samples with a low U-content and/or very young samples), it can be next to impossible to measure a statistically acceptable number of confined tracks for the construction of a length distribution. In these cases, one can try to increase the number of confined tracks by irradiating the sample with a fission source such as a $^{252}\text{Cf}$-source or by heavy ion irradiation (Jonckheere *et al.* 2007 and Min *et al.* 2007). These techniques create artificial channels from the sample surface through which the etchant can penetrate the internal part, augmenting the possibility that confined tracks are etched.
1.3: Principles of the fission track-dating method

1.3.1 The fundamental and practical age equations

The principle of the FT-dating method is based on the spontaneous decay of an unstable mother isotope to a stable daughter at a fixed rate called the decay constant \( \lambda \). In this case, a \(^{238}\text{U}\)-nuclide decays to fission fragments at a fixed (and known) rate \( \lambda_f \) and subsequently, FT will accumulate spontaneously through time. In order to determine a FT-age, not only the number of spontaneous tracks has to be known, but also the current \(^{238}\text{U}\)-concentration. The \(^{238}\text{U}\)-concentration is determined indirectly through a count of induced FT. These induced FT are the result of the induced fission of \(^{235}\text{U}\) in a nuclear reactor. In short, a \(^{235}\text{U}\) nuclei will capture thermal or slow neutrons, become unstable and then fission. The concentration of \(^{235}\text{U}\) can be related to the \(^{238}\text{U}\) content, because the natural \(^{235}\text{U}/^{238}\text{U}\)-ratio, is considered to be a constant value (see below).

So, the FT-age of the sample can be calculated from the ratio of the number of spontaneous tracks to induced tracks (or \(N_s/N_i\)). This forms the basis of the fundamental age equation (equation 1.1) as derived by Price and Walker (1963). For a complete derivation of the equation the reader is referred to Wagner and Van den haute (1992).

The fundamental age equation is given by:

\[
t = \lambda_{\alpha}^{-1} \ln \left[ \frac{\lambda_{\alpha}}{\lambda_f} \frac{N_s}{N_i} \phi_{th} + 1 \right]
\]

1.1

With

\( \lambda_{\alpha} \) = the \( \alpha \)-decay constant \((a^{-1})\), this values takes into account that \(^{238}\text{U}\) does not only disintegrate by means of fission.

\( \lambda_f \) = the fission decay constant \((a^{-1})\)

\( N_s \) = the number of spontaneous tracks (spontaneous track densities per unit of volume)
Chapter 1: The apatite fission track dating method

\[ N_i = \text{the number of induced tracks (induced tracks per unit of volume)} \]

\[ I = \text{the natural } ^{235}\text{U/}^{238}\text{U-ratio, which is considered to be a constant value of } 7.2527 \times 10^{-3} \]

(Cowan and Adler, 1976; Steiger and Jäger, 1977)

\[ \phi_{th} = \text{the fluence of thermal (or slow) neutrons (neutrons cm}^{-2}). \]

\[ \sigma = \text{the conventional effective cross-section of } ^{235}\text{U for thermal-neutron induced fission. This cross-section can be seen as the ratio of the number of neutrons which actually produce a fission reaction per unit of time to } \phi_{th}. \]

Given a certain neutron energy, \( \sigma \) can be considered as a constant value of 570.8 \text{b} (Wagner and Van den haute, 1992) for a specific nuclear reaction. (This cross-section is expressed in barn (b), a unit of area with \( 1 \text{b} = 10^{-24} \text{cm}^2 \)).

In other words, in determining the \( N_s/N_i \)-ratio and the thermal neutron fluence used during irradiation, the FT-age can be derived.

Since \( N_s \) and \( N_i \) are defined as tracks per units of volume, whilst in practice areal or surface densities \( \rho_s \) and \( \rho_i \) are determined, equation 1.1 is further transformed by means of the theoretical relationship between areal and spatial FT-densities (as defined by Fleischer et al., 1975). For more details, see Wagner and Van den haute (1992).

Starting from:

\[ \rho_s = g_s N_s R_s \eta_s q_s \]

\[ \rho_i = g_i N_i R_i \eta_i q_i \]

And by defining the following parameters:

\[ G = \frac{g_i}{g_s} \]

\[ Q = \frac{R_{s_i} \eta_i q_i}{R_{s_s} \eta_s q_s} \]

With:

\( g_{s,i} = \text{geometry factor (1 for an internal surface, 0.5 for an external surface)} \)

\( N_{s,i} = \text{the number of respectively spontaneous and induced tracks per unit of volume} \)

\( R_{s,i} = \text{the average etchable range} \)

\( \eta_{s,i} = \text{the etch efficiency factor} \)

\( q_{s,i} = \text{the observer efficiency} \)

\( G = \text{the geometry ratio} \).
Chapter 1: The apatite fission track dating method

Q = The procedure factor, it encloses several aspects such as observation factors, etching characteristics, the etchable range and (specifically for the ED-method) the difference in length deficit between the mineral and the external detector. Hence, this factor is difficult to determine and age standards are often used to eliminate this factor from the equation, (see section 1.5).

The \( N_s/N_i \)-ratio can now be written as:

\[
\frac{N_s}{N_i} = \frac{\rho_s \, g\, R_s\, q_i}{\rho_i \, g\, R_e\, q_s} = \frac{\rho_s}{\rho_i} \, G \, Q
\]

Combining the former equation yields the following practical FT age equation:

\[
t = \lambda_a^{-1} \ln \left[ \frac{\rho_s}{\rho_i} \, Q \, G \, \phi_{th} \, t + 1 \right]
\]

The geometry ratio \( G \) depends on the type of surface used for counting the spontaneous and induced tracks. Seeing as this study only uses the external detector method (section 1.4.1.), this value will be set to 0.5 (Gleadow and Lovering, 1977). Based on this equation, the age of a sample can be calculated if \( \rho_{s,i}, \phi_{th} \) and \( Q \) are known, seeing as the remaining parameters are empiric or natural and physical constants:

- \( \lambda_a = 1.55125 \times 10^{-10} \text{ a}^{-1} \) (Jaffey et al., 1971; Steiger and Jäger, 1977)
- \( \lambda_i = 8.46 \times 10^{-17} \text{ a}^{-1} \) (e.g. Galliker et al., 1970)
- \( I = 7.2527 \times 10^{-3} \) (Cowan and Adler, 1976; Steiger and Jäger, 1977)
- \( \sigma = 570.8 \text{ b} \) (Wagner and Van den haute, 1992)

1.3.2 The thermal neutron fluence

As mentioned in the previous section, the current \( ^{238}\text{U} \) concentration is determined indirectly through irradiating the sample with thermal neutrons. These neutrons can be captured by the \( ^{235}\text{U} \)-nuclei and hence these nuclei will fission. However, the neutron spectrum and fluence (the amount of neutrons that cross a unit area) of a nuclear reactor do not solely consist of thermal neutrons. In fact, the neutron spectrum comprises three components (De Grave, 2003): (1) the fast or high energy neutrons; (2) the epithermal neutrons of intermediate energy and (3) the slow or thermal neutrons. The reason why thermal neutrons are used for irradiations of the samples is simply because the other two do not only cause the fission of \( ^{235}\text{U} \) but also the fission of \( ^{238}\text{U} \) and \( ^{232}\text{Th} \). Meaning that if
epithermal or fast neutrons were used for irradiation, it is impossible to tell which nuclei would have fissioned and hence the $^{235}\text{U}$-concentration (and the $^{238}\text{U}$ concentration) could not be determined. The moderator (graphite or water), is responsible for the required deceleration of fast neutrons by absorbing their energy (see also section 5.2.1.3). As a result, these fast neutrons are decelerated to an epithermal state. When a thermal equilibrium between the neutron energy and the moderator has been reached, the neutrons will be in a slow or thermal state. In practice this boils down to using a well thermalized channel (section 5.2.1.3), with a thermal/epithermal fluence ratio >50, in order to avoid interfering fission reaction caused by epithermal and fast neutrons (De Grave, 2003).

1.4. Dating procedures and techniques
This sections gives a brief overview of common procedures used to determine the $\rho_s/\rho_i$-ratio, with a special focus on the external detector (ED) method, which was applied in this study. For an extensive review the reader is referred to Wagner and Van den haute (1992). In general, an ideal fission-track dating method allows a measurement of spontaneous and induced tracks with identical properties under identical registration, etching and counting conditions (Wagner and Van den haute, 1992).

The distinction between the various techniques is based on how the induced tracks are counted. Although each technique has its own characteristics, the existing procedures can be categorized into two groups: the single-grain methods and the multi-grain or population methods. The ED-method is a the single-grain method.

1.4.1 The External Detector (or ED) method
This method (described by Price and Walker, 1963) and is currently the most widely used FT-track dating method (Tagami and O’Sullivan, 2005). Generally, it is used in combination with the $\zeta$-calibration method (see next section). The “external detector” refers to a uranium-poor muscovite or polymer sheet which is attached to the polished (and etched) sample surface and co-irradiated with this sample (figure 1.5a). Hence, induced tracks registered by the ED only originate in or just below the ED/sample-contact surface (figure 1.5b). After irradiation, only the ED is etched. Consequently, the spontaneous tracks can be counted in the polished internal mineral surface, whilst induced tracks are counted on the etched ED-surface (figure 1.5c). Note that for geological applications muscovite is typically used as an ED.

Since the spontaneous tracks are counted in an internal surface and induced tracks are counted in an external surface, $g_s$ equals 1 and $g_i$ equals 0.5. Thus the geometry ratio of this approach ($G = g_i/g_s$) equals 0.5 (Gleadow and Lovering, 1977).
The factor $Q$ (equation 1.8) typically differs from 1 due to differences between the mineral surface and the ED, e.g. difference in etching efficiency and optical counting efficiency between (in this case) apatite and muscovite (Gleadow and Lovering, 1977).

![Figure 1.5: A schematic representation of the principle of the ED method, in which spontaneous tracks are counted in the internal mineral surface and the induced tracks are counted in the ED. a) the situation before irradiation: the ED is in close contact with the polished and etched sample surface; b) the situation after co-irradiation. The ED registers tracks originated in close contact to surface; c) the situation after etching. Since only the ED is etched, spontaneous track can be counted in the internal mineral surface and induced tracks on the ED (after Tagami and O’Sullivan, 2005).](image)

1.5: The $\zeta$-calibration method: a calibration with age standards

Due to a historic controversy surrounding the values of the decay constant $\lambda_f$ and effective cross-section $\sigma$ (section 1.3), as well as the potential problems and difficulties with a precise determination of the thermal neutron fluence $\phi_{th}$ and factor $Q$, an empirical calibration method based on age standards was developed by Hurford and Green (1982, 1983). This is one of the most commonly used calibration methods to date.

1.5.1 Principles

In this approach, a user-specific empirical calibration factor or $\zeta$-factor independent of irradiation and based on age standards and U-doped glass monitors is determined.

The first step in the $\zeta$-method is a co-irradiation of age standards (in this case of the mineral apatite) with specifically designed glass monitors. Induced FT from the U-doped glass dosimeters are then counted in the attached ED. The resulting induced track density $\rho_0$ can then be directly related to the thermal neutron fluence $\phi_{th}$ by means of the proportionality factor $B$ (expressed in neutrons per track):
\[ \Phi_{th} = B \rho_d \] 1.9

This factor depends on the type of glass dosimeter, etching and observation conditions (De Grave, 2003).

In the following steps, the spontaneous and induced tracks densities of the age standard are determined by means of the ED-method.

By combining equation 1.8 and 1.9, the practical age equation for the age standard (s) now yields:

\[ t_s = \lambda_a^{-1} \ln \left[ \frac{\lambda_a}{\lambda_f} \left( \frac{\rho_s}{\rho_d} \right)_{s} QG \sigma \Phi_{th} + 1 \right] = \lambda_a^{-1} \ln \left[ \frac{\lambda_a}{\lambda_f} \left( \frac{\rho_s}{\rho_d} \right)_{s} QG \sigma B (\rho_d)_{s} + 1 \right] \] 1.10

The \( \zeta \)-factor (expressed in \( \text{a.cm}^{-2} \)) is defined as:

\[ \zeta = \frac{QG \sigma B}{\lambda_f} \] 1.11

Meaning it incorporates the factors \( \sigma, \lambda_f \) and \( \Phi_{th} \) (through the proportionality factor \( B \)). Equation 1.10 now becomes:

\[ t_s = \lambda_a^{-1} \ln \left[ \lambda_a \left( \frac{\rho_s}{\rho_d} \right)_{s} G \zeta (\rho_d)_{s} + 1 \right] \] 1.12

As the age of the standard is known and since \( \zeta \) is the only unknown parameter in this equation, \( \zeta \) can be determined as follows:

\[ \zeta = \frac{e^{\lambda_a t_s} - 1}{\lambda_a \left( \frac{\rho_s}{\rho_d} \right)_{s} G (\rho_d)_{s}} \] 1.13

The final step is the determination of the age of an unknown sample (u) which has been co-irradiated with a similar glass dosimeter by means of the following equation:

\[ t_u = \lambda_a^{-1} \ln \left[ \lambda_a \left( \frac{\rho_u}{\rho_d} \right)_{u} G \zeta (\rho_d)_{u} + 1 \right] \] 1.14

With \( (\rho_d)_u \) being the induced track density for the glass dosimeter co-irradiated with the unknown sample.
Chapter 1: The apatite fission track dating method

The ζ-factor is a personal calibration factor that can also vary with experimental set-up and conditions, mineral type and U-content in the glass dosimeter (De Grave, 2003). Hence, the calibration must be carried out for every individual researcher and for every type of mineral at certain optical settings (Wagner and Van den haute, 1992).

To enhance reproducibility, it is preferred to work with a weighted average ζ-factor. Two different factors can be defined: a *sample weighted mean zeta* (SWMZ) and an *overall weighted mean zeta* (OWMZ) (Hurford and Green, 1983). The former refers to a weighted mean for a specific age standard. The latter refers to a weighted mean of several SWMZ for different age standards, all acquired using the same glass dosimeter and under the same etching and observation conditions. This OWMZ the value is the value used for the ζ-factor in the age equation 1.14. (For the appropriate equations section 6.3.1.2)

1.5.2 Apatite age standards

Seeing as age standards are of crucial importance to the ζ-calibration method, Hurford and Green (1981) defined 6 essential requirements for a mineral set to be acknowledge as an age standard: (1) The standard should come from a formation which is geologically well documented; (2) the sample should be homogeneous in FT-age, meaning only minerals of a single age population can be present; (3) the rock unit should be readily accessible and contain reasonable amounts of the proposed standard; (4) independent ages should be unambiguously known from different methods such as relative dating by means of stratigraphy, as well as by using absolute dating methods such as the K-Ar, and Rb-Sr-systems; (5) the FT-age should represent a formation age instead of a cooling event; and (6) no correction for tracks fading should be necessary.

Wagner and Van den haute (1992) provide an overview of the various age standards used for different reference materials. For apatite, two age standards are recommended by the IUGS subcommission on Geochronology (Hurford, 1990): apatite from the *Fish canyon tuff* and the *Durango apatite*.

The Fish canyon tuff is an Oligocene tuff from Colorado, USA. $^{40}\text{Ar}/^{39}\text{Ar}$-dating of biotite by Hurford and Hammerschmidt, (1985) yielded an age of 27.9 ± 0.5 Ma.

The Durango apatite originates from a Tertiary ore body from the Carpintero volcanic group, Cerro de Mercado, Mexico. Kr-Ar dating by McDowell and Keizer (1977) yielded an age of 31.4 ± 0.5 Ma.

Other widely used standards are *Mount Dromedary* (Australia) and *Kaiserstuhl* (Germany) apatite, but the latter two were not used in this study.
Chapter 2: Geological interpretation of FT data

This chapter gives a brief overview of the principles of interpretation of FT data. For a more elaborate account, the reader is referred to Wagner and Van den haute (1992) and De grave (2003).

2.1 Thermal stability of FT

As mentioned in section 1.1, the creation of a fission tracks distorts the crystal lattice of the detector. Since a latent FT represents an energetically metastable solid state, this distortion or lattice defect will be restored with time under the influence of external factors. This process is known as fading. Fading can result in a reduction of the etchable length (shrinking) and/or a reduction of the etching rate (pitching), which in turn can reduce the areal FT density. Hence, the apparent FT age is lowered and will not represent the formation age. Although fading violates the prerequisite of a closed isotopic system, it forms the principle for the geological interpretation of FT data (Wagner and Van den haute, 1992).

Usually, the degree of fading is expressed as the reduction of track length \( l \) or track density \( p \) relative to the original length \( l_0 \) or density \( p_0 \) before fading. The \( l/l_0 \) or \( p/p_0 \)-ratio is called the retention factor \( r \). This factor varies between 1, meaning no fading has occurred and \( l=l_0 \) or \( p=p_0 \), and 0, meaning fading is complete and no tracks are retained (or \( l=0 \) and \( p=0 \)). Nowadays, the retention factor expressed in track length is preferred, seeing as it is more precise and density reduction is a secondary response compared to length reduction (De Grave, 2003).

Several geological parameters (e.g. time, temperature, hydrostatic pressure, ionizing radiation) can influence FT-fading. Of these, temperature and time appear to be the two major controlling factors (Wagner and Van den haute, 1992 and references herein) and their combined effect on FT-fading is termed annealing. Consequently, FT-ages, mean lengths and length distribution (of confined tracks) provide a unique insight into the thermal history of a sample, even allowing a t-T path reconstruction (see section 2.2.2). Especially apatite is of great importance, due to its sensitivity to low temperatures and to the extensively studied annealing characteristics of the FT in this mineral.

2.1.1 First order approximation of annealing and Arrhenius diagrams

Track fading in apatite has been extensively studied both experimentally and under geological conditions (Wagner and Van den haute, 1992). The following section will focus on the experimental research followed by the extrapolation to geological time scales.

In experiments, it is generally assumed that diffusion contributes to the reordering of the crystal lattice. Hence, it is in part responsible for the fading of the fission tracks. By considering the
annealing process solely as a diffusion process, the following equation has been derived (Märk et al., 1973)

\[
\ln t = \frac{E_a}{kT} + \ln \left[ -\ln \left( \frac{l}{l_0} \right) \right] - \ln \alpha_0
\]

With:

- \( t \) = the age of the sample (a)
- \( T \) = the absolute temperature (K)
- \( E_a \) = the activation energy for the diffusional process (J)
- \( k \) = the Boltzmann constant (= 8.616*10^-5 eV/K)
- \( l/l_0 \) = the retention factor, the ratio of the reduced track length to the original length
- \( \alpha_0 \) = a material specific resonance frequency

Note that this is a rather simple first order equation and that at present, the parameters of the empirically derived equation have no physical meaning (Wagner and Van den haute, 1992; De Grave, 2003). Based on equation 2.1, there is a linear relationship between \( \ln t \) (AFT-age) and \( 1/T \) (its thermal history). Therefore, experimental data are typically plotted in an Arrhenius diagram (figure 2.1), which is a diagram with a logarithmic time axis against an axis with a reciprocal absolute temperature (\( \ln t \) versus \( 1/T \)). The linear relationship of equation 2.1 implies that the straight lines in this diagram represent lines of equal retention factors or of an equal annealing degree. The slope of the lines is proportional to the activation energy (De Grave, 2003). Seeing as \( E_a \) and \( \alpha_0 \) are material specific, each type of mineral has its own Arrhenius diagram. From the annealing fan in figure 2.1 it is also apparent that the slope increases with increasing annealing (or with decreasing retention), indicating an increased activation energy. Usually, the variation in slope, between lines representing total retention (right-hand side of figure 2.1) and total annealing (left-hand side of figure 2.1), equals a factor two or three (Wagner and Van den haute, 1992). Mind that some authors (e.g. Green et al., 1985) dispute the idea of an annealing fan, claiming it is an artefact created by the superposition of near parallel Arrhenius diagrams.

The main advantage of an Arrhenius diagram (and equation 2.1) is that it allows the extrapolation of experimental values to geological timescales, (e.g. \( 10^6 \) a or more). As indicated in figure 2.1, it suffices to extend the straight lines to the desired time domain. For apatite, borehole data have been used to calibrate the Arrhenius curves at geological relevant time scales. Hence, this mineral can be employed to reconstruct thermal histories of rocks (Wagner and Van den haute, 1992).
Chapter 2: Geological interpretation of FT data and applicability

Figure 2.1: An example of an Arrhenius diagram, based on experimental values that were extrapolated to geological time scales. The retention actor \( r \) is a measure for the degree of annealing, with 100% meaning total retention. The lines of equal retention form an annealing fan. The left-hand side represents total annealing, while the right-hand side represents total retention. The total annealing is related to the exposure to high temperatures (Wagner and Reimer, 1972).

2.1.2 A more complex approach by incorporating the effects of chemical composition and crystallographic orientation

The above mentioned first-order approach is an oversimplification, since the chemical composition as well as the crystallographic orientation (more specifically the orientation of tracks versus the crystallographic c-axis) can both have an influence on track fading in apatite (Carlson et al., 1999; Donelick et al., 1999). (The latter is also referred to as the anisotropy of annealing). Therefore, several other models have been developed, such as the Laslett et al. (1987) and Ketcham et al. (1999) and the Ketcham et al. (2007) models. In this thesis, the Ketcham et al. (2007) annealing model was used to reconstruct the time-temperature (or t-T-paths) (section 2.1.2.3). This is a purely empirical model based on isothermal annealing data in apatite. For a review of the other models, the reader is referred to Ketcham (2005).

2.1.2.1 The effects of the chemical composition of apatite

The general crystal-chemical formula of apatite is \( \text{Ca}_5(\text{PO}_4)_3(\text{OH}, \text{F}, \text{Cl}) \), with hydroxy apatite (\( \text{Ca}_5(\text{PO}_4)_3(\text{OH}) \)), fluoapatite (\( \text{Ca}_5(\text{PO}_4)_3\text{F} \)) and chlorine apatite (\( \text{Ca}_5(\text{PO}_4)_3\text{Cl} \)) as endmembers. Of which the fluorine endmember is the most common (rock-forming) variant and OH-apatite endmember is very rare. The rate of thermal annealing of AFT varies with the varying composition (Green et al., 1985; Carlson et al., 1999). The chlorine-endmember is more resistant to annealing than the (more common) fluorine-endmember (e.g. Green et al., 1985). Therefore, these authors suggested to use the Cl/F or Cl/(F+Cl)-ratio as a measure for the effect of the chemical composition of apatite on annealing. The Ketcham et al. (1999) and (2007) multi-kinetic models allow the incorporation of the Cl-content as a measurable kinetic parameters (see section 2.1.2.3) for this purpose.
These models also allow the use of the so-called $D_{\text{par}}$ kinetic parameter as a proxy for the effects of the chemical composition of apatite. $D_{\text{par}}$ is the etch pit diameter of AFT etch figures parallel to the crystallographic c-axis (or the maximum diameter). It is an indirect measure of the composition as it depends on the etching condition and therefore also on the chemical composition of apatite. More specifically, a high $D_{\text{par}}$ generally implies a high Cl content under a specific etching condition (Barbarand et al., 2003; De Grave, 2003 and references herein).

2.1.2.2 Crystallographic orientation and the c-axis projection model

Only taking quantities depending on the chemical composition into account, such as the Cl-content, does not fully explain the observed variability in annealing properties (Carlson et al., 1999). The annealing process is also anisotropic. According to Donelick (1991) the AFT-length appears to decrease with increasing angle of the AFT with the crystallographic c-axis. In other words, AFT parallel to the c-axis will be longest, whilst AFT perpendicular to the c-axis will be the shortest. Moreover, this anisotropy appears to increase with an increasing degree of annealing (Donelick et al., 1999). Donelick (1991) defined an empirical model in order to quantify this anisotropy: the so-called elliptical model (see figure 2.2). This model considers an ellipse with as the largest semi-axis the mean length of AFT parallel to the c-axis and with the smallest semi-axis as the mean length of AFT perpendicular to the c-axis. The etchable length of an AFT at any angle to the crystallographic c-axis is given by the corresponding axis of this ellipse. Unfortunately, this model fails for mean track lengths of about 11 µm or smaller, as starting from 11 µm AFT at high angles will experience an accelerated reduction (Donelick et al., 1999). Therefore, Donelick et al. (1999) defined the empirical c-axis projection model, which is an extension of the previously mentioned elliptical model that is applicable to all degrees of annealing, even when accelerated track reduction has occurred. This model has been incorporated by the Ketcham et al. (1999) and Ketcham et al. (2007) models. So besides the chemical composition, these models also account for the effects of crystallographic orientation on annealing.

2.1.2.3 The Ketcham et al. (2007) model.

Several thermal annealing models for apatite have been developed. The Ketcham et al. (1999) and Ketcham et al. (2007) multi-kinetic annealing models are the two most widely used. (Multi-kinetic refers to a variation in annealing behaviour between apatites.) For this thesis, the Ketcham et al. (2007) model was chosen.
The Ketcham et al. (2007) model is an empirical model based on the Ketcham et al. (1999) annealing equations. The authors constructed this so-called super model (sic) in order to predict the measurement from two major AFT data sets: that of Barbarand et al. (2003) and that of Carlson et al. (1999). In other words, it allows an estimation of a mean track length and standard deviation that is no longer specific to a single dataset and hence to specific apatite etching conditions, as was the case with the Ketcham et al. (1999). The model uses the c-axis projection technique to normalize track length if the user so wishes. Hence it takes the anisotropy of annealing into account. Note that this was not necessary for this thesis, as the applied etching procedure is associated with a near-isotropy. So in this case, correcting for an anisotropic annealing would induced errors (Glorie and De Grave pers. comm.).

Similar to Ketcham et al. (1999), the effects of compositional variations are considered by means of Equation 2.2:

\[ r_{tr} = \left( \frac{r_{mr} - r_{min}}{1 - r_{mr0}} \right) ^ \kappa \]  

2.2

With

\[ r_{trmr} = \text{the reduced length of apatites that are less resistant (lr) and more resistant (mr) to annealing} \]

\[ r_{mr0}, \kappa = \text{empirical parameters.} \]
By relating the empirical parameters in equation 2.2 to measurable kinetic parameters, the Ketcham et al. (2007) model can estimate the annealing kinetics. These kinetic parameters include composition (e.g. Cl-content), etch-figure dimension ($D_{par}$) and unit-cell parameters (Ketcham et al., 2007). The precise relationship between the empirical and measurable parameters forms another major difference between the Ketcham et al. (1999) and (2007) models. For more information, the reader is referred to the publications of Ketcham et al. (1999; 2007).

In this thesis, the Ketcham et al. (2007) annealing model was used in order to reconstruct time-temperature or t-T paths, which are a valuable aid for interpreting AFT-ages and lengths (see next section). The modelling was performed by means of the HeFTy-programme.

2.2 Geological interpretation of FT ages

Due to the sensitivity of AFT-geochronometer to low temperatures, AFT-ages rarely approximate the formation age (see below). Yet, they often hold important geological information. In order to interpret (A)FT-age with respect to geological events, two concepts have to be defined: the closing temperature and the Partial Annealing Zone concept.

2.2.1 The closure temperature and Partial Annealing Zone concepts

2.2.1.1 The closure temperature concept

For this section, it is important to bear in mind that the fading of apatite fission tracks is a gradual process. In other words, values such as the closure temperature $T_c$ (generally defined as the temperature at the time given by the apparent age) are not fixed, but can vary in function of the cooling rate (De Grave, 2003 and references herein).

Similar to most geochronological systems three temperature thresholds can be defined for the (A)FT-method (figure 2.3) (De Grave, 2003): (1) $T_A$ or the minimal temperature of total annealing. This temperature represents a threshold above which no tracks are retained; (2) $T_R$ or the temperature of total retention. In principle, this should represent the temperature threshold below which no annealing occurs. However, fission tracks can still be annealed at ambient temperatures. Consequently, total retention can never be achieved in geological conditions. Therefore in terms of FT-dating, $T_R$ is considered as the temperature above which annealing rates increase significantly (Wagner and Van den haute, 1992) (figure 2.3), c) the closing temperature $T_c$, which in terms of FT-dating has been defined as the temperature at which 50 % of the tracks are retained or annealed (Wagner and Reimer, 1972). $T_c$ is located within the transition zone between $T_A$ and $T_R$ (figure 2.3).
As previously mentioned, $T_C$ but also $T_A$ and $T_R$ can depend on the cooling rate. They all increase with an increasing cooling rate (De Grave, 2003). Moreover, $T_A$, $T_C$ and $T_R$ values can also vary in function of the chemical composition of apatite.

For a normal constant cooling rate of ca. 10 °C/Ma, Wagner and Van de haute (1992) calculated a $T_C$ of 100 ± 20°C, a $T_A$ of around 120-125 °C and a $T_R$ of around 60 °C for fluoapatite (which is the most common type of apatite).

Mind that apatites can have complex geological histories e.g. several heating and cooling phases, meaning that they have not always known a single cooling rate (De Grave, 2003).

![Figure 2.3: Schematic and idealized representation of the PAZ-concept. Three temperature zones are defined based on the stability of the FT expressed as the retention factor $r (l/l_0$ or $\rho/\rho_0)$. The TSZ or total stability zone is characterised by a complete retention or $r = 1$. The PAZ or partial annealing zone is characterised by an intermediate behaviour varying from total retention at the top to total annealing at the bottom. The TAZ or that annealing zone is characterised by a complete annealing of the FT or $r = 0$. Based on Wagner and Van den haute (1992).](image)

### 2.2.1.2 The Apatite Partial Annealing Zone concept or APAZ

The Earth’s crust can geothermally be divided into three depth zones, based on the three aforementioned temperatures (figure 2.3) (Wagner, 1981; Wagner and Van den haute, 1992; De grave, 2003). With decreasing temperature these zones are: (1) a bottom zone of instability known as the Total Annealing Zone (TAZ), of which the top is defined by $T_A$. As temperature increases from top to bottom (or with increasing depth), $T_A$ forms the low-temperature boundary; (2) an intermediate zone where tracks are partially stable. This zone is known as the Apatite Partial Annealing Zone.
Chapter 2: Geological interpretation of FT data and applicability

(APAZ). It is bounded at the bottom by \( T_A \) and at the top by \( T_R \); (3) a top zone of total-track stability known as Total Retention Zone (TRZ) or Total Stability Zone (TSZ). In depth, this zone is bounded at the bottom by \( T_R \).

As previously mentioned, \( T_C (100 \pm 20 \, ^\circ C) \) is located centrally in the APAZ (Wagner and Van den haute, 1992). Under normal cooling rates, the APAZ is bounded by the \( \sim 60 \, ^\circ C \) and \( \sim 120 \, ^\circ C \). As indicated in figure 2.4, this translates to a depth interval for the APAZ between ca. 2 km and ca. 4 km (assuming a “normal geothermal gradient” of 25-30 \(^\circ C/\text{km}\)) (De Grave, 2003).

The annealing behaviour of the tracks differs from zone to zone (Wagner and Van den haute, 1992): (1) in the TAZ, AFT do not accumulate as they are instantaneous annealed after their formation due to the high temperatures. Hence, both the retention factor \( r (l/l_0 \text{ or } \rho/\rho_0) \) and the AFT-age will equal zero; (2) in the APAZ, the temperature decreases from bottom to top. Consequently, the track stability, track accumulation and track length increase from bottom to top. More specifically, the retention factor varies from 0 at the bottom (total annealing) to 1 at the top (total retention) (figure 2.3). As a result, AFT accumulation can take place, but the tracks are subjected to shortening. Due to the accumulation, the AFT-geochronometer can start registering time in this zone; (3) in the TRZ or TSZ the temperatures are low enough for the tracks to remain stable. So basically, all the tracks that survived the APAZ and TAZ will be retained and no further shortening will occur.

2.2.1.3 \( t-T \) paths and the associated type of AFT-age

Based on the previous section, it can be concluded that the thermal (or \( t-T \)) history of a sample determines the track density and hence the AFT-age. Several possible cooling (or \( t-T \)) paths are depicted in figure 2.4 (Wagner, 1981; Wagner and Van den haute, 1992; De Grave, 2003).

Path 1 in figure 2.4 represents a fast steady-cooling throughout the entire APAZ with no further thermal disturbance, which is the case for e.g. undisturbed volcanic rocks. In this scenario, track accumulation quickly follows mineral formation and continues at a constant rate. Throughout this cooling event, track fading is negligible. Hence, the AFT will not have been significantly shortened and the AFT-age will typically approximate either the formation age \( t_f \) or the “fast cooling event age” (meaning age of the cooling event) \( t < t_f \).

Path 2 in figure 2.4 represents a slow steady-cooling through the APAZ, typical for uplifted basement rocks, e.g. plutonic rocks. In comparison to path 1, the samples did spend a significant amount of time in the APAZ. Hence, tracks will only have become stable and started to accumulate (long) after mineral formation \( t_f \). And the tracks will clearly have been shortened. Consequently, the AFT-age represents a cooling age \( t_c \): moment in time when the \( T_C \) isotherm was crossed and tracks were effectively retained.
Path 3 in figure 2.4 represents a reheating that was insufficient to completely anneal all pre-existing AFT and hence did not entirely reset the AFT-geochronometer. Due to the thermal event, tracks that had accumulated before this event were partially annealed and shortened during this event. On the other hand, the accumulation of tracks formed after this thermal event is comparable to that along path 1 or 2 (depending on cooling rates). As a result, the age $t_n$ will plot somewhere between the time of formation and the moment in time when the TSZ was reached. This is called a mixed age and is geologically meaningless, as it neither approximates the time of formation or the time of the thermal event. However, a mixed age might help to determine the intensity of the thermal event. Several geological processes can create such a path, amongst which sedimentary burial, tectonic subsidence and hydrothermal activity.

Path 4 in figure 2.4 also represents a thermal event. However, contrary to the third path, this reheating event was intense enough to anneal all pre-existing AFT and hence to completely reset the AFT-geochronometer. If the cooling following the thermal event was fast, the scenario would be
similar to that of path 1 and the AFT-age would yield the age of the thermal event $t_\alpha$. This age referred to is an overprint age. If the post-event cooling was slow, the post-event situation would be comparable to path 2 and the AFT-age would represent a cooling age.

2.2.2 AFT-length as thermochronological tool: the reconstruction of t-T-paths

The distribution of AFT-lengths is a powerful tool for reconstructing thermal histories, seeing as every t-T-path results in its own specific length distribution (Wagner and Van den haute, 1992). Hence, these length distributions can help interpret the AFT-age (Gleadow et al., 1986). Moreover, not only the distribution is of importance, but the mean track length ($l_m$) and standard deviation ($\sigma$) also hold information on the thermal history. After all, progressive annealing (e.g. through the APAZ) results in a reduction of the track lengths (see previous section) and an increase in the standard deviation due to a broadening of the length distribution (Gleadow et al., 1986; Wagner and Van den haute, 1992; De Grave, 2003).

![Figure 2.5](image)

**Figure 2.5:** An overview of the five different end-member length distributions of confined tracks (in apatite): a) freshly induced track distribution; b) undisturbed volcanic type-distribution; c) the undisturbed basement type-distribution: d) mixed distribution and e) bimodal distribution (De Grave, 2003).

In general, five end-member types of length distributions (of confined tracks) can be distinguished (figure 2.5) (Gleadow et al., 1986; De Grave, 2003): (1) freshly induced (figure 2.5a) fission tracks yield a narrow, symmetrical distribution. This distribution is typically characterised by a (large) $l_m$ of 16.3 $\mu$m (see section 1.1) and a $\sigma$ of 0.9 $\mu$m; (2) samples which were rapidly cooled after formation and that were not reheated (like volcanic samples, see section 2.2.1.3) exhibit a so-called undisturbed volcanic distribution (figure 2.5b; 2.6b). Similar to scenario (1), this is a narrow, symmetrical distribution. The main difference with freshly induced tracks is that however brief the passage through the APAZ might have been, the tracks will still have been slightly annealed and hence will have been shortened. The resulting length distribution is centred around a $l_m$ of 14.0 and 15.7 $\mu$m with a $\sigma$ of 0.8 to 1.3 $\mu$m; (3) a slow cooling of apatite through the APAZ, without being reheated results in the so-called undisturbed basement distribution (figure 2.5c; 2.6c). Seeing as a slow cooling
implies a progressive annealing, the AFT will typically be shorter than in the two previous cases ($l_m$ between 12.5 and 13.5 µm) and the resulting length distribution will be somewhat broader with a σ between 1.3 and 1.7 µm. As can be seen in figure 2.5c the distribution is somewhat negatively skewed, meaning it has a tail of small tracks. This skewness is attributed the older tracks. This distribution is typical for uplifted basement rocks such as plutonic rocks; (4) when apatite has been reheated, AFT-lengths strongly vary and the resulting distribution is called a mixed distribution (figure 2.5d). Typically these are very broad distributions, due to the shortening of the tracks pre-dating the thermal event(s). In these distributions, the younger unaffected tracks are the longer ones. Mixed distributions usually have a $l_m < 11.5$ µm and a σ of around 2 µm; (5) a bimodal distribution (figure 2.5e; 2.6e) is a mixed distribution and is characterised by two distinct modes in the length distribution and by a $l_m < 13$ µm and a σ > 2 µm. The two (well preserved) length distributions can be seen as the result of a two-stage thermal evolution, the smaller tracks represent partially annealed tracks that predate the thermal event, the larger ones are the unannealed tracks that post-date the thermal event. Note that an intense thermal event, strong enough to completely anneal every pre-existing FT, will probably yield a narrow, symmetrical distribution.

![Figure 2.6: This figure illustrates how AFT-length distributions, mean track length and the standard deviation can help interpret FT ages. For more information: see text.](image)

### 2.2.3 Geological interpretation of AFT-data

As illustrated in the previous sections, AFT-ages can be interpreted in various ways. In order to determine whether the AFT-age represents a cooling age, additional information is needed (Wagner and Van den haute, 1992). A first additional information source has already been outlined: the track
length distribution. Independent geological information on the thermal history of the region is another important information source. Usually other radiometric ages (of co-existing minerals) with different closing temperatures are used. The lower the closing temperature of the system, the younger the radiometric age is expected to be.

Once a cooling age has been determined, the mean cooling rate of the sample can be calculated (De Grave, 2003):

\[ u_c = (T_c - T_0) t_{AFT}^{-1} \]  (2.3)

With:
- \( u_c \) = the mean cooling rate in the temperature interval \([T_{C_{AFT}}, T_0]\)
- \( T_c \) = the effective closure temperature for the AFT-system
- \( T_0 \) = the ambient surface temperature of the sample
- \( t_{AFT} \) = the AFT-cooling age

AFT-ages of basement rocks are often interpreted in terms of uplift and denudation (De Grave, 2003). Therefore it is important to unambiguously define these terms. Uplift refers to a displacement of rocks or a surface in opposite sense of the gravity vector in a fixed reference framework (often the mean sea level is used). Both tectonic forces and isostasy can result in uplift. Denudation is the displacement of rocks with respect to the surface. Two types of denudation are recognised: erosional and tectonic. The major difference between uplift and denudation is indicated in figure 2.7: uplift alone does not disturb the thermal structure of the uplifted rock column, whilst denudation does (figure 2.7). Therefore, in the absence of conclusive geological indications for uplift, it is preferred to interpret the AFT-ages in terms of cooling and denudation (De Grave, 2003 and references herein).

2.2.3.1 Cooling through denudation
Tectonic uplift can often induce denudation. When there is no considerable time-lag between the two, denudation can cause a continuous cooling of the rock column on its way to the surface. In this case the mean uplift rate \( (u_x) \) can be calculated from the cooling rate \( (u_c) \), provided the geothermal gradient stays constant \((dT/dx)\):

\[ u_x = u_c (dT/dx)^{-1} \]  (2.4)

The AFT-age has become an uplift age, representing the moment in time when the \( T_c \)-isotherm was crossed (Wagner and Van den haute, 1992; De Grave 2003). It is also possible that the time lag
between tectonic uplift, uplift induced denudation and the thermal relaxation of these isotherms is no longer negligible. This is the scenario depicted in figure 2.7. As a result, the ages can only be interpreted in terms of denudation (De grave, 2003).

Denudation does not have to be triggered by a tectonic uplift, other possible triggers are rifting processes, which result in a base-level change and denudation, climatically controlled base-level drops, etc. (De Grave, 2003).

![Figure 2.7](image)

**Figure 2.7**: Cartoon representing the causal relationship between uplift, denudation and thermal relaxation relative to the sea-level. The 100 °C-isotherm is the Tc-isotherm. At time t₁ uplift occurred, without disturbing the thermal structure of the uplifted block. Denudation does not take place until time t₂. Time t₃ signifies the end of the denudation process, which lowered the surface to the normal base-level. Due to the associated compression of the isotherm, there will be an increased heat transport to the surface in order to restore the original thermal structure at time t₄. In other words, the geothermal gradient will have increased. From Wagner and Van den haute, (1992).

### 2.2.3.2 Tectonic versus thermal model

In the discussion up till now, the geothermal gradient was considered to be constant. However, this is not always the case, for example when a heat source is involved (figure 2.8). Hence, a horizontal age difference such as the one in figure 2.8 can be interpreted in two ways: (1) a tectonic interpretation, with a constant geothermal gradient. The difference in AFT-age then might indicate a differential uplift/denudation: the lower the AFT-age, the faster the uplift/denudation; (2) a thermal interpretation, with a constant uplift/denudation rate. The difference in FT-age now indicates a variability in the geothermal gradient. The younger FT-ages, the higher the geothermal gradient (Wagner and Van den haute, 1992).
2.2.3.3 The influence of topography

In our overview, the temperature has been considered to increase linearly with depth. Implying a thermal structure of the Earth’s crust as represented in figure 2.7 (t₁) and 2.9 (y₀), with steady-state isotherms parallel to the surface. Even though this might hold in areas with a limited surface topography during periods of tectonic quiescence and erosional stability, this will no longer hold in regions with active erosion and denudation (figure 2.9) (De Grave, 2003). In other words, both the wavelength of the topography and the rates of erosion and denudation can perturb the structure of the steady-state isotherms. As indicated on figure 2.9, the isotherms will be stretched out under hilltops and compressed underneath valleys. According to Braun (2002) this perturbation decreases exponentially with depth, meaning that the influence is mainly restricted to the low-temperature isotherms (figure 2.9).

Given the dependence of the interpretation of AFT-ages on temperature, it is not surprising that this perturbation can affect the AFT-geochronometer (which is sensitive to low temperatures). According to Stüwe et al. (1994), the effect on the AFT T_C-isotherm (the 100°C-isotherm) becomes noteworthy for denudation rates larger than 500-1000 m/Ma and for a topography with amplitudes greater than 3 km and wavelength greater than 20 km.
Figure 2.9: The influence of topography and active erosion $U$ on the steady-state isotherms in the Earth’s surface. $T_s$ represents the surface temperature. $Y_0$ represents the initial elevation before the active erosion in which the steady state isotherms are parallel to the surface. $Y_1$ and $y_2$ illustrate the effects of active erosion: the low-temperature ($T_L$) isotherms are perturbed whilst the high temperature ($T_H$) isotherms remain unaffected. Under hilltop, the $T_L$ isotherms are stretched out ($y_1$), whilst they are compressed underneath valleys ($y_2$). From Stüwe et al. (1994).
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A first set of samples was collected on the eastern passive Indian margin. All sample sites are located in the state of Tamil Nadu, situated in the most southeastern part of Peninsular India, see figure 3.1a. As indicated in figure 3.1b, the (four) samples analysed in this work were located along an E-W profile on the Tamil Nadu margin. This profile was chosen as a reference for the Thailand samples (see next chapter).

This chapter will focus on the aspects of the geological history of India relevant for these samples.

3.1 Introduction

The main part of the Indian basement is constructed of Archean and Proterozoic crustal blocks and cratonic nuclei. This is also the case for southern India (and Tamil Nadu in particular). This Archean and Proterozoic basement is locally covered by Cenozoic sediments of various nature (figure 3.1b).

![Figure 3.1: a) Main setting plate motions (black arrow) of India. The red box indicates the study areas. Also the state of Tamil Nadu in southern Peninsular India is indicated (after from Simons et al.,2007, Metcalfe, 2011); b) a part of the geological map of northern Tamil Nadu with the location of our sample profile. (Source: the Geological Survey of India; see also appendix B).](image-url)
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Figure 3.2: Overview of the tectonic history of India, after Hodges et al. (2001); Santosh et al. (2009); and Chatterjee et al. (2013).
The tectonic history of (Southern Peninsular India) is summarised in figure 3.2. The pre-Mesozoic tectonic history of southern Peninsular India is dominated by an orogenic phase of contrasting nature: a progressive sequence from a Neoproterozoic Pacific-type subduction-accretion orogeny, to a Himalayan-type Cambrian orogeny (Santosh et al., 2009; Saitoh et al., 2011). This orogenic sequence resulted in not only the closure of an intervening ocean and the development of a Himalaya-scale orogenic belt, but also in the amalgamation of the different crustal blocks that build up Southern Peninsular India (Santosh et al., 2009; Saitoh et al., 2011). Moreover, the deformation of Southern Peninsular India, and more specifically that of the Southern Granulite Terrane (see below) can be considered in the larger geological framework of the suturing of Gondwana, as India did not coalesce with other Gondwana continents until the Latest Neoproterozoic to Cambrian (Collins et al., 2007; 2013 and references herein).

The Mesozoic history of India was dominated by its break-up from Gondwana in the Jurassic, followed by its northward drift during the Cretaceous, which resulted in the opening of the Indian Ocean. The Cenozoic history of India was subsequently dominated by the gradual closure of Tethyan Basins between the Indian and (Eur)Asian plates and the associated India-(Eur)Asia collision, which invoked the formation of the Himalaya orogeny and associated far-field effects (such as strike-slip faulting and shearing in Southeast Asia including Thailand) (Chatterjee et al., 2013).

3.2 The crustal blocks in Peninsular India
As indicated on figure 3.3, southern Peninsular India is constructed of a mosaic of crustal blocks, dissected by Late Neoproterozoic-Cambrian shear (or suture) zones (Mukhopadhyay et al., 2011). Generally, a distinction between a northern and southern metamorphic terrane is made. Both terranes have their own specific protolith origins and thermotectonic histories (Collins et al., 2013). The northern terrane consist of a low-grade metamorphic facies, dominated by the Archean Dharwar Craton. The southern terrane (which is the focus of this thesis) is a high-grade terrane which is made up of several Proterozoic (to Cambrian) granulite blocks and dissected by shear zones including the Palghat Cauvery Shear Zone (PCSZ) (figure 3.3) (Santosh et al., 2003; Collins et al., 2013). This shear zone isolates most of Tamil Nadu from the rest of India. It also forms the boundary between the Dharwar Craton and the granulite blocks (Saitoh et al., 2011). The granulite block north of the PCSZ and west of the Moyer-Bhavani shear zone (MBSZ) is called the Salem Block (also referred to as the Northern Granulite Terrane, not to be confused with the previously mentioned terranes). The block to the east of the MBSZ is known as the Madras Granulite Block. The block situated to the west to the PCSZ is known as the Nilgiri Granulite Block (the latter two are not always considered as separate blocks, e.g. Santosh et al., 2009). The Madurai, Trivandrum and Nagercoil Blocks are all located to the
south of the PCSZ. These three blocks are also called the Southern Granulite Terrane or SGT (Collins et al., 2010; Santosh et al., 2003). Our study area is located in the Madras Block (figure 3.3). In the following section a brief overview of the most important blocks will be given, based on the reviews presented by Santosh et al. (2009) and Collins et al. (2013). For more elaborate information, the reader is referred to their works.

![Geological map of southern peninsular India](image)

**Figure 3.3:** Geological map of southern peninsular India, depicting the different blocks and suture zones (from Kooijman et al., 2011). The red box indicates the study area, which is located in the Madras block close to the MBSZ or Moyer-Bhavani shear zone. The other abbreviations are: PCSZ = Palghat-Cauvery Shear Zone; ACSZ = Achankovil Shear Zone and KKPTSZ = Karur-Kambam-Pianavu-Trichur Shear Zone.

### 3.2.1 The Salem Block

The Salem block is considered to be composed of Archean and Neoproterozoic rocks (Santosh et al., 2009). The gradual boundary between this block and the Dharwar Craton, which shows no clear lithological or tectonic contacts, indicates that the granulites of the Salem Block are the metamorphosed counterparts of the Dharwar Craton (Collins et al., 2013 and references herein). Based on geochronological data, protoliths of the Salem Block appear to be Meso- to Neoarchean in age and have experienced two major phases of metamorphism. A first metamorphic phase occurred during the Latest Archaean to Earliest Proterozoic, with pressures and temperatures up to the granulite facies. The second phase was a high-grade metamorphic phase of Latest Neoproterozoic to Cambrian age. The P-T-evolution during this phase is currently still poorly understood (Santosh et al., 2009; Collins et al., 2013 and references herein). The tectonic setting of the Salem Block is interpreted in terms of a suprasubduction zone. In similar settings, a mantle wedge is typically
created above the subducting plate. Recent studies on the Salem block suggest that this block could be fragment of such mantle wedge (Santosh et al., 2009 and references herein).

### 3.2.2 The Madurai Block

The Madurai Block is the largest crustal block in southern India (Santosh et al., 2009). The northwest part mainly consists of Neoarchean (2.55 and 2.50 Ga) charnockite (or orthopyroxene-bearing granites) massifs intercalated with gneisses. The bulk of the massive, however, is dominated by metasediments. The lithological boundary between the two more or less follows the Karur-Kamban-Painavu-Trichur shear zone or KKPTSZ (figure 3.3) (Santosh et al., 2009; Collins et al., 2013 and references herein). The metasedimentary belts were interpreted as being developed in an accretionary setting. Subsequently, they were subjected to (and hence overprinted by) high grade metamorphism during the aforementioned ancient orogenic sequence. The geochemical signature of these charnockites also indicate a component of arc magmatism, implying that the Madurai Block might have originated from a long-lived Neoproterozoic arc. Hence, the present day exposures are interpreted as the eroded roots of this arc (Santosh et al., 2009, Plavsa et al., 2012 and references herein).

Next to these Neoarchean granites, a suite of younger (Cambrian-Ordovician) granites have also been described in the Madurai Block. Based on their alkaline geochemical nature, they are related to post-orogenic magmatism in an extensional setting where the magmas predominantly followed pre-existing conduits along this orogenic fabric (Santosh et al., 2009 and references herein).

The Madurai Block is characterised by the combined occurrence of the aforementioned arc magmatic rocks with HP-HT metamorphosed rocks (Santosh et al., 2009). This juxtaposition differs from the situation in most present day subduction zones, where the two are separated in space. Generally, the HP-HT metamorphosed rocks occur were the two plates collide. The magmatic arc on the other hand, is located on the overriding plate. Although the mineralogy of the HP mafic rocks indicates a supra-subduction zone affinity, deep seated subduction and subsequent extrusion alone cannot explain the juxtaposition of these rock units (Santosh et al., 2009). Therefore, the authors invoke tectonic erosion: due to this kind of erosion, the active margin would be obliterated into the deep mantle. As a result, the younger accretionary complex would develop downwards only to be periodically exhumed.
3.3 The final amalgamation of Gondwana

3.3.1 The role of southern India in the amalgamation of Gondwana

According to Proterozoic palaeogeographic reconstructions (e.g. Collins and Pisarevsky, 2005), India did not coalesce with any of the Gondwana blocks the Latest Neoproterozoic to Cambrian (Collins and Pisarevsky, 2005; Collins et al., 2013). During Neoproterozoic times, the Mozambique Ocean separated Neoproterozoic India (India north of the PCSZ) from other continental blocks (more specifically the Congo/Tanzania/Bangweulu Block, figure 3.4a,b). This ocean is assumed to have been over 3000 km wide (Santosh et al., 2009 and references herein).

Moreover, these reconstructions show that southern India and the bordering regions of Madagascar, Sri Lanka and Antarctica are situated along a number of orogenic belts (figure 3.4b,c,d). These belts are assumed to have formed during the amalgamation of Gondwana due to the collision of India, Australia, Azania (see below), Kalahari and Antarctica (Collins et al., 2007). As represented in figure 3.4, (southern) India was situated near the convergence zone between the East and West Gondwana, therefore (southern) India is said to have formed the “heart” of Gondwana (Santosh et al., 2009; Collins et al., 2013). Within Gondwana, a Himalaya-scale orogen, the East African Orogen, can be traced starting from southern India, through Madagascar, East Africa all the way north to Arabia (Santosh et al., 2009; Collins et al., 2013). This orogen is believed to have formed in two main phases: a Neoproterozoic East African Orogeny s.s. (between approximately 650-630 Ma) and a Neoproterozoic to Cambrian Malagasy Orogeny (between 550-510 Ma) (Santosh et al., 2009; Collins et al., 2013 and references herein). During the East African orogeny, Azania (which consists of Archaean and Palaeoproterozoic crust of Madagascar, Somalia, Ethiopia and Arabia) collided with the Congo/Tanzania/Bangweulu Block (Collins and Pisarevsky, 2005). The Malagasy Orogeny on the other hand, is said to be the result of the closure of the Mozambique Ocean (Collins and Pisarevsky, 2005; Santosh et al., 2009). Hence, it resulted in the amalgamation between India and the already unified Congo/Tanzania/Bangweulu–Azania Block (Collins and Pisarevsky, 2005; Santosh et al., 2009; Collins et al., 2013). Collins and Pisarevsky (2005) also mentioned a third orogenic phase, coeval with the Malagasy Orogeny: the Kuunga Orogeny. According to these authors, the Kuunga Orogeny led to the amalgamation of India with the Australia/Mawson Block (figure 3.4d). At the end of this time period, Gondwana had completely amalgamated.
Figure 3.4: Palaeogeographic reconstruction of Gondwana: a) at around 750 Ma (before the East African Orogeny); b) at around 630 Ma (at the final phase of the East African Orogeny); c) at around 570 Ma (after the East African Orogeny and before the Malagasy Orogeny); d) at around 530 Ma (during the Malagasy and Kuunga Orogeny) (from Collins and Pisarevsky, 2005). Deep yellow indicates continental plates, pale yellow indicates the hypothesised extent of continental crust. The following abbreviations were used: Amazon=Amazonia; Aus-Maw = Australia/Mawson Block; Az = Azania; Congo = Congo/Tanzania/Bangweulu Block; Kal = Kalahari Block; Laur=Laurentia; RP = Rio de la Plata; SF = São Francisco; WA = West Africa, Adola, Adamastor, Brasiliano, Mozambique; Pacific = oceanic basins.
3.3.2 Terrane Assembly in Southern Peninsular India

The mosaic of crustal blocks in the SGT in southern India is considered to have been assembled during the Latest Neoproterozoic to Cambrian within the context Gondwana supercontinent along the PCSZ (Santosh et al., 2003; Saitoh et al., 2011; Collins et al., 2013). This amalgamation is a direct result of the Malagasy Orogeny. Consequently, the PCSZ is interpreted as the trace of the Cambrian suture, which represents the closure of the Mozambique Ocean (Santosh et al., 2003; 2009; Saitoh et al., 2011).

To the best of our knowledge, the most recent model for the amalgamation of the southern Indian blocks, and the related closure of the Mozambique Ocean, is the one proposed by Santosh et al. (2009). This model is summarised in figure 3.5 and comprises an entire orogenic cycle from ocean rifting to closure.

Figure 3.5a represents the first stage of their model: oceanic rifting due to mantle plume activity. This rifting phase signified the beginning of the Mozambique Ocean and created passive continental margins. Although the exact timing of this continental rifting phase is still uncertain, it most likely coincided with the rifting of the Rodinia supercontinent some 1000 to 750 Ma ago (Santosh et al., 2009).

The actual amalgamation of southern India did not start until the Neoproterozoic (around 540 Ma). It commenced as a Pacific-type subduction-accretion orogeny (Figure 3.5b), which coincided with the collision between the Madurai Block (Azania) and the Salem Block (Neoproterozoic India) (Collins et al., 2013). The Pacific-type orogeny had actually anteriorly been initiated as an Andean-type by the subduction of the oceanic lithosphere carrying the Mozambique Ocean underneath the continental margin of the Dharwar Craton. This southwards subduction beneath the Dharwar Craton was responsible for the eventual closure of the intervening Mozambique Ocean (Santosh et al., 2009). This closure was accompanied by the collision of the Archean Dharwar Craton with the already amalgamated Madurai and Salem Blocks. This collision in turn invoked the development of a Himalayan-type (Cambrian) Orogenic belt (figure 3.5c). During this Himalaya-stage, post-orogenic extension and emplacement of late stage intrusives transpired in the different blocks (Santosh et al., 2009).

Note that a similar model has been suggested for the eastern margin of the Dharwar Craton, whereby ocean closure commenced in the Mesoproterozoic and a final collision occurred in the Neoproterozoic (Saitoh et al., 2011 and references herein).
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Figure 3.5: Cross-sections illustrating the a) the formation of the Mozambique Ocean around 1 Ga to 750 Ma; b) the Neoproterozoic Pacific stage and c) the Neoproterozoic to Cambrian Himalaya stage of the orogeny that would lead to the closure of the Mozambique Ocean, the amalgamation of the different crustal blocks of India as well as the final amalgamation of Gondwana. The following abbreviations were used: MB = Madurai Block and MOR = Mid Oceanic Ridge. (from Santosh et al. 2009).
3.4 The break-up of the Indian plate from Gondwana and its northward drift

3.4.1 Introduction

The break-up of eastern and western Gondwana initiated around 167 Ma (Middle to Late Jurassic) and was completed around 65 Ma (Cretaceous-Cenozoic boundary) with the separation of India from the Seychelles. This implies that Africa, Antarctica, Australia and India had separated and the Indian Ocean had opened (Chatterjee et al., 2013).

Generally, a distinction between two types of rifting, leading to continental break-up, is made: (1) active rifting which results from mantle plume activity and (2) passive rifting which is associated with far-field stresses. Based on the relative timing of rifting and rift-related volcanism, the two processes can be distinguished. Volcanism predates rifting in the active rifting scenario, but post-dates it in the passive rifting scenario (Chatterjee et al., 2013 and references herein). Concerning India, it is probably a combination of the two, since the subduction of the Neotethys beneath the southern margin of Eurasia is considered to have triggered the break-up of Gondwana around 167 Ma. But the mantle plume activity from the Jurassic onwards is considered to have been the underlying driving force, as it substantially weakened the lithosphere (Chatterjee et al., 2013).

During the Mesozoic, the Indian craton has known five episodes of continental flood magmatism accompanied by rifting (figure 3.2, 3.6) (Chatterjee et al., 2013). These episodes of continental magmatism are: (1) Karoo–Ferrar basalts, associated with the Bouvet plume (~182 Ma); (2) Kerguelen–Rajmahal basalts, associated with the Kerguelen plume (~118 Ma); (3) Morondava–St. Mary basalts, associated with the Marion plume (~88 Ma); (4) Somnath Ridge basalt (~70 Ma), and (5) Deccan–Réunion basalts, associated with the Réunion plume (~65 Ma) (Chatterjee et al., 2013).

The associated rifting phases are: (1) the separation of eastern Gondwana from western Gondwana around 167 Ma; (2) the separation of India from Antarctica-Australia around 130 Ma; (3) the separation of Madagascar from India around 90 Ma; (4) the separation of Seychelles-Laxmi Ridge Block around 70 Ma and (5) the separation of Seychelles from India around 65 Ma (Chatterjee et al., 2013).

3.4.2 The separation of eastern Gondwana from western Gondwana

The break-up of Gondwana initiated around 167 Ma (Middle to Late Jurassic). As indicated on figure 3.6b, this first rifting phase resulted in two more or less equidimensional continents: East and West Gondwana, separated by a narrow seaway. East Gondwana was comprised of Madagascar, India, Australia, and Antarctica, whilst West Gondwana consisted of Africa and South America. This rifting phase was accompanied by a first phase of continental magmatism, which resulted in the Karoo Group in South Africa and the Ferrar Group in Antarctica. This magmatism is associated with the Bouvet plume as previously mentioned (Chatterjee et al., 2013 and references herein).
3.4.3 The separation of India from Antarctica-Australia

During the second major rifting phase, India was separated from Antarctica-Australia. This phase transpired during the Cretaceous (around 130 Ma) and is associated with the opening of the Indian Ocean (around 130-125 Ma) (Powell et al., 1988; Chatterjee et al., 2013). As can be seen in figure 3.6c, both East and West Gondwana had broken up even further at the end of this rifting phase. East Gondwana had now separated into two segments, with the southern segment consisting of Antarctica and Australia and the northern segment of Sri-Lanka, India, Laxmi Ridge, Seychelles and Madagascar. The two segments were separated by the Central Indian Ocean, which was the result of spreading along the South East Indian Ridge or SEIR. This second rifting phase was associated with plume activity of the Kerguelen plume, which started around 118 Ma and continued throughout the Cenozoic. This plume activity emplaced continental flood basalts such as the Rajmahal–Sylhet basalts in eastern India and Bunbury basalts in Australia (Chatterjee et al., 2013 and references herein).

3.4.4 The separation of Madagascar from India

During the Late Cretaceous, around 88 Ma, the third major rifting phase occurred (figure 3.6d). During this rifting phase, Madagascar broke away from India-laxmi Ridge-Seychelles though the opening of the Mascarene Basin and seafloor spreading along the Central Indian Ridge (CIR). Concurrently, Australia started to separate from Antarctica. The associated continental flood basalts are mainly restricted to the Madagascar Morondava flood basalts. Contrary to the two earlier rift phases, the plume activity slightly preceded the actual rifting phase as the Marion plume became active around 93 Ma (Chatterjee et al., 2013 and references herein).

After the separation from Madagascar, India(-Laxmi Ridge-Seychelles) began its northward drift towards (Eur)Asia due to continued spreading along the CIR. This drift phase implies the subduction of the Neotethys, which is considered to have been comprised of two oceanic plates. Hence two subduction zones developed between India and Asia (figure 3.6e): (1) the Shyok–Tsangpo Trench, situated along the southern margin of Asia (now Tibet) and (2) the more southern Oman–Makran–Indus Trench, which divided the Neothethys in a northern Neotethys segment and a southern Indotethys segment. Subduction of the Indian plate beneath this second trench led to the development of the Kohistan–Ladakh (KL) Arc. Around 85 Ma, India(-Laxmi Ridge-Seychelles) would collide with this arc (Chatterjee et al., 2013 and references herein).
3.4.5 The separation of Seychelles-Laxmi ridge

Contrary to the other major rift phases, this phase of limited rifting and seafloor spreading was aborted and resulted in a failed rift, known as the Gop Rift. Moreover, the Laxmi Ridge (a NW–SE trending basement high in the Arabian Sea) had not completely separate from India at the end of this phase. This short-lived phase (73-64 Ma) is the first of two phases in the opening of the Arabian Sea. It predates the emplacement of the Deccan Traps and is probably associated with the buried Somnath volcanic platform (75-68 Ma) (Chatterjee et al., 2013 and references herein).

3.4.6 The separation of Seychelles from India

The final major rifting phase occurred around 65 Ma. As a result of seafloor spreading in which the Seychelles microcontinent was separated from India (figure 3.6f). This phase of rifting and seafloor spreading is the second phase of the opening of the Arabian Sea and is associated with the activity of the Réunion plume and emplacement of the Deccan Traps. Emplacement of the Deccan flood basalts occurred both in western India as in eastern Seychelles (or in other words at the two rifted margins) (Chatterjee et al., 2013 and references herein).

3.5 The India-Asia collision

During most of the Cretaceous, India continued its northward movement at a rate of 3-5 cm/a. Around the Cretaceous-Tertiary boundary, India’s movement was accelerated to 20 cm/a. In the Late Palaeocene to Early Eocene (55-50 Ma), India was slowed down again to 5 cm/a. This deceleration is assumed to be associated with the India-(Eur)Asia collision (Van Hinsbergen et al., 2012; Chatterjee et al., 2013 and references herein).

The India-Asia collision, and the associated closure of the Neotethys, are a direct result of the subduction of the Neotethys beneath the Lhasa Block (which is the most southern continental fragment of Mesozoic Asia). Both the timing of the India-(Eur)Asia collision and the different stages of the collision are still under debate (e.g. Klootwijk et al., 1992; DeCelles et al., 2002; Curray, 2005; Ali and Aitchison, 2008 van Hinsbergen et al., 2012). Aitchison et al. (2007) refute the generally accepted continent-continent collision-age of 55 Ma, by pointing out that the Indian and Tibetan margins, which are the two margins that partook in the collision, were not situated in close proximity to each other at the time (figure 3.7a,b). Based on the position of India and Asia in palaeogeographic reconstructions, these authors propose a Late Eocene-Early Oligocene age (more specifically around 35 Ma) for the continent-continent collision (figure 3.7c).
Figure 3.6: Palaeogeographic reconstruction of the break-up of Gondwana from a) the Late Triassic to d) the KT-Boundary. The number in figure a correspond with the locations of the future mantle plumes: 1) Bouvet plume (ca. 180 Ma); 2) Kerguelen plume (ca. 180 Ma); 3) Marion plume (ca. 88 Ma); 4) Reunion plume (ca. 65 Ma). The following abbreviations were used: Af = Africa, An = Antarctica, Au = Australia, SA = South America; CIR = Central Indian Ridge and SEIR = Southeast Indian Ridge. From Chatterjee et al. (2013). For more information see text.
In the traditional model (figure 3.8a), the (Neo)Tethys is considered as a single oceanic plate and hence, the collision was associated with a single subduction zone. In other words, there was a collision continuum from 55 Ma onwards. This traditional “one ocean two continents” view is also refuted by Aitchison et al. (2007). Instead they support a more complex model of two separate collisions, with the Tethys ocean consists of two oceanic plates (figure 3.8b). According to Aitchison et al. (2007) and Ali and Aitchison (2008), a first India-Asia contact occurred during the Late Palaeocene around 57.5 Ma when the northeastern corner of Greater India\(^1\) bumped with western Indonesia (figure 3.7b) (Ali and Aitchison, 2008). This contact was accompanied by the collision of Greater India with the intraoceanic island arc system of Shyok-Tsangpo Trench. This collision gave rise to a short-lived orogeny. This India-Asia contact is referred to as the “soft collision” (figure 3.7b) (Ali and Aitchison, 2008; Chatterjee et al., 2013).

As India continued its northward drift, it “tracked” along the western margin of Southeast Asia and eventually a second collision ensued, in which India finally collided with the southern margin of Asia (the Lhasa lock) around 35 Ma (figure 3.7c) (Aitchison et al., 2007; Ali and Aitchison, 2008). This is the so-called “hard collision”. This continental collision caused far-field deformation in Central Asia (see next section), such as crustal shortening and thickening as well as a rapid exhumation of Himalaya crystalline rocks. This processes eventually led to the development of the Himalaya-Tibet Plateau. (DeCelles et al., 2002; van Hinsbergen et al., 2012 Chatterjee et al., 2013). Note that several other variants to this model exist as well (e.g. van Hinsbergen et al., 2012).

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\(^1\)Greater India is term used for the part of the Indian plate that has been subducted underneath Tibet since the onset of Cenozoic continental collision of India and Asia (van Hinsbergen et al., 2012).
Chapter 3: The tectonic history of the Southern peninsular India

3.6 (Tectonic) Far field effects of the India-Asia collision

An obvious and well-known effect of the India-Asia collision is the development of the Himalaya-Tibet orogen. The processes driving the evolution of the orogen are considered to have occurred in two phases (Yin, 2010). The first phase consisted of crustal shortening across distinct thrust belts and had initiated from the moment India and Asia collided (55 Ma or 35 Ma). During the Miocene a major change occurred as the accommodation of the collision and crustal shortening was dominated by lateral extrusion (Chatterjee et al., 2013). During this latter phase (around 15-5 Ma), major strike-slip faults developed. Concurrently, the Himalaya-Tibet Plateau began to rise. This Plateau had reached it present-day elevation at around 8 Ma (Hodges et al., 2001; Yin, 2010, Chatterjee et al., 2013). After which gravitational spreading is assumed to have induced an east-west extension (Liu and Yang, 2003).

Besides the development of this orogen, the effects of the India-Asia collision and especially the lateral extrusion can be traced across southeast Asia. The lateral extrusion of crustal blocks along
long-lived strike-slip/shear zones (or escape tectonics) is mainly directed E-wards, to the less constraining Pacific Boundary. Contrary to the Himalayan-Tibetan Orogen, the extrusion in the rest of Southeast Asia was already dominant at around 32 Ma and lasted until approximately 17 Ma (Yin, 2010). As a result, several major strike-slip faults developed or were reactivated across Southeast Asia such as the Three Pagodas Fault and the Ranong Fault in Thailand (see section 4.1.3) (Yin, 2010; Morley et al. 2011).

Another example of far-field effects is the development of the Central Asian, North China and eastern Asian margin and the Eastern margin deformation Domains. A discussion of these domains falls outside the scope of this thesis, for more information the reader is referred to Yin (2010).
This chapter is relevant for the second set of samples taken in Peninsular Thailand. The geological history of Thailand is summarized in figure 4.1. Only the phases relevant for the current study will be considered below.

This chapter will be divided into three parts, the first part covers the tectonic history of Thailand, whilst the second part focusses on a more regional context. The final part focusses on the igneous history.

4.1 The tectonic history of Thailand

4.1.1 Introduction

The geological basement of Thailand is a part of a larger geodynamic entity known as Sundaland (Ridd et al., 2011). Currently, geographic Sundaland consists of Java, a part of Borneo, Sumatra, the Peninsular Malaysia, Thailand, eastern Myanmar, Cambodia, Vietnam, Laos and the Sunda shelf, see figure 4.2 (Ridd et al., 2011, Metcalfe, 2011). The region encompassing the three latter countries is also referred to as Indochina (figure 4.2). Sundaland’s position is indicated by the convergence zone between the Indian-Australian, Philippine and Eurasian plates (figure 4.2; Simons et al., 2007).

Tectonically, Sundaland (and SE Asia) consists of several continental blocks, volcanic arcs and suture zones representing the remnants of closed oceanic basins. Two of these continental blocks (in the literature these are also referred to as plates or terranes) and one island arc terrane can be recognized in Thailand (figure 4.3): an eastern Indochina Block, western Sibumasu Block and the Sukhothai island arc terrane (Ridd et al., 2011). From figure 4.1 it can be concluded that the Indochina and Sibumasu Blocks have played an important part in the geological history of Thailand. Note that for this thesis only the latter block is of importance.

Despite their different geological histories, both continental blocks originated from the Indian-Australian margin of Gondwana during the Late Palaeozoic (Metcalf, 2011). The two blocks separated from Eastern Gondwana and distanced themselves by drifting northwards. Eventually, the two blocks collided and fused together in the Middle Triassic. This collision provoked the assembly of Sundaland and effectively implied the complete closure of the oceanic basin of the Palaeotethys (Barber et al., 2011; Ridd et al., 2011; Searle et al., 2012). The line of closure of the Palaeotethys is currently represented by a north-south oriented suture zone in Thailand, although there is still some disagreement over its exact location (Ridd et al., 2011; Metcalfe, 2011).
Chapter 4: The tectonic and igneous history of Thailand

Figure 4.1: A schematic representation of the phases of the tectonic and igneous history of Thailand which are relevant for this thesis. Note that some ages are still under debate (see text). If this was the case, the most recently published age was chosen (based on Ridd et al., 2011).
Figure 4.2: Current Topography and plate motions (black arrows) of Southeast Asia. The most relevant aspect of this map is the location of Sundaland at the convergence zone of the Eurasian, Philippine and Indian–Australian plates (Simons et al., 2007, Metcalfe, 2011). Also the state of Tamil Nadu in southern Peninsular India, where the TN-samples were collected, is indicated. The two red boxes indicate the two study areas. The Sagaing Fault zone is marked in yellow simply for the readers convenience.

4.1.2 The tectonic evolution of the Sibumasu Block

Sibumasu is an acronym for Southwest China (Si for Sino), Burma (Bu), Malaysia (Ma) and Sumatra (Su) (Metcalfe, 2011). Metcalfe (2002) defined Sibumasu as the area west of the Palaeotethyan suture zone in north and southwest Thailand, eastern Burma and western Malaysia. The term was proposed to replace other inadequate and often cited terms such as Shan-Tai, Sinoburmalaya and West Malaya for example (Metcalfe, 2011).
Figure 4.3: A simplified representation of the Fault zones of Thailand. The Indochina (yellow) and Sibumasu Block (green) are also indicated. The red box represents the study area. "XXXX" indicate major tectonic lines or zones, the following abbreviations were used: MYF = Mae Yuam Fault; MPFB = Mae Ping Fault Belt; TPFB = Three Pagodas Fault Belt; TMF = Tha Mai Fault; RF = Ranong Fault; KMF = Khlong Marui Fault; BRS = Bentong-Raub Suture (Ridd et al., 2011).
There is a general consensus that Indochina rifted and separated from Gondwana due to the opening of the Palaeotethys ocean in the Late Silurian to Early Devonian (Barber et al., 2011; Metcalfe, 2011 and references herein). Nevertheless the Precambrian and Early Palaeozoic history of the Indochina and Sibumasu Block is not well understood. Barber et al. (2011) propose two possibilities based on where Indochina would have rifted from Gondwana. In their first scenario Sibumasu evolved as the passive rifted (north-western) margin of Gondwana in the assumption that the Indochina Block rifted from the present eastern edge of the Sibumasu Block. In their second scenario, the (present) eastern margin of Sibumasu would have remained passive during the Palaeozoic, implying that the Indochina Block rifted from a different part of Gondwana. Due to the absence of Palaeotethyan ocean-floor rocks older than Lower Devonian, neither one of these options can be excluded.

From palaeontological data, Sibumasu appeared to be a part of Gondwana from the Precambrian to the Early Permian, figure 4.4a, 4.5a (Barbet et al., 2011). However, in the Latest Carboniferous–Earliest Permian (during the Gondwana glaciation), the Sibumasu block had already started to rift from Gondwana as evidenced by rift grabens filled with glacial-marine strata (Eyles et al., 2003; Metcalfe, 2011). Nevertheless, it took until the Late Early Permian for the Sibumasu Block to become entirely separated from Gondwana due to the opening of the Mesotethys (figures 4.4b, 4.5b) (Barber et al., 2011). As previously mentioned, this separation was followed by a progressive northward drift. This drifting phase can be related to the expanding Mesotethys between Sibumasu and the remainder of Gondwana (Barbet et al., 2011).

Concurrently, the Palaeotethys was being subducted beneath or Cathaysia (which corresponds to the amalgamated South China-Indochina-East Malaya terrane), northern Pangea and North China. This subduction resulted in the development of the Sukhothai Arc (figures 4.4a, b and c) (Metcalfe, 2011). The aforementioned expansion of the Mesotethys and the subduction of the Palaeotethys, provoked the collision of the Sibumasu Block with the southern margin of the Sukhothai Arc and Indochina (figure 4.4d, 4.5c). This collision in turn resulted in to the Indosinian Orogeny (Barber et al., 2011) and is associated with granite emplacement in the Eastern and Central Granite Provinces. However, there is some discrepancy concerning the timing of this collision. Metcalfe (2000, 2011) and Barber and Crow (2009) support a Late Permian to Early Triassic age, whilst a younger (Late Triassic) collision is supported by e.g. Kamata et al. (2009) and Ridd (2009a).

A final important phase in the tectonic history of Sibumasu is the amalgamation with the West Burma block. Again, there is some debate concerning this amalgamation (Barber et al., 2011 and references herein).
A first scenario, supported by Charusiri et al. (1993), claims that the collision of the West Burma Block with the western margin of Sibumasu occurred after the closure of the Mesotethys in the Cretaceous. Consequently, this collision could have invoked the shearing along the Khlong Marui and Ranong faults in Peninsular Thailand (see next section and figure 4.3) (Charusiri et al., 1993; Barber et al., 2011).

This scenario is contradicted by Barber and Crow (2009) and Barber et al. (2011), who suggest that both West Burma and West Sumatra, which had been amalgamated since the Permian, were emplaced by strike-slip fault movements against the western margin of Sibumasu during the Latest Permian to Earliest Triassic. Hence, this emplacement could not result in the Cretaceous strike-slip movements along these faults (Barber et al., 2011). During the Miocene, the opening of the Andaman Sea separated West Burma from West Sumatra (Curray, 2005; Barber et al., 2011).

**Figure 4.4**: Section across northern and northeastern Thailand, illustrating the closure of the Palaeotethys, opening of the Mesotethys and overall tectonic evolution of Sibumasu, Indochina and the Sukhothai Arc from the Latest Carboniferous to the Early Jurassic. From Metcalfe (2011).
Figure 4.5: Plate reconstruction for the a) Early Permian: the beginning of the rifting of Sibumasu and the subduction of the Palaeotethys; b) Late Permian: Sibumasu had fully separated from Gondwana and the Mesotethys had formed. The Palaeotethys was still being subducted; c) Late Triassic: Sibumasu had collided with the Sukhothai Arc (Wakita and Metcalfe, 2005). The following abbreviations were used: EM = East Malaya; I = Indochina; L = Lhasa; NC = North China; S = Sibumasu; SC = South China; WB = West Burma; WS = West Sumatra.
4.1.3 The tectonic evolution of the Ranong and Khlong Marui Faults

Throughout the mainland of Southeast Asia, large strike-slip faults occur. These faults have been interpreted as the expression of widespread intraplate deformation (Watkinson et al., 2008). As can be seen of figure 4.3, Thailand is also strongly dissected by similar faults. In northern Thailand, the two most important strike-slip faults are the Mae Ping Fault (MPF), and the Three Pagodas Fault (TPF). Activity along these faults is associated with the Late Cretaceous to Palaeocene uplift and an orogenesis in Northern Thailand, Eastern Myanmar and Laos (Morley, 2004). In peninsular Thailand, the two most important faults are the Khlong Marui Fault (KMF) and Ranong Faults (RF). As indicated on figure 4.3, these two NE-SW tending strike-slip faults bound the study area. Although neither faults have been traced offshore, it is plausible that they have offshore analogues in the Andaman Sea and the northern Gulf of Thailand. Moreover, these analogues are presumed to be related to the extension of the Andaman Sea (see further) and the Gulf of Thailand (Watkinson et al., 2008).

4.1.3.1 Fault movement

Both the KMF and RF have a complex history ranging back to the Mesozoic (figure 4.6) (Ridd et al., 2011). Hence these fault zones represent inherited structural fabrics, that were reactivated in a protracted fashion throughout younger events. The overall tectonic history of these two faults can be summarized in two phases of dextral ductile shearing, followed by sinistral strike-slip brittle faulting. The first dextral shear phase occurred during the Late Cretaceous and was accompanied by widespread granite intrusion. The second phase transpired during the Middle Eocene (Watkinson et al., 2008; Barber et al., 2011; Morley et al., 2011; Ridd et al., 2011).

There has been some debate concerning the cause and setting of these transcurrent (shear and strike-slip) movements, as already briefly mentioned in the previous section (Barber et al., 2011). Based on their experiments, Tapponnier et al. (1982) reasoned that the fault activity could be a response to the India – Asia collision. Especially since this large-scale continent-continent collision is deemed responsible for widespread escape tectonics of lesser constraint tectonic blocks in Southeast Asia (see section 3.6). (And since this collision-induced lateral extension is bounded by large strike-slip faults and shear zone). Although the precise timing of this collision is still under debate, a Tertiary age is generally accepted (see section 3.5) (e.g. Klootwijk et al., 1992; Curray, 2005; Metcalfe 2011; van Hinsbergen et al., 2012). Consequently, the idea that this collision resulted in the aforementioned fault activity has been refuted, seeing as the strike-slip faults were already active during the Late Cretaceous (Watkinson et al., 2008; Barber et al., 2011). An alternative cause has already been discussed (and refuted) in the previous section, namely the collision of the West Burma Block with the Sibumasu Block which is now believed to have transpired in the Permian instead of during the Cretaceous (Barber and Crow, 2009; Barber et al., 2011).
Watkinson et al. (2011) tried to conclude this debate by thoroughly dating the Khlong Marui and Ranong fault zones. They concluded that both faults were activated before (approximately) 80 Ma. The authors link this first phase of dextral shearing to the Andean-type western margin of Sundaland, which had developed during the Late Cretaceous. More specifically, they claim that dextral shear stresses were conveyed from the subducting slab towards the continental margin (Watkinson et al., 2008). This Andean-type margin is the result of subduction of Neotethyan oceanic crust beneath West Burma and Western Thailand (Late Jurassic-Early Cretaceous), followed by nappe emplacement (during the Middle Cretaceous) (See further). This nappe was termed the Mawgyi nappe, after the Mawgyi Andesites2 (Mitchell, 1993; Barber and Crow, 2009; Barber et al., 2011).

After a period of inactivity both faults were reactivated during the Middle Eocene (between around 48-40 Ma), which led to the second ductile dextral deformation phase. The collision between West Burma and India is proposed as the cause of this reactivation (Watkinson et al., 2011). Hence, this reactivation can be associated with a major deformation phase in northern Thailand and eastern Myanmar as well as with major fault movement in Northern and Upper Peninsular Thailand (the Mae Ping and Three Pagodas faults). Watkinson et al. (2011) do stress that this reactivation was aided by the steeply dipping fabrics formed during the previous shear phase combined with a weakening from anatexis and the development of magmatic conduits in the shear zones.

During the remainder of the Middle Eocene the faults were inactive, only to be reactivated during the Late Eocene-Early Oligocene (37-30 Ma). This time however, the deformation is brittle and sinistral. Some authors (e.g. Watkinson et al., 2008) claim that this reactivation was responsible for the deformation of the Cretaceous and Eocene granites and the exhumation of the ductile fault core (figure 4.6). This reactivation is considered to result from continued sinistral strike-slip movement along the northern Mae Ping and Three Pagodas faults and is believed to accommodate lateral extrusion associated with the India-Asia collision and the early stages of the Himalayan orogeny (section 3.6) (Watkinson et al., 2008; 2011).

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2 The Mawgyi Andesites are an association of basaltic andesites and pillow lavas in eastern Myanmar. These andesites have been considered to be part of a Jurassic to Cretaceous intraoceanic arc. They have been correlated all the way to Sumatra, where they are referred to as the Woyla nappe (Mitchell, 1993; Barber and Crow, 2009).
Figure 4.6: A schematic representation of the tectonic history of the Khlong Marui and Ranong Fault and the magmatic history of Peninsular Thailand, the red box indicates the study area (after Watkinson et al., 2011). See figure for extra information.
4.1.3.2 Exhumation of the ductile shear zones

Watkinson et al. (2011) recognised at least five exhumed ductile fault cores along the RF and only a single one along the KMF. Since the ductile phases of the RF and KMF zones were dominated by simple shear movement, the exhumation of a fault zone (over more than 10 km) either requires a significant component of pure shear deformation or an independent mechanism such as regional uplift (Morley et al., 2011). As there are no indication for a regional uplift from e.g. the MPF zone, a component of pure shear is considered to be the underlying cause (Morely et al., 2011).

In general, Morley et al. (2011) propose five mechanisms that can cause the exhumation of a ductile shear zone (figure 4.7). In their two first scenarios (figure 7.4a,b) the geometry of the shear zone is responsible for regions of localised stress, such as jogs and bends. The jog or bend can either be releasing or restraining. Figure 4.7a represents a releasing or dilatational bend, in which case the cover material can be removed by extension. Figure 4.7b represents a restraining or compressional bend, in which case uplift and erosion could transpire. If exhumation is not restricted to jogs or bends, uplift can be explained by incorporating a transtensional (figure 4.7c) or transpressional lateral stress (figure 4.7d). Whereby exhumation occurs in a similar fashion as at a dilatational jog and a compressional jog respectively, but on a larger scale (Morley et al., 2011).

Another possibility is that the reactivation of the fault zone after a period of inactivity was responsible for the exhumation, or a combination of the above (figure 4.7e), which would result in a two-stage exhumation. In figure 4.7e, the ductile core was only partially exhumed during a first stage. After a period of quiescence, the uplift was completed exhumed due to brittle faulting (Morley et al., 2011).

According to Watkinson et al. (2008); Morley et al. (2011) and Watkinson et al. (2011), the exhumation of the ductile shear zones along the RF and KMF zones requires a similar two-stage process. According to these authors, the final uplift of the two dextral ductile shear zones transpired during the Late Eocene-Oligocene brittle deformation phase (figure 4.6). Hence, this final uplift phase occurred simultaneous with sinistral deformation and exhumation (between 44-20 Ma) along the TPF and MPF fault zones (Upton, 1999; Watkinson et al., 2008; 2011).

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3In this case, the definition given by Watkinson et al. (2011) for ductile fault cores is followed. They define exhumed ductile fault cores as exposed slivers of ductile fault rocks that are bounded by brittle faults and surrounded by non-metamorphic rocks.

4In this context, exhumation is considered as the emplacement of rocks to shallower crustal levels.
Figure 4.7: Five possible mechanisms for the exhumation of a ductile shear core: a) releasing or dilatational bend; b) restraining or compressional bend; c) transtension; d) transpression; e) a two-stage exhumation: during a first stage the core was only partially exhumed. After a period of quiescences, the uplift was completed exhumed due to brittle faulting (from Morley et al., 2011).
4.2 The tectonic history of Thailand in a more regional context.

Searle and Morley (2011) divided South-East Asia in 13 Cenozoic tectonic provinces, including the Indochina Block, the Andaman Basin and Central Thailand Basins, of which only Andaman Sea is important for this thesis.

As indicated on figure 4.2, the Andaman Sea is part of the Indian Ocean and is situated off the west coast of Thailand. Depending on the author, the Andaman Sea is defined slightly different (e.g. Curray, 2005; Khan and Chakraborty, 2005). Here the Andaman Sea is considered to be the watermass located to the north of Sumatra, to the south of Myanmar, to the west of Peninsular Thailand and to the east of the Andaman Nicobar Islands. In this thesis, the Andaman basin will be treated separately from the Adaman Sea. The (Central) Andaman Basin is a young back-arc basin which is accommodates in the Andaman Sea. This back-arc basin is associated with the (oblique) subduction between the overriding Southeast Asian plate and the downgoing Indian-Australian plate (figure 4.8a; Curray, 2005).

4.2.1 The Andaman Sea

The Andaman Sea is a part of the larger Burma-Java subduction complex (Chakraborty and Khan, 2009) and located above and behind the active Sunda (or Java) trench (figure 4.8a,b). This active subduction complex is well-known by the larger public due to the tragic Boxing Day Sumatra earthquake and tsunami of 2004. Both Tethyan and Indian oceanic crust have been subducted during much of the Mesozoic and Cenozoic by this subduction system and its precursors (Curray, 2005; Khan and Chakraborty, 2005 and references herein). As is generally the case for oblique subduction, it is associated with a partitioning motion: an arc-parallel sliver fault which accommodates the lateral component and a normal component which is related to subduction (Chakraborty and Khan, 2009).

In this case the Sagaing Fault, which continues as the West Andaman fault (figure 4.8b) forms the arc-parallel sliver fault (on the overriding plate) (Chakraborty and Khan, 2009). Eventually, the oblique subduction will invoke basin formation through back-arc extension, amongst which the active Andaman back-arc basin and Mergui Basin (which is an offshore continuation of the North Sumatra Basin) (figure 4.8b; Curray, 1989, Raju et al., 2004) are main examples. Generally, back-back basins are the result of flexure (or stretching) in the overriding plate combined with secondary convection which accompanied the ocean-continent convergence. Therefore the Andaman Basin differs from typical back-arc basins because it formed by transtension and oblique subduction (Curray, 2005).

5 The Sagaing Fault is a major fault in Myanmar located between the Indian plate and the Southeast Asian plate. It connects the spreading centres of the Andaman Sea (Tsutsumi and Sato, 2009).
4.2.1.1 Morphotectonic elements of the Andaman Sea

The Andaman Sea can be subdivided into three main morphotectonic elements (Raju et al., 2004; Curray, 2005): an accretionary prism, a central basin and an eastern zone of continental crust. (See figure 4.8a,b for a morphotectonic map and cross-section).

The Andaman-Nicobar ridge forms as the accretionary (outer-arc) prism of the Sunda subduction zone (Curray, 2005; Chakraborty and Khan, 2009; Cochran, 2010), with the Andaman Islands as its subareal part (Searle and Morley, 2011). As can be seen in fig 4.8b, this trench paralle ridge is situated in the northern part of the subduction zone. The Andaman outer arc was built up during the Cretaceous to Miocene period. Structurally, it consists of tectonically imbricated fold-thrust packages, including accretionary sediments and ophiolite slices (Pal et al. 2003; Chakraborty and Khan, 2009).

The central basin encompasses a north-south trending fault system. This basin can be further subdivided into a forearc basin, a magmatic arc and a back-arc basin (the Andaman Basin) with a back-arc spreading centre (which has been active since approximately 4 Ma) (Raju et al., 2004; Curray, 2005). (Note that spreading in the Andaman Sea region, not the Andaman back-arc spreading centre, is considered to have started in the Early to Middle Miocene). During the Neogene, these parts were subjected to strike-slip sliver faulting and transtensional extension (Curray, 2005).

The eastern zone can be seen as the previously defined Sibumasu Block, meaning it contains (amongst others) the continental margin of the eastern Andaman Sea (Raju et al., 2004; Curray, 2005).
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Figure 4.8: a) A cross-section trough the Andaman Sea, the major tectonic elements are indicated on the cross-section (Chakraborty and Khan, 2009); b) A tectonic map of the Andaman Sea Region and adjacent southern Myanmar and northern Sumatra. The WAF stands for West Andaman Fault (Curray, 2005). The red line indicates the position of cross-section a. The following abbreviations are used: Ba = Barren Island, Co = continental crust refraction determination, DF = Diligent Fault, EMF = Eastern Margin Fault, GC = Great Cocos Island, KMF = Khlong Marui Fault, LC = Little Cocos Island, MPF = Mae Ping Fault, NS = North Sentinel Island, NSR = North Sumatra Ridge, Oc = oceanic crust refraction determination, Pr = Preparis Island, R = Rutland Island, RF = Ranong Fault, RR = Ranong Ridge, SEU = Seulimeum strand of SFS, SFS = Sumatra Fault System, SSF = Shan Scarp Fault, TPF = Three Pagodas Fault and WAF = West Andaman Fault.
4.2.1.2 The opening of the Andaman Sea

Both the formation of the passive margin of Peninsular Thailand and the associated opening of the Andaman Sea can be related to effects of oblique subduction of the Indo-Australian plate under the Southeast Asian plate, which include strike-slip faulting, back-arc extension and basin formation (Curry, 1989; Raju et al., 2004; Cochran, 2010).

Subduction in the larger Burma-Java subduction complex probably started in the Early Cretaceous, at least by the time India had broken off from Australia and Antarctica in eastern Gondwana (Curry, 2005; Chakraborty and Khan, 2009). However, the Andaman–Sumatra margin has only has been active since the middle Tertiary (Chakraborty and Khan, 2009). Assuming the model of Aitchison et al. (2007) a first India-Asia contact probably occurred around 57.5 Ma with the soft-collision phase (see section 3.5). This contact resulted in a clockwise rotation of India, which invoked a clockwise rotation of the subduction zone. This rotation in turn led to an increase in the subduction-obliquity. After the initial contact with Sumatra, India tracked along western South-East Asia, resulting in an amalgamation with the West Burma Block and eventually causing the so called “hard collision” around 35 Ma (or during the Late Eocene to Early Oligocene) (Aitchison et al., 2007; Ali and Aitchison, 2008). The transtension caused by the amalgamation of the West Burma Block and India and the subsequent drifting of India-Western Burma resulted in the development of the passive western continental margin of Peninsular Thailand and Burma (Searle and Morley, 2011). Meanwhile, the obliquity of the subduction zone increased even further. At the end of this period (at around 32 Ma), the oblique subduction lead to the formation of sliver faults in the Sagaing and West Andaman fault systems (Raju et al., 2004; Curry, 2005). This coincides with the second phase of dextral shearing along the RF and KMF within our study area. Concurrently, both the Mergui and North Sumatra rift basins were formed during the Early Oligocene (around 32 Ma) due to a combination of rift extension of the Southeast Asian continental crust and (increased) strike-slip motion along the Sagaing and West Andaman Fault (Curry, 2005; Searle and Morley, 2011). Continental extension of these basins lasted until the Middle Miocene (Searle and Morley, 2011). According to Morley et al. (2011) the deceleration of this extension is due to spreading in the Andaman back-arc area. Miocene post-rift deposition facies in the Mergui Basins show a transition from predominantly deepwater to more shallow marine environments (Searle and Morley, 2011). Sedimentary structures indicate deltaic deposits, which indicate erosion of and sedimentary input from Peninsular Thailand during the Late Miocene (Curry, 2005; Searle and Morley, 2011).

During the Middle Miocene (around 11 Ma), the first phase of the opening of the Andaman Sea transpired (figure 4.9a,b), which was follow by a second extensional phase during the Late Miocene to Early Pliocene (around 4-5 Ma, figure 4.92c). Both extensional phases were the result of an
increase in subduction rate and subduction angle in the Indian-Australian plate (Khan and Chakraborty, 2005). These events in turn, are considered to cause trench retreat combined with extensive extension, eventually inducing back-arc spreading around 5 Ma (Khan and Chakraborty, 2005). The associated (Andaman) spreading centre is still active today (figure 4.9d) (Curry, 2005; Morley et al., 2011; Searle and Morley, 2011). Moreover, these three processes were accompanied by the opening of the Andaman back-arc Basin (around 5-4 Ma).

![Figure 4.9: Schematic representation of the opening of the Andaman Sea (Khan and Chakraborty, 2005): a) the situation before the opening of the Andaman Sea, the region is dominated by a phase large scale forearc subsidence; b) the first extension phase, which consisted of stretching and rifting; c) the initiation of the present spreading centre in the Andaman Sea; d) the present situation. The indicated region more or less corresponds to the area shown of figure 4.8b.]

### 4.3 Igneous history of Peninsular Thailand.

#### 4.3.1 Introduction

The granites of Thailand form a ca. 1500 km long sinuous complex of batholights and plutons (Charusiri et al., 1993; Cobbing, 2011). This complex is a part of the larger Southeast Asia Tin belt, which in fact forms a southeastern extension of the Alpine-Himalayan mountain range (Charusiri et al., 1993). This belt is one of the most productive and largest sources of metallic mineral deposits, such as tin and copper, in the world (Searle, 2012). These granites can be subdivided into three or four (depending on the authors) Provinces based on geological setting, lithological characteristics,
and geochronology (e.g. Cobbing et al., 1986; Charusiri et al. 1993 and references herein). This section only focusses on the subdivision specific for Thailand.

In Thailand, three Provinces were recognised: a Western, Eastern and a Central Granite Province (Charusiri et al., 1993, Cobbing, 2011) (figure 4.10a,b). All three Provinces form more or less linear north-south belts. Each Province is characterised by distinctive plutons (sometimes composed of smaller batholiths and stocks), with specific granite types, emplacement ages and patterns of mineralisation (Cobbing et al., 1986; Cobbing 2011). According to Charusiri et al. (1993), the Provinces are mainly Mesozoic in age. Whereby the Western Province was formed during the Late Cretaceous (to Middle Palaeogene), whilst the other two are considered to be mainly of Triassic age (Cobbing 2011). Besides a difference in geochronological evolution, these provinces also mark different geotectonic environments (Charusiri et al., 1993). It should be noted that, at least in lower Peninsular Thailand, much of the information concerning the chronology is based on relative bench marks and that absolute age information is relatively scarce.

Peninsular Thailand is dominated by the Western Province, which stretches out into Myanmar and Sumatra (Cobbing, 2011) (figure 4.10). In Thailand, the granites of this Province lie in between (and in the immediate area surrounding) the KMF and RF (Watkinson et al., 2008). In other words, they form a curvature which follows the Myanmarese border (Cobing et al., 1986). The Eastern Province spreads out from Laos to Peninsular Malaysia and Indonesia (Charusiri et al. 1993). In Thailand, this Province extends from north (Chiang Rai and Loei) to southernmost Peninsular Thailand (Cobbing, 2011). The central Province, which is the largest of the three, covers most of Thailand: it spreads out south from the Myanmarese border towards Bangkok, then it runs further southwards across the Gulf of Thailand towards Peninsular Malaysia (Charusiri et al. 1993; Cobbing, 2011). Cobbing et al. (1986) further divide the Granites of the Central Province into a North Thai Province and a Main Range Province (figure 4.10a).

So to conclude, these granites are omnipresent, except for in the north-eastern part of the country, more specifically the (volcanic) Khorat Plateau (figure 4.3) (Charusiri et al., 1993).

In general, the granite Provinces are the result of subduction and accretion related magmatism. The closure of the Palaeotethys (and the resulting Indosinian orogeny, section 4.1.2) in the Permian or Triassic, as well as the closure of the Neotethys during the Mesozoic and Palaeogene are main driving events in this framework (Ridd et al., 2011; Searle et al., 2012). All samples studied in this work belong to the Western Province (figure 4.11a, b and Table 4.1), only these granites will be considered further.
4.3.2 Western Province granites

4.3.2.1 General characteristics

The granites of the Western (or Peninsular Thailand-Burma; Cobbing et al., 1986) Province form a chain of small granitic bodies such as batholiths and isolated plutons and stocks that spread out from Phuket into Myanmar (figure 4.10, 4.11) (Cobbing, 2011). In comparison to the other provinces, the granites (or more correctly the granitoids) of the Western Province enclose a much smaller area in Thailand (Charusiri et al., 1993). Geochemically, these granites vary from Sn and/or W-bearing S-types to barren (or Sn, W-poor) I-types. Compositonally, they range from granodiorite to syenogranite and they are strongly potassic (Cobbing, 2011).

In general, the larger batholiths consist of S-type Sn-bearing granites, whilst the smaller plutons consist of I-type granites (Cobbing et al. 1986 and references therein) (Figure 4.10a). As previously mentioned, they are mainly of Late Cretaceous to Middle Palaeogene age (80-50 Ma), making them approximately 200-250 Ma younger than their Carboniferous to Early Permian Kaeng Krachan Group host rocks, which mainly consist of pebbly mudstone or sandstone (table 4.1; Cobbing et al., 1986; Cobbing, 2011). The mineralisation pattern typically consists of greisen(-bordered) vein swarms and
Chapter 4: The tectonic and igneous history of Thailand

pegmatites rich in cassiterite (SnO$_2$) and wolframite ((Fe,Mn)WO$_4$) (Cobbing et al., 1986). Extensive kaolinisation is not uncommon (Cobbing et al., 1986: Searle et al., 2012). Cobbing (2011) provides a review of the different batholith bodies (figure 4.11a,b). In the following paragraph, only those relevant to this study are mentioned. The Phuket-Takua Pa Batholith, in which most of the samples are situated (table 4.1; figure 4.11b), is the most southerly batholith. It contains both S- and I-type granites. As indicated in table 4.1, other characteristics can vary strongly. The Khao Phanom Bencha Granite (fig. 4.11b) is an isolated I-type granite. It is situated to the east of the Khlong Marui Fault zone. As can be derived from table 4.1, these greyish granites can contain biotite-hornblende and some pyroxene. They have a medium-grain size and their texture is equigranular (Cobbing, 2011).

This Granite Province can be further subdivided based on geochemical or metallogenic characteristics (Cobbing, 2011).

4.3.2.2 S-type versus I-type granites and their tectonic implications

Cobbing et al. (1986), Cobbing 2011 and Searle et al. (2012) describe an evolutionary series of granites from a parental mafic precursor I-type through leucocratic main phase granites to S-type batholiths. The latter are often related to tin (Sn), tungsten (W) and Rare Earth Element (REE) mineralisations (Charusiri et al., 1993: Searle et al., 2012).

A possible explanation is given by Searle et al. (2012), who state that I-type granites became more felsic due to an increased crustal influence. The tin mineralisation is considered to be the result of the later stages of plutonism along the greisen-type vein networks (Searle et al., 2012).

Generally, I-type granites can be seen as being metaluminous, whilst S-type granites can be seen as peraluminous. I-type granites typically contain hornblende, titanite and biotite and have low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (Frost et al., 2001: Cobbing 2011 and Searle et al., 2012). They are often considered to have originated from the remelting of igneous sources (Ridd, 2009a), which explains the low $^{87}\text{Sr}/^{86}\text{Sr}$-value. I-type granites can be of significant economic value as they are often associated with porphyry copper (Cu)-molybdenum (Mo) deposits, as well as epithermal ores, such as Au, Ag, Pb, Fe and Zn (Robb, 2004; Searle et al., 2012). Tectonically, these granites are related to with magmatism at a continental margin in Andean-type subduction setting zone, see further (Barber et al., 2011; Cobbing, 2011; Searle et al., 2012).
Figure 4.11: a) The Granite Provinces and tectonic divisions of southern Thailand, the red square indicated the study area; b) a focus on the exact location of the samples. (After Cobbing, 2011 and Cobbing et al. 1992).
**Table 4.1:** Overview of the samples and their larger geological entities. The third column was based in the map shown in the Appendix. Following abbreviations were used: Al = allanite; Ap =apatite, Bt = biotite, Hbl = hornblende, K-fs = alkali-feldspar Mus = muscovite, Or = orthoclase, Px = pyroxene, Sn = tin, Ttn = titanite, Tur = tourmaline and Zr = zircon. (adamellite is a quartz monzonite).

<table>
<thead>
<tr>
<th>Sample</th>
<th>Field description</th>
<th>Information based on the geological map (edition 1893), Department of mineral resources.</th>
<th>Information based on Cobbing (2011)</th>
<th>Age (based on the geological map)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ST-33</td>
<td>Granite boulder</td>
<td>Porphyritic Bt granite, granodiorite Hbl-adamellitite and fine-grained Mus-Tur granite</td>
<td>Khao Phanom Bencha Batholith Primary Texture Granitoid (pink-grey) Or,(green-brown) Hbl, (Bt),(Px),Ttn, Zr and Ap I-type</td>
<td>Cretaceous-Tertiary</td>
</tr>
<tr>
<td>ST-34</td>
<td>Sandstone host rock</td>
<td>grey, brown, greenish-gray sandstone, bedded chert and conglomerate, mudstone and greywacke, sandstone</td>
<td>/</td>
<td>Carboniferous-Permian (Kaeng Krachan Group)</td>
</tr>
<tr>
<td>ST-35</td>
<td>Sheared granite</td>
<td>Porphyritic Bt granite, granodiorite Hbl-adamellite and fine-grained Mus-Tur granite</td>
<td>Phuket-Takua Pa Batholith: Ban Lam Ru Primary Texture Granitoid (white) Microcline, (dark-brown) Bt, Ttn, Al, Zr and Ap S-type</td>
<td>Cretaceous-Tertiary</td>
</tr>
<tr>
<td>ST-36</td>
<td>Granite</td>
<td>Porphyritic Bt granite, granodiorite Hbl-adamellite and fine-grained Mus-Tur granite</td>
<td>Phuket-Takua Pa Batholith: Ban Lam Ru Primary Texture Granitoid (white) Microcline, (dark-brown) Bt, Ttn, Al, Zr and Ap S-type</td>
<td>Cretaceous-Tertiary</td>
</tr>
<tr>
<td>ST-37</td>
<td>Granite</td>
<td>Porphyritic Bt granite, granodiorite Hbl-adamellite and fine-grained Mus-Tur granite</td>
<td>Phuket-Takua Pa Batholith: Ban Lam Ru Primary Texture Granitoid (white) Microcline, (dark-brown) Bt, Ttn, Al, Zr and Ap S-type</td>
<td>Cretaceous-Tertiary</td>
</tr>
<tr>
<td>ST-38</td>
<td>Granite</td>
<td>Porphyritic Bt granite, granodiorite Hbl-adamellite and fine-grained Mus-Tur granite Sn-deposit</td>
<td>Phuket-Takua Pa Batholith: Ban Lam Ru Primary Texture Granitoid (white) Microcline, (dark-brown) Bt, Ttn, Al, Zr and Ap S-type</td>
<td>Cretaceous-Tertiary</td>
</tr>
<tr>
<td>ST-39</td>
<td>Granite</td>
<td>Porphyritic Bt granite, granodiorite Hbl-adamellite and fine-grained Mus-Tur granite</td>
<td>Phuket-Takua Pa Batholith: Ban Lam Ru Primary Texture Granitoid (white) Microcline, (dark-brown) Bt, Ttn, Al, Zr and Ap S-type</td>
<td>Cretaceous-Tertiary</td>
</tr>
<tr>
<td>ST-40</td>
<td>Weathered coarse grained granite</td>
<td>Porphyritic Bt granite, granodiorite Hbl-adamellite and fine-grained Mus-Tur granite</td>
<td>Phuket-Takua Pa Batholith: Lam Pi Primary Texture Granitoid (Grey) K-Fs megacryst, perthite and Microcline (Red-brown) Bt, Px, Zr and Ap S-type</td>
<td>Cretaceous-Tertiary</td>
</tr>
<tr>
<td>ST-41</td>
<td>Granite</td>
<td>Porphyritic Bt granite, granodiorite Hbl-adamellite and fine-grained Mus-Tur granite</td>
<td>Phuket-Takua Pa Batholith: Ao Nai Yang. Two options: Primary Texture Granitoid (white) Perthite, (dark-brown) Bt, Al, Zr and Ap or Two phase texture (white) Or, (dark-brown) biotite, Mus, Zr and Ap I-type</td>
<td>Cretaceous-Tertiary</td>
</tr>
<tr>
<td>------</td>
<td>--------</td>
<td>----------------------------------------------------------------------------------</td>
<td>-----------------------------------------------------------------</td>
<td>------------------</td>
</tr>
<tr>
<td>ST-42</td>
<td>Granite</td>
<td>Porphyritic Bt granite, granodiorite Hbl-adamellite and fine-grained Mus-Tur granite</td>
<td>Phuket-Takua Pa Batholith: Kata Beach Primary Texture Granitoid, weakly deformed (Grey) K-Fs megacryst, microcline, (red-brown) Bt, Ttn, Al, Zr and Ap S-type</td>
<td>Cretaceous-Tertiary</td>
</tr>
<tr>
<td>ST-43</td>
<td>Granite</td>
<td>Porphyritic Bt granite, granodiorite Hbl-adamellite and fine-grained Mus-Tur granite</td>
<td>Phuket-Takua Pa Batholith: Kata Beach Primary Texture Granitoid, weakly deformed (Grey) K-Fs megacryst, microcline, (red-brown) Bt, Ttn, Al, Zr and Ap S-type</td>
<td>Cretaceous-Tertiary</td>
</tr>
<tr>
<td>ST-44</td>
<td>White granite</td>
<td>Porphyritic Bt granite, granodiorite Hbl-adamellite and fine-grained Mus-Tur granite</td>
<td>Phuket-Takua Pa Batholith: Ban Katha Primary Texture Granitoid, (Grey), microcline, Bt (red-brown), Ttn, Al, Zr and Ap S-type</td>
<td>Cretaceous-Tertiary</td>
</tr>
<tr>
<td>ST-45</td>
<td>Granite</td>
<td>Porphyritic Bt granite, granodiorite Hbl-adamellite and fine-grained Mus-Tur granite</td>
<td>Phuket-Takua Pa Batholith: Khao Prathiu Primary Texture Granitoid (grey) Or, (green- brown) Bt, green Hbl, Ttn, Zr and Ap I-type</td>
<td>Cretaceous-Tertiary</td>
</tr>
<tr>
<td>ST-53</td>
<td>Quartzite</td>
<td>Kaeng Krachan Group: greenish-grey, grey, brown sandstone, shale, pebbly sandstone, mudstone and greywacke, sandstone</td>
<td>/</td>
<td>Carboniferous-Permian Kaeng Krachan Group</td>
</tr>
<tr>
<td>ST-54</td>
<td>Granite</td>
<td>Porphyritic Bt granite, granodiorite Hbl-adamellite and fine-grained Mus-Tur granite</td>
<td>Phuket-Takua Pa Batholith: Ban Lam Ru Primary Texture Granitoid (white) Microcline, (dark-brown) Bt, Ttn, Al, Zr and Ap S-type</td>
<td>Carboniferous-Permian</td>
</tr>
</tbody>
</table>
S-type granites, on the other hand, often contain (Al and Ti-rich) garnet, muscovite and biotite (due to the higher Al-content, biotite will be richer in the S-types) (Frost et al., 2001, Searle et al., 2012). They typically have high initial $^{87}\text{Sr}/^{86}\text{Sr}$-ratios. This type of granite can also be of economic value due to their association with Sn- and W-mineralizations (Searle et al., 2012).

S-type granites are often considered to result from the melting or remelting of a crustal source, generally (meta)sedimentary rocks (Ridd, 2009a), explaining the higher $^{87}\text{Sr}/^{86}\text{Sr}$-ratio. However, Cobbing (2011) claims that the initial $^{87}\text{Sr}/^{86}\text{Sr}$-ratios of both the I- and S-type granites are low, indicating a mantle origin. Hence, these S-type granites are linked to oceanic subduction, more specifically they can occur in a transpressional continent margin volcanic arc setting (Cobbing et al., 1986; 2011).

Although various theories concerning the cause of this Cretaceous-Eocene magmatism exist (e.g. Charusiri et al., 1993; Mitchell, 1993), the current consensus is that this magmatism originated from the Andean type subduction of the Neotethyan oceanic crust under western Thailand and thus also under Sibumasu (see above) (figure 4.12), which resulted in the collision of the Neotethyan Mawgyi island arc with the Sibumasu Block during the Cretaceous (Crow and Khin Zaw, 2011; Barber et al., 2011; Watkinson et al., 2011 and references herein). This collision invoked compression, crustal thickening and overthrusting of the (Mawgyi) nappe over West Burma and Western Thailand (Barber et al., 2011) (see section 4.1.3.1), which in turn caused anatexis and hence granite emplacement (Barber et al., 2011; Crow and Khin Zaw, 2011). The presence of volatile components, derived from the subducting slab, probably contributed to the partial melting of the asthenospheric mantle, which would explain the low initial $^{87}\text{Sr}/^{86}\text{Sr}$ values. As indicated on figure 4.1, the S- and I-type granite igneous episode, which led to the emplacement of the Western granites, is concurrent with the first dextral shear phase of the KM and RF zones. As could be expected, since this Andean-type subduction is considered to trigger the first dextral shear phase of the RF and KMF (see section 4.1.3).
Figure 4.12: Palaeogeographic reconstructions of the subduction of the Neotethys underneath Thailand during a) the Early Cretaceous and b) the Late Cretaceous (from Metcalfe, 2013). The following abbreviations were used: EM = East Malaya, I = Indochina, L = Lhasa, PA = Incipient East Philippine ar, S = Sibumasu, SC = South China, SWB = Southwest Borneo, WB = West Burma, WSu = West Sumatra. Again, our study area is located in the Sibumasu Block (S).
5.1 Overview of the analysed samples

Sample description, geographical location and methods applied to each sample are listed in table 5.1. For this thesis, samples were taken from two distinct areas (figure 5.1): (1) an E-W profile in the Tamil Nadu State in southeastern India (TN-samples) and (2) a more or less N-S transect in (Lower) Peninsular Thailand (ST-samples), which was situated in between the Ranong and Khlong Marui faults. During the sampling of both sections, special attention was given to lithologies that were suitable geochronological applications.

In February 2010, the TN-samples used in this study were sampled by Dr. Stijn Glorie, Prof. Johan De Grave and Dr. Tejpal Singh. The samples are all located within the vicinity of the city of Chennai (formerly known as Madras) and located in the tectonic unit of the Madras Block. According to the geological map of the Tamil Nadu region (figure 3.1b) the samples are mainly Archean migmatites or granites. Although recently crystallisation ages and ages of metamorphism of both the Salem and Madurai Blocks have been published (e.g. Saitoh et al., 2011; Plavsa et al., 2012), this is not the case for the Madras Block (see section 3.2).

The ST-samples were collected during three weeks of fieldwork in the period of January-February 2011. This fieldwork was performed by Dr. Stijn Glorie, Prof. Dr. Johan De Grave, Dr. Pitsanupong Kanjanapayont and Prof. Dr. Punya Charusiri. The 13 granites that were used in this thesis were sampled from two distinct batholiths, the Khao Phanom Bencha and the Phuket-Takua Pa Batholith, which have already been extensively discussed in section 4.3. These Cretaceous to Tertiary granites (table 4.1) intruded a Carboniferous to Early Permian of pebbly mudstone or sandstone host rock, which belongs to the Kaeng Krachan Group. Samples ST-34, ST-53 represent this host rock. The magmatism responsible for the emplacement of these granites is associated with the activity along the Ranong and Khlong Marui fault zones, which bound this study area (figure 4.3). Pictures of samples can be found in appendix A. Table 5.2 gives an overview of the geochronological information already available for these batholiths.
Table 5.1: Overview of the location, lithology and applied methods (with AFT = Apatite fission track dating and/or thermochronology; chemical analysis = major element analysis by means of ICP-OES). ST stands for southern Thailand, TN for Tamil Nadu.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Country</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Altitude</th>
<th>Locality</th>
<th>Lithology</th>
<th>AFT-Dating</th>
<th>AFT-thermochronology</th>
<th>Thin section analysis</th>
<th>Chemical analysis</th>
</tr>
</thead>
<tbody>
<tr>
<td>TN-35</td>
<td>India</td>
<td>N 12°29'14.64&quot;</td>
<td>E 79°07'19.7&quot;</td>
<td>179 m</td>
<td>Just South of Polur</td>
<td>Granite/ Diorite/ gneiss</td>
<td>V</td>
<td>V</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TN-36</td>
<td>India</td>
<td>N 12°28'6.9&quot;</td>
<td>E 79°22'60.0&quot;</td>
<td>175 m</td>
<td>Nedugunnam Village Road to Vandvasi</td>
<td>Diorite/ Charnokite</td>
<td>V</td>
<td>v</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TN-37</td>
<td>India</td>
<td>N 12°31'52.9&quot;</td>
<td>E 79°35'15.7&quot;</td>
<td>109 m</td>
<td>East of Vandvasi-road to Cheyur</td>
<td>Granite</td>
<td>V</td>
<td>v</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TN-38</td>
<td>India</td>
<td>N 12°36'47.5&quot;</td>
<td>E 80°3'25.8&quot;</td>
<td>45 m</td>
<td>~25 km West of Mammalapuram</td>
<td>Weathered granite</td>
<td>V</td>
<td>v</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-33</td>
<td>Thailand</td>
<td>N 08°13'14.0&quot;</td>
<td>E 098°58'14.3&quot;</td>
<td>73 m</td>
<td>Close to Khao Phanom</td>
<td>Granite</td>
<td>V</td>
<td>v</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-34</td>
<td>Thailand</td>
<td>N 08°13'14.0&quot;</td>
<td>E 98°58'14.3&quot;</td>
<td>73 m</td>
<td>Same as ST-33</td>
<td>Sandstone host-rock</td>
<td>V</td>
<td>V</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-35</td>
<td>Thailand</td>
<td>N 08°34'07.8&quot;</td>
<td>E 98°32'32.5&quot;</td>
<td>159 m</td>
<td>North of Phang Nga along road 4</td>
<td>Sheared granite</td>
<td>V</td>
<td>v</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-36</td>
<td>Thailand</td>
<td>N 08°36'46.1&quot;</td>
<td>E 98°32'58.1&quot;</td>
<td>320 m</td>
<td>3-4 km North of Song Phraek</td>
<td>Granite</td>
<td>V</td>
<td>v</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-37</td>
<td>Thailand</td>
<td>N 08°37'01.4&quot;</td>
<td>E 98°14'30.7&quot;</td>
<td>16 m</td>
<td>Along road 4 to Khao Lak</td>
<td>Granite</td>
<td>V</td>
<td>v</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-38</td>
<td>Thailand</td>
<td>N 08°29'48.5&quot;</td>
<td>E 98°17'00.9&quot;</td>
<td>45 m</td>
<td>Close to Khanim waterfall</td>
<td>Granite</td>
<td>V</td>
<td>v</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-39</td>
<td>Thailand</td>
<td>N 08°26'08.1&quot;</td>
<td>E 98°18'28.2&quot;</td>
<td>40 m</td>
<td>Close to Thai Mueang</td>
<td>Granite</td>
<td>V</td>
<td>v</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-40</td>
<td>Thailand</td>
<td>N 08°16'32.0&quot;</td>
<td>E 98°19'24.7&quot;</td>
<td>53 m</td>
<td>At junction Phuket-Phang Nga</td>
<td>Granite</td>
<td>V</td>
<td>V</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-41</td>
<td>Thailand</td>
<td>N 08°05'24.0&quot;</td>
<td>E 98°19'32.9&quot;</td>
<td>41 m</td>
<td>Road to Phuket airport, Thalang College</td>
<td>Granite</td>
<td>V</td>
<td>v</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-42</td>
<td>Thailand</td>
<td>N 07°54'55.0&quot;</td>
<td>E 98°17'31.8&quot;</td>
<td>2 m</td>
<td>North Patong Beach</td>
<td>Granite</td>
<td>V</td>
<td>v</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-43</td>
<td>Thailand</td>
<td>N 07°53'00.0&quot;</td>
<td>E 98°16'44.5&quot;</td>
<td>61 m</td>
<td>South Patong Beach</td>
<td>Granite</td>
<td>V</td>
<td>v</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-44</td>
<td>Thailand</td>
<td>N 07°48'44.3&quot;</td>
<td>E 98°18'00.8&quot;</td>
<td>56 m</td>
<td>Close to Kata Noi Beach</td>
<td>White granite</td>
<td>V</td>
<td>v</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-45</td>
<td>Thailand</td>
<td>N 07°55'45.1&quot;</td>
<td>E 98°22'23.7&quot;</td>
<td>66 m</td>
<td>Close to Phuket City</td>
<td>Granite</td>
<td>V</td>
<td>v</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-53</td>
<td>Thailand</td>
<td>N 09°07'34.2&quot;</td>
<td>E 98°18'00.8&quot;</td>
<td>95 m</td>
<td>Along road 4</td>
<td>Quartzite</td>
<td>V</td>
<td>v</td>
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<tr>
<td>ST-54</td>
<td>Thailand</td>
<td>N 08°39'30.7&quot;</td>
<td>E 98°14'44.4&quot;</td>
<td>1 m</td>
<td>Khao Lak Beach</td>
<td>Granite</td>
<td>V</td>
<td>v</td>
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### Table 5.2: An overview of the geochronological data available for Peninsular Thailand, TFT = titanite FT, WR = whole rock (Cobbing et al., 2011 and references herein).

<table>
<thead>
<tr>
<th>Batholith or locality</th>
<th>Dating method</th>
<th>Age (Ma)</th>
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<tr>
<td></td>
<td>WR Rb-Sr</td>
<td>114 ± 10</td>
</tr>
<tr>
<td></td>
<td>WR Rb-Sr</td>
<td>109 ± 40</td>
</tr>
<tr>
<td></td>
<td>WR Rb-Sr</td>
<td>98 ± 7</td>
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<tr>
<td></td>
<td>WR Rb-Sr</td>
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<tr>
<td></td>
<td>WR Rb-Sr</td>
<td>82 ± 4</td>
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<tr>
<td></td>
<td>WR Rb-Sr</td>
<td>78 ± 2</td>
</tr>
<tr>
<td></td>
<td>WR Rb-Sr</td>
<td>79 ± 4</td>
</tr>
<tr>
<td>Phuket-Takua Pa Batholith</td>
<td>K-Ar</td>
<td>59</td>
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<tr>
<td></td>
<td>K-Ar</td>
<td>57.4 ± 1.5</td>
</tr>
<tr>
<td></td>
<td>K-Ar</td>
<td>54 ± 2</td>
</tr>
<tr>
<td></td>
<td>K-Ar</td>
<td>56</td>
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<tr>
<td></td>
<td>TFT</td>
<td>55-57</td>
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<tr>
<td></td>
<td>TFT</td>
<td>54 ± 7</td>
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<td>TFT</td>
<td>53</td>
</tr>
<tr>
<td>Khao Phanom Bencha Batholith</td>
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<td>43</td>
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<tr>
<td></td>
<td>Ar-Ar</td>
<td>41</td>
</tr>
<tr>
<td>Phuket area</td>
<td>WR Rb-Sr</td>
<td>108 ± 5</td>
</tr>
</tbody>
</table>

![Figure 5.1: A morphotectonic map of South-East Asia, the two red boxes indicate the study area.](image-url)
5.2 AFT-analysis

5.2.1 Sample preparation

5.2.1.1 Separation of apatite

In short, the apatite minerals were separated from the original whole rock samples, embedded in epoxy and prepared for irradiation. The entire procedure was carried out at Ghent University in the laboratories for Mineralogy and Petrology (MINPET).

First of all, the whole rock samples were crushed, milled and sieved to a desired grain size, namely between 250 µm and 75 µm, by means of a jaw crusher and disc mill.

![Figure 5.2](image-url)

*Figure 5.2: a) The jaw-crusher, used for an initial crushing. By decreasing the distance between the two jaws, a grain size of about 5mm can be reached.; b) a close-up of the discs of the disc mill; c) the disc mill, used for further grinding of the samples. Grinding occurs by means two discs at an adjustable distance from each other until a grain size between 250 and 75µm is acquired.*
A jaw crusher (figure 5.2a) is constructed out of two jaws. A sample is crushed in between these two jaws. Typically, one of the jaws is fixed. By decreasing the distance between the two jaws, an increasingly smaller grain size (in our case up to around 5 mm is achieved).

Further crushing was accomplished by means of a disc mill (figure 5.2b,c). A disc mill consists of two counter-rotating discs in between which samples are crushed. Hence, the spacing of the discs determines the resulting grain size. The samples are then sieved between 250 µm and 75 µm. This crushing and dry sieving was followed by a wet sieving and decantation procedure in order to remove the fine dust. For this step, a sieve with a mesh size of 60 µm was used.

The two following separation steps are based on the physical properties of apatite, namely its magnetic susceptibility and density (ρ).

Seeing as apatite is a diamagnetic mineral, it can be separated from ferro- and paramagnetic minerals by means of a Frantz Isodynamic Magnetic Barrier Separator (Figure 5.3a). Essentially, this separator consists of an electromagnet, a feeding tube, a ramp with a dividing wedge and a power supply. The sample throughput can be regulated by a valve at the bottom of the feeder. Whilst travelling through an electromagnetic field, grains with a stronger remnant magnetism are separated from those with a weaker one and collected in different containers. Generally, the separation was performed in four steps, with an increased electrical current (and thus an increased magnetic field). Starting from 0.1 A over 0.5 A and 1 A going to 1.5 A. Note that this automatic separation was preceded by removal of the strongest magnetic fraction with a small hand-held magnet to prevent a blockage in the magnetic separator. The resulting fraction (> 1.5 A) is composed of diamagnetic minerals such as apatite, zircon, quartz and feldspar. The following heavy liquid (figure 5.3b,c) separation step aims to separate the ‘heavy minerals’ such as apatite (p = 3.1-3.2 g/cm³) from lighter minerals such as quartz (p = 2.65-2.66 g/cm³, depending on the amount of impurities) by means of a heavy liquid. Traditionally, bromoform with density of 2.89 g/cm³ is used as heavy liquid. However, this liquid is quite toxic. Therefore the less toxic “LST FastFloat” with a density of 2.81 g/cm³ was used. (Beside its low toxicity, this is relatively new technique offers several other advantages over the older bromoform-technique: http://www.polytungstate.co.uk/LST_Fastfloat.pdf.)

The principle remains the same, namely a heavy liquid with a density of around 2.81 g/cm³ enables the desired heavy minerals (ρ > 2.81 g/cm³) to sink, whilst the lighter minerals stay afloat (ρ ≤ 2.81 g/cm³).
At this point in the preparation phase, the apatites could be hand-picked: a selection of approximately 100 apatite grains were placed in a grid of ten by ten grains on a sticky tape. The selection was performed by means of a Leica M16 FA stereo zoom microscope. The apatites were picked based on their typical hexagonal symmetry, prismatic habit as well as its translucent colour and the absence of a characteristic lustre.

5.2.1.2 Preparation for irradiation

The apatites were then embedded in an epoxy resin. More specifically a mixture of araldite epoxy (AY103) and a hardening compound (HY956) was used in a five to one ratio (at room temperature). After a 48 hour hardening period, 0.5 cm high cylindrical apatite-epoxy mounts were obtained (figure 5.4a). The samples were subsequently ground and polished. The grinding procedure consisted of two steps. First a rather rough mechanical approach, by means of a DELTA LF 350 vertical axis grinding machine (figure 5.4b). Basically the “grainfree” side of the 0.5 cm high cylindrical epoxy tubes was cut off to about 2 mm thick discs. To expose of the internal grain surfaces, the grain side of the samples was ground by means of three types of FEPA P SIC-grinding paper, with a mesh size varying from #500, #1000 to #2400. The subsequent mechanical polishing step, by means of a Struers DP-U4 polishing machine (figure 5.4c), sought to further polish the grain surfaces. Three different polishing pads, and a corresponding diamond paste, were used in the following order: 6 µm, 3 µm and 1 µm. After polishing, the samples were etched in 2.5 % HNO₃ solution for 70 s at a temperature of 25 °C in order to reveal the spontaneous tracks under a microscope.

After etching, the grain side of the samples, standards and glasses, which will be co-irradiated with the samples (see section 1.5), were covered with an external detector (ED), in this case a Goodfellow clear ruby muscovite, making the samples ready for irradiation.

Figure 5.5 shows the irradiation package, which contains samples, standards and glasses.
5.2.1.3 The BR1 nuclear research reactor

The irradiation with thermal neutrons (section 1.3) occurred at the Belgian Nuclear Research Centre (or SCK-CEN) in Mol, more specifically in the BR1 or Belgian Reactor 1. This reactor is the world’s oldest (non–military) research reactor that is still in operation today. The BR1 is a so-called “natural uranium-graphite-air” type reactor. The term “natural uranium” refers to the use of uranium in natural isotopic ratios as nuclear fuel, which is loaded in 569 channels. The moderator surrounding the nuclear fuel (see section 1.3.2) is constructed of 14,500 graphite (or carbon) blocks. The term “air” refers to the cooling system, which is based on a forced convection of air. Cool air is injected and circulated by means of a fan. During convection the air is heated and eventually this warm air is discharged through a chimney. Furthermore, this reactor is shielded off by a 2 m thick concrete wall.

For more information the reader is referred to the SCK-CEN-website (http://www.sckcen.be/nl/Ons-Onderzoek/Research-facilities/BR1-Belgian-Reactor-1).

The irradiation of our samples occurred in the well thermalized X26 channel, with a thermal/epithermal fluence ratio of 98 ± 3. An irradiation time of 10 hours was chosen in order to obtain a thermal fluence in the of $10^{15}$ neutrons/cm² order of magnitude. The axial fluence gradient, determined by counting FT in ED from U-doped glass dosimeters (IRMM-540), is less than 1.8 %. 
Chapter 5: Samples and methodology

5.2.1.4 The final preparation step after irradiation

After irradiation the now slightly activated samples were left to cool in order for the radiation levels to decay. After safety thresholds were reached, the micas (ED) of the samples, standards and glasses were removed. Subsequently these micas were etched in 40% HF solution at 20 °C for 40 minutes. This step is necessary in order to reveal the induced tracks for optical microscopy.
5.2.2 Apatite fission track dating: counting procedures

Counting was conducted with an Olympus BH2 binocular microscope and followed the same routine. The number of spontaneous or induced surface tracks was counted at a magnification of 1250X. This magnification was achieved by a 10X ocular, a 100X dry objective and a 1.25X drawing tube module. This counting grid consisting of 100 numbered square fields had a total surface area of 80.5 \times 80.5 \mu m^2. Hence, this grid could also be used for surface area measurements, which are necessary to convert the counts into densities (see section 1.3).

Six U-doped glass IRM-540 monitors were counted: FCT-1-M7, FCT2-M7, GL-5, GL-6, GL-7 and GL-8 (figure 5.5). For each of these glass monitors, a total of 100 grids were counted in the ED. As mentioned in section 1.5.1, the track density of the glass monitors or \( r_d \) can be related to the thermal neutron fluence. Hence, these values were used to construct a \( r_d \)-calibration curve as a measurement of the thermal neutron fluence and to take the possible fluence gradients into account (see section 6.3.1.1).

In order to determine the \( \zeta \)-factor, counts on apatite age standards had to be carried out (section 1.5). The following six apatite age standards were used: FCT1, FCT2, DUS 16, DUR 31, DUR 34, DUR 35 (figure 5.5). FCT stands for Fish Canyon Tuff apatites and DUS/DUR stands for Durango apatite. Contrary to the glass monitors, both spontaneous and induced tracks had to be counted. This was achieved by the repositioning procedure, i.e. meticulously placing the ED (track-side down) on the epoxy mount that contained the grains and by shifting the focus between the two (Jonckheere et al., 2003). The spontaneous tracks were counted on the apatite grains, whilst the induced tracks were counted on the ED. For statistical reasons (and whenever possible) approximately a 1000 spontaneous tracks were counted.

Finally, the \( r_d/r_s \) ratio of the samples was determined. The following criteria were met: only grains with (1) with a homogeneous track distribution; (2) randomly oriented tracks were counted.

5.2.3 AFT Thermochronology and modelling

5.2.3.1 Length measurements

In order to reconstruct the thermal history of the sample (chapter 2), the length of 100 horizontal confined tracks (if possible) were measured by means of an KONTRON-MOP-AMO3 image analysing system. This image analysing system consists of a processing unit, a digitizing tablet and a cursor with a built-in LED. By projecting this light onto the microscopic image with the drawing tube attachment, the length of a confined track can be measured by marking the two endpoints of the confined track on the digitizing tablet. By means of a micrometre scale, the resulting values could be calibrated. An incident light source was implemented to recognise the confined tracks. In order to make sure that the measurements were restricted to horizontal confined tracks, the following criteria were used: (1)
the tracks had to be etched out over their entire length; (2) both endpoints had to be in focus at the same time.

5.2.3.2 The HeFTy programme

The t-T paths of the samples were modelled by means of the HeFTy-programme. Every TN-sample had a sufficiently high density of confined tracks and could be modelled. This was not the case for the ST-samples, as only one of these samples (ST-41) had enough confined tracks to enable modelling. HeFTy is freeware package developed by Ketcham, R.A. in cooperation with the Apatite to Zircon inc. This programme can be applied for both forward and inverse modelling of the thermal history. This modelling is based on AFT-age, the AFT-length distribution and mean track length. Other low temperature thermochronometric systems can also be modelled by means of this programme, such as the (U-Th)/He-system and vitrinite reflectance. A major advantage of this programme is that it gives its user the choice between various annealing models. For this thesis, the inverse modelling approach in combination with the Ketcham et al. (2007) annealing model (section 2.1.2.3) was chosen.

The advantage over forward modelling is that unknown t-T paths are reconstructed and hence it is no longer restricted to the input of a thermal history (De Grave, 2003). In general, inverse modelling boils down to making predictions of a system based on measurements and fixed boundary conditions. For thermochronometry, this translates in determining a set of t-T-paths that correspond to measured AFT-data and that are bounded by at least two so-called constraints, a high temperature initial constraint and an low-temperature final constraint (Ketcham et al., 2003; Kethcam 2005). In HeFTy, these constraints are entered as boxes delineating a certain t-T-window. For the initial constraint, results of previous research were used. For example, a K-Ar age for the pluton of sample ST-41 was available (see details in section 5.1, table 5.2). The final constraint is considered to represent the conditions during sampling and is automatically fixed by the programme. If additional information is available, this can also be entered as an intermediate constraint. Since two constraints were insufficient to model the thermal history of the TN-samples, additional constraints were used to test several possible scenarios, such as a heating event around 100 Ma.

Several searching algorithms for reconstructing t-T-paths exist. The HeFTy-programme works with the Monte Carlo algorithm, which generates a large number of t-T-paths (between 10,000 and 100,000 in this thesis) and only retains those with a statistically “acceptable” or “good” fit (Ketcham, 2005; Ketcham et al., 2007). The retained paths can be displayed as individual paths or as path
envelopes. The programme will also display the statistically best-fit path. In general, the path envelopes are used for the thermochronologic interpretation and not a single path. Besides these t-T-paths, the goodness of fit with respect to the AFT-age and AFT-length data is also displayed.

5.3 Chemical analysis (ICP-OES)

Weathered edges and surfaces of the fist-sized samples were removed by a diamond cutting disc. The samples were crushed to the proper grain size by means of the Jaw crusher and disc mill (see section 5.2.1.1) and then the samples were further pulverised by means of an agate mill (figure 5.6). The set-up is quite simple: an agate bowl filled halfway with the sample and with six agate balls, was rotated in a stationary unit, the “Fritsch Pulverisette”. The idea behind the agate mill is that by a fast rotation (400 rotations per minute) of the agate bowl, the six balls will pulverize the sample to the silt fraction. Depending on the sample, a certain duration was chosen. Usually this varied between 90 and 120 minutes.

ICP-OES stand for Inductively Coupled Plasma-Optical Emission Spectroscopy. It is a powerful analytical tool that allows the determination of major, minor and trace elements. For this thesis, only the major elements were analysed.

Emission spectroscopy is based on the principle that, during the atomisation of a sample, electrons in these atoms are excited. When these excited electrons fall back to their ground state, the atoms will emit electromagnetic radiation. The wavelength of this radiation is element-specific and the intensity depends on the concentration of the elements within the sample. Hence, analysing the emission spectrum allows a qualitative and quantitative determination of the elements present (Harris, 2009).
ICP-OES is a type of emission spectroscopy, in which an aerosol is atomised by means of an argon plasma: the plasma was created by ionising high-purity Ar-gas by a spark from a Tesla coil. The sample (which has been transformed into an aerosol) is transported to the plasma by an Ar-gas flow, where it will be vaporised due to the high temperature of the plasma. Through collision with free electrons and ions, the vapour will then be atomised in the plasma (Harris, 2009).

Note that prior to the chemical analysis, LOI or loss on ignition (which is a measurement for the amount of volatiles) was determined by heating or pre-igniting the previously pulverised samples at 850°C in a furnace for 30 minutes. After this pre-ignition, the samples were melted by mixing them with lithium metaborate flux and were fused by heating this mixture to 950 °C for 15 minutes. The result is a glasslike substance. This substance is then dissolved in 70-80 ml of 4% nitric acid (HNO₃) for 4 hours. This solution is further diluted to 100 ml and filtered through a (10% HNO₃) filter. Subsequently, the samples are introduced in a peristaltic pump, which leads this solution to an analytic nebulizer by means of a gas flow (Harris, 2009). In this nebulizer, the sample is transformed into an aerosol. Subsequently, this aerosol is introduced into the plasma in a VARIAN 720-ES apparatus. After atomisation and excitation, the optical emission spectrum can then be analysed on an array detector.

This procedure was performed by the staff of the Laboratory of Soil Sciences under the supervision of professor Eric Van Ranst at Ghent University.
Chapter 6: Results

6.1 Thin section analysis

In total 11 thin sections restricted to the Thai Peninsula were analysed, of which 9 were granites and 2 were quartzites (this distinction is based on mineralogy and texture) (table 5.1). The thin sections were studied by means of an Olympus BH-2 transmitting light microscope, thus opaque minerals could not be identified. The terminology used in the following section is based on the works by Mackenzie et al. (1982); Pichler and Schmitt-Riegraf (1997); Vernon (2004); Passchier and Trouw (2005).

6.1.1 Minerals and textures

Tables 6.1 and 6.2, 6.3 give an overview of the observed minerals and textures. First a restricted overview of the granites is given, followed by an overview of the quartzites.

As can be derived from table 6.1, quartz and feldspars (plagioclase, orthoclase and microcline) were omnipresent in the granite samples. These crystals often formed symplectic (or vermicular) intergrowths (figure 6.1a,b), such as a granophyric texture (quartz and alkali-feldspar figure 6.1a) or a myrmekitic texture (quartz and plagioclase, see figure 6.1b). In most samples, exsolution lamellae of albite in orthoclase or in microcline were observable (figure 6.1c). This texture is known as a microperthitic texture, with the “micro” referring to the microscopic scale of this unmixing. Besides these common minerals, biotite and/or chlorite were often present. They typically occurred as parallel intergrowths (figure 6.1d), although chlorite could also form small needles. Biotite and chlorite were also often intergrown with muscovite.

Although the presence of zircon and apatite was suggested from the picking process (5.2.1), they were not always recognised or present in the thin sections. This is probably due to the accessory nature and the limited dimensions of these minerals. They were most often recognised as inclusions in biotite, in which case zircon was typically surrounded by a pleochroic halo.
As indicated in table 6.2, most granites had a holocrystalline, phaneritic, inequigranular anhedral to subhedral texture (figure 6.2a). In other words, the samples were entirely constructed of macrocrystalline minerals of various grain sizes with poorly developed crystal faces. In most samples large subhedral to euhedral orthoclase phenocrysts surrounded by anhedral smaller grains were observed (figure 6.2c). Some authors refer to this kind of texture as a porphyritic texture (e.g. Vernon, 2004).

In most samples, quartz had a pronounced undulose extinction and sutured boundaries. However, when this undulose extinction was absent (sample ST-40, 44), the quartz grains showed straight boundaries and triple junctions (an interfacial angle of 120° between the quartz grains, see figure 6.2b).
As can be expected, quartz was the dominant mineral in both quartzites. Samples ST-34 and ST-53 are characterised by a granoblastic or interlocking fabric (figure 6.3a,b), which is typical for quartzite. In other words quartz occurred as more or less polygonal equant grains. This texture is sometimes also referred to as a foam texture. In sample ST-53 a (quartz) vein was also recognised (figure 6.3b).
6.1.2 Nomenclature

Table 6.4 shows the nomenclature based on the thin section analysis. In order to identify and name specific rock types, the following point-counting method was used: a right-left and a top-bottom transect through each thin section was studied: every mineral that occurred at a fixed distance along this transect was counted. In total approximately 100 counts per sample were made. This way, the volumetric percentage of the minerals was determined. From this count, the percentages of quartz, plagioclase and alkali-feldspar were calculated and normalised. Subsequently, these percentages were plotted in a QAPF-diagram (figure 6.4), making it clear that the samples are divided over two distinct groups: ST-33, ST-37 are monzogranites, whilst the others are syenogranites.

![QAPF-plot and nomenclature of the ST-samples.](image)

Figure 6.4: QAPF-plot and nomenclature of the ST-samples.
Table 6.1: An overview of the mineralogical composition based in the thin section analysis. "?" means that the identification of the mineral is uncertain. Note that ST-34 and ST-53 were quartzites. Every other sample had been previously classified as a granite.

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<td></td>
<td></td>
</tr>
<tr>
<td>Chlorite</td>
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<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
</tr>
<tr>
<td>Muscovite</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td></td>
<td></td>
</tr>
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<td>V</td>
<td>V</td>
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<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
</tr>
<tr>
<td>Apatite</td>
<td>V</td>
<td>?</td>
<td>V</td>
<td>V(?)</td>
<td>V(?)</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
</tr>
<tr>
<td>Titanite</td>
<td>V</td>
<td>V(?)</td>
<td>V</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Calcite</td>
<td>V</td>
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<td></td>
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<td></td>
<td></td>
<td></td>
<td></td>
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<td></td>
</tr>
<tr>
<td>Opaque minerals</td>
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<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
<td>V</td>
</tr>
</tbody>
</table>
Table 6.2: Overview of the observed textures in thin section of the granites. The following abbreviations were used: Bt = biotite, Chl = chlorite, Fs = the general feldspar-term, Hbl = hornblende, Pl = plagioclase, Ms = muscovite, Or = orthoclase, Qz = quartz, Ser = sericite, Tnt = titanite or sphene, Zr = zircon. A granophyric texture is an intergrowth structure between Qz and K-feldspar, a myrmekitic texture is an intergrowth between Qz and Pl, a microperthitic texture is characterised by the presence of albite exsolution lamellae in Or. The terminology was based on Mackenzie et al. (1982).

<table>
<thead>
<tr>
<th>Texture</th>
<th>Sample</th>
<th>ST-33</th>
<th>ST-36</th>
<th>ST-37</th>
<th>ST-38</th>
<th>ST-40</th>
</tr>
</thead>
<tbody>
<tr>
<td>Crystallinity</td>
<td></td>
<td>Holocrystalline</td>
<td>Holocrystalline</td>
<td>Holocrystalline</td>
<td>Holocrystalline</td>
<td>Holocrystalline</td>
</tr>
<tr>
<td>Granularity, crystal size</td>
<td></td>
<td>Phaneritic</td>
<td>Phaneritic</td>
<td>Phaneritic</td>
<td>Phaneritic</td>
<td>Phaneritic</td>
</tr>
<tr>
<td>Fabric</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Crystal faces</td>
<td></td>
<td>Subhedral to euhedral</td>
<td>Anhedral</td>
<td>Anhedral</td>
<td>Predominantly anhedral, although Or can be subhedral to euhedral</td>
<td>Predominantly anhedral, although Or can be subhedral to euhedral</td>
</tr>
<tr>
<td>Shape of crystal aggregates</td>
<td></td>
<td>Inequigranular-polygon</td>
<td>Seriate-polygon</td>
<td>Seriate-polygon</td>
<td>Inequigranular (to seriate)-polygon</td>
<td>Inequigranular-polygon</td>
</tr>
<tr>
<td>Zoning</td>
<td></td>
<td>Pl: both oscillating and discontinous</td>
<td>/</td>
<td>Pl: oscillating</td>
<td>/</td>
<td>/</td>
</tr>
<tr>
<td>Orientation</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Intergrowth textures</td>
<td></td>
<td>Parallel intergrowth of Bt ad Chl</td>
<td>Granophyric, inclusion growth of Bt and Chl</td>
<td>Granophyric, Parallel intergrowth of Chl and Mus</td>
<td>Parallel intergrowth of Ms and Bt/Chl; Microperthitic</td>
<td>Parallel intergrowth between Ms and Bt; Microperthitic</td>
</tr>
<tr>
<td>Inclusions</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Alteration</td>
<td></td>
<td>Hbl to Chl; Bt to Chl</td>
<td>Bt to Chl; Fs to Ser</td>
<td>Bt to Chl; Fs to Ser (limited)</td>
<td>Pl to clay minerals (probably kaolinite)</td>
<td>Bt to Chl; Pl to clay minerals (probably kaolinite)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
### Table 6.2: Continued

<table>
<thead>
<tr>
<th>Texture</th>
<th>Sample</th>
<th>ST-33</th>
<th>ST-36</th>
<th>ST-37</th>
<th>ST-38</th>
<th>ST-40</th>
</tr>
</thead>
<tbody>
<tr>
<td>Other</td>
<td>Qz: undulose extinction (less pronounced than in the others); Sutured texture; Hbl: single twins; Fs: two fraction, the smaller ones are aligned and less weathered); Idem for Qz; Chl can occur as small needles</td>
<td>Qz: undulose extinction; Sutured texture; Oxidation along fractures</td>
<td>Qz: undulose extinction; Sutured texture; Fs: two fraction, the smaller ones are aligned and less weathered; Idem for Qz</td>
<td>Qz: undulose extinction; Sutured boundaries between Qz-grains; Fs: two fraction, the smaller ones are aligned and less weathered); Idem for Qz</td>
<td>Qz: undulose extinction, those without had straight boundaries and triple junctions; Fs: two fraction, the smaller ones are aligned and less weathered); Idem for Qz; Oxidation along fractures</td>
<td></td>
</tr>
</tbody>
</table>

| Colour of Bt | Pleochroic from yellow to dark brown green | Pleochroism less pronounced from dark brown green to lighter green | Very pale brown to reddish brown | Very pale brown to dark brown | Very pale brown to reddish brown |

<table>
<thead>
<tr>
<th>Texture</th>
<th>Sample</th>
<th>ST-42</th>
<th>ST-43</th>
<th>ST-44</th>
<th>ST-45</th>
<th>ST-54</th>
</tr>
</thead>
<tbody>
<tr>
<td>Crystallinity</td>
<td>Holocrystalline</td>
<td>Holocrystalline</td>
<td>Holocrystalline</td>
<td>Holocrystalline</td>
<td>Holocrystalline</td>
<td>Holocrystalline</td>
</tr>
<tr>
<td>Granularity, crystal size</td>
<td>Phaneritic</td>
<td>Phaneritic</td>
<td>Phaneritic</td>
<td>Phaneritic</td>
<td>Phaneritic</td>
<td></td>
</tr>
<tr>
<td>Fabric</td>
<td>Crystal faces</td>
<td>Predominantly anhedral</td>
<td>Predominantly anhedral</td>
<td>Subhedral</td>
<td>predominantly anhedral</td>
<td>predominantly anhedral, exceptionally Or was Subhedral</td>
</tr>
<tr>
<td>Shape of crystal aggregates</td>
<td>Inequigranular-polygonal</td>
<td>Inequigranular-polygonal</td>
<td>equigranular-polygonal</td>
<td>Inequigranular-polygonal</td>
<td>Seriate-polygonal</td>
<td></td>
</tr>
<tr>
<td>Zoning</td>
<td>/</td>
<td>PI: discontinuous</td>
<td>PI: discontinuous</td>
<td>PI: discontinuous</td>
<td>/</td>
<td></td>
</tr>
<tr>
<td>Mutual relations of crystals</td>
<td>Orientation</td>
<td>Non-preferential</td>
<td>Non-preferential</td>
<td>Non-preferential</td>
<td>Non-preferential</td>
<td>Non-preferential</td>
</tr>
</tbody>
</table>
## Mutual relations of crystals

<table>
<thead>
<tr>
<th>Intergrowth textures</th>
<th>Inclusions</th>
<th>Alteration</th>
<th>Other</th>
<th>Colour of bt</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Granophyric, Myrmekitic; Microperthitic; Parallel intergrowth between Ms and Bt</strong></td>
<td>In bt, Mus: Zr pleochroic haloes</td>
<td>Bt to opaque minerals; Fs to Ser, Fs to clay minerals</td>
<td>Qz: undulose extinction; Fs: two fractions, both have a more or less random orientation; Chl: occur as small needles; Oxidation along fractures</td>
<td>Very pale yellow (almost colourless) to reddish brown</td>
</tr>
<tr>
<td><strong>Granophyric, Myrmekitic; Microperthitic; Parallel intergrowth between Ms and Bt</strong></td>
<td>In Bt: Zr in brownish pleochroic haloes, apatite</td>
<td>Bt to opaque minerals; Fs to Ser, Fs to clay minerals</td>
<td>Qz: undulose extinction; Sutured boundaries between Qz grains; Fs: two fraction, both have a more or less random orientation; Idem voor Qz; Oxidation along fractures</td>
<td>Orange brown to reddish brown</td>
</tr>
<tr>
<td><strong>Granophyric, Myrmekitic; Microperthitic</strong></td>
<td>/</td>
<td>In Bt/Chl: Zr in brownish pleochroic haloes</td>
<td>Or-Carlsbad (or simple twins) were absent; Chl also present as small needles; Oxidation along fractures; Texturally different from the rest: Qz had straight boundaries and triple junctions between the Qz grains, Undulose extinction of quartz was hardly observable</td>
<td>Very pale brown to reddish brown</td>
</tr>
<tr>
<td><strong>Granophyric, Parallel intergrowth of Ms and Chl</strong></td>
<td>In Chl: Zr in brownish pleochroic haloes, apatite; In Qz: Zr</td>
<td>Bt to Chl; Fs to Ser, Fs to clay minerals</td>
<td>Qz: undulose extinction; Sutured boundaries between Qz grains; Fs to Ser, Bt to Chl; Chl: as small needles; Oxidation along fractures</td>
<td>Very pale yellow (almost colourless) to dark brown</td>
</tr>
<tr>
<td><strong>Granophyric, Microperthitic</strong></td>
<td>In Chl: Zr in brownish pleochroic haloes</td>
<td>Chl to opaque minerals; Fs to Ser, Bt to opaque minerals; Fs to clay minerals</td>
<td>Texturally different from the rest, more weathered, dominant grain size was smaller</td>
<td>Hard to tell due to the intergrowth with Chl, but very pale brown to dark brown</td>
</tr>
</tbody>
</table>

*Inclusions:* Inclusions in Bt, Mus: Zr pleochroic haloes in Bt: Zr in brownish pleochroic haloes, apatite / In Bt/Chl: Zr in brownish pleochroic haloes.

*Alteration:* Bt to opaque minerals; Fs to Ser, Fs to clay minerals. Bt to opaque minerals; Fs to Ser, Fs to clay minerals. In Bt/Chl: Zr in pleochoric haloes, Bt to Chl; Fs to Ser, Fs to clay minerals. Chl to opaque minerals; Fs to Ser, Bt to opaque minerals; Fs to clay minerals.

*Other:* Qz: undulose extinction; Fs: two fractions, both have a more or less random orientation; Chl: occur as small needles; Oxidation along fractures. Qz: undulose extinction; Sutured boundaries between Qz grains; Fs: two fraction, both have a more or less random orientation; Idem voor Qz; Oxidation along fractures. Or-Carlsbad (or simple twins) were absent; Chl also present as small needles; Oxidation along fractures; Texturally different from the rest: Qz had straight boundaries and triple junctions between the Qz grains, Undulose extinction of quartz was hardly observable. Qz: undulose extinction; Sutured boundaries between Qz grains; Fs to Ser, Bt to Chl; Chl: as small needles; Oxidation along fractures. Texturally different from the rest, more weathered, dominant grain size was smaller.

*Colour of bt:* Very pale yellow (almost colourless) to reddish brown. Orange brown to reddish brown. Very pale brown to reddish brown. Very pale yellow (almost colourless) to dark brown. Hard to tell due to the intergrowth with Chl, but very pale brown to dark brown.
### Table 6.3: Description of the non-granite samples based on thin section analysis.

<table>
<thead>
<tr>
<th>Texture</th>
<th>Sample</th>
<th>ST-34</th>
<th>ST-53</th>
</tr>
</thead>
<tbody>
<tr>
<td>Description</td>
<td>Quartzitic, granoblastic texture</td>
<td>Larger grain size fraction</td>
<td>Smaller grain size fraction</td>
</tr>
<tr>
<td>ST-34</td>
<td>ST-53</td>
<td>veins</td>
<td>Quartzitic, granoblastic texture of microcrystalline Qz</td>
</tr>
</tbody>
</table>

### Table 6.4: Nomenclature of the samples based on the QAPF-diagram. The percentages are based on approximately 100 counts of the minerals present in the thin sections.

<table>
<thead>
<tr>
<th>Sample</th>
<th>% Qz</th>
<th>% K-feldspar</th>
<th>% Pl</th>
<th>Name</th>
</tr>
</thead>
<tbody>
<tr>
<td>ST-33</td>
<td>0.25</td>
<td>0.44</td>
<td>0.31</td>
<td>Monozogranite</td>
</tr>
<tr>
<td>ST-36</td>
<td>0.42</td>
<td>0.45</td>
<td>0.13</td>
<td>Syenogranite</td>
</tr>
<tr>
<td>ST-37</td>
<td>0.40</td>
<td>0.30</td>
<td>0.30</td>
<td>Monozogranite</td>
</tr>
<tr>
<td>ST-38</td>
<td>0.29</td>
<td>0.49</td>
<td>0.22</td>
<td>Syenogranite</td>
</tr>
<tr>
<td>ST-40</td>
<td>0.28</td>
<td>0.62</td>
<td>0.10</td>
<td>Syenogranite-(Alkali feldspar granite)</td>
</tr>
<tr>
<td>ST-42</td>
<td>0.34</td>
<td>0.51</td>
<td>0.15</td>
<td>Syenogranite</td>
</tr>
<tr>
<td>ST-43</td>
<td>0.35</td>
<td>0.57</td>
<td>0.08</td>
<td>Syenogranite-(Alkali feldspar granite)</td>
</tr>
<tr>
<td>ST-44</td>
<td>0.44</td>
<td>0.47</td>
<td>0.09</td>
<td>Syenogranite</td>
</tr>
<tr>
<td>ST-45</td>
<td>0.45</td>
<td>0.45</td>
<td>0.10</td>
<td>Syenogranite</td>
</tr>
<tr>
<td>ST-54</td>
<td>0.37</td>
<td>0.52</td>
<td>0.11</td>
<td>Syenogranite</td>
</tr>
</tbody>
</table>
## 6.2 Chemical analysis

In table 6.5 the results of the ICP-OES analysis are presented as weight percentages of the elementary oxides of the major elements. Altogether, these results are very similar to the result of those from the surrounding area studied by the MINPET group (e.g. Soens, 2012-2013; Yao, 2012-2013).

Note that this analysis does not differentiate between Fe\(^{3+}\) and Fe\(^{2+}\). The measured Fe\(_2\)O\(_3\)-percentage is in fact a combination of the actual weight percentages of Fe\(_2\)O\(_3^*\) and FeO*\(^*\). Hence these percentages had to be calculated.

<table>
<thead>
<tr>
<th></th>
<th></th>
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<th></th>
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<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>(\text{SiO}_2)</td>
<td>64.96</td>
<td>75.95</td>
<td>69.94</td>
<td>73.49</td>
<td>72.73</td>
<td>73.62</td>
<td>73.11</td>
<td>74.23</td>
<td>72.76</td>
<td>72.33</td>
<td>62.26</td>
</tr>
<tr>
<td>(\text{Al}_2\text{O}_3)</td>
<td>15.97</td>
<td>12.93</td>
<td>13.61</td>
<td>13.82</td>
<td>13.91</td>
<td>14.35</td>
<td>14.19</td>
<td>13.56</td>
<td>14.00</td>
<td>14.55</td>
<td>10.62</td>
</tr>
<tr>
<td>(\text{Fe}_2\text{O}_3)</td>
<td>3.72</td>
<td>0.94</td>
<td>3.56</td>
<td>1.90</td>
<td>2.17</td>
<td>1.51</td>
<td>2.07</td>
<td>1.66</td>
<td>2.13</td>
<td>2.24</td>
<td>15.46</td>
</tr>
<tr>
<td>(\text{FeO}^*)</td>
<td>0.86</td>
<td>0.22</td>
<td>0.82</td>
<td>0.44</td>
<td>0.35</td>
<td>0.48</td>
<td>0.38</td>
<td>0.49</td>
<td>0.52</td>
<td>3.57</td>
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</tr>
<tr>
<td>(\text{MnO})</td>
<td>0.08</td>
<td>0.03</td>
<td>0.06</td>
<td>0.04</td>
<td>0.04</td>
<td>0.05</td>
<td>0.03</td>
<td>0.03</td>
<td>0.02</td>
<td>0.25</td>
<td></td>
</tr>
<tr>
<td>(\text{MgO})</td>
<td>0.94</td>
<td>0.08</td>
<td>0.93</td>
<td>0.24</td>
<td>0.33</td>
<td>0.38</td>
<td>0.48</td>
<td>0.26</td>
<td>0.35</td>
<td>0.66</td>
<td>2.21</td>
</tr>
<tr>
<td>(\text{CaO})</td>
<td>3.54</td>
<td>0.29</td>
<td>1.96</td>
<td>0.75</td>
<td>0.47</td>
<td>0.93</td>
<td>1.07</td>
<td>1.17</td>
<td>0.29</td>
<td>0.17</td>
<td>4.49</td>
</tr>
<tr>
<td>(\text{Na}_2\text{O})</td>
<td>2.86</td>
<td>3.21</td>
<td>2.43</td>
<td>2.73</td>
<td>2.18</td>
<td>2.88</td>
<td>2.50</td>
<td>2.65</td>
<td>2.48</td>
<td>2.21</td>
<td>2.43</td>
</tr>
<tr>
<td>(\text{K}_2\text{O})</td>
<td>5.87</td>
<td>5.25</td>
<td>5.36</td>
<td>5.57</td>
<td>6.32</td>
<td>5.13</td>
<td>5.82</td>
<td>5.97</td>
<td>5.95</td>
<td>5.88</td>
<td>0.47</td>
</tr>
<tr>
<td>(\text{TiO}_2)</td>
<td>0.42</td>
<td>0.06</td>
<td>0.58</td>
<td>0.20</td>
<td>0.29</td>
<td>0.23</td>
<td>0.32</td>
<td>0.15</td>
<td>0.29</td>
<td>0.31</td>
<td>1.27</td>
</tr>
<tr>
<td>(\text{P}_2\text{O}_5)</td>
<td>0.18</td>
<td>&lt;</td>
<td>0.14</td>
<td>0.07</td>
<td>0.11</td>
<td>0.09</td>
<td>0.10</td>
<td>0.13</td>
<td>0.09</td>
<td>0.15</td>
<td></td>
</tr>
<tr>
<td>(\text{LOI})</td>
<td>0.90</td>
<td>0.62</td>
<td>0.70</td>
<td>0.84</td>
<td>1.27</td>
<td>0.84</td>
<td>0.79</td>
<td>0.46</td>
<td>1.42</td>
<td>1.99</td>
<td>0.14</td>
</tr>
<tr>
<td>Total</td>
<td>99.45</td>
<td>99.36</td>
<td>99.25</td>
<td>99.66</td>
<td>99.74</td>
<td>100.03</td>
<td>100.48</td>
<td>100.24</td>
<td>99.82</td>
<td>100.45</td>
<td>99.74</td>
</tr>
</tbody>
</table>

## 6.3 AFT-analysis

### 6.3.1 Calibration

#### 6.3.1.1 The glass monitor interpolation curve

As mentioned in section 1.5 and 5.2.2, the (axial gradient of the) thermal neutron fluence can be determined from the induced track density in co-irradiated glass monitors. As mentioned in the previous chapter, the number of induced tracks \(N_d\) and the track density \(\rho_d\) were determined in an ED attached to the glass monitors during irradiation. In this thesis, six U-doped IRMM-540 glass monitors were used (figure 5.5, table 6.6). IRMM-540 glass monitors are reference glasses with a fixed U-content of 13.9 ± 0.5 ppm. Note that two of these glass dosimeter had been embedded together with a Fish Cayon Tuff apatite age standard (FCT).

As these glass monitors were positioned at regular intervals within the irradiation package, the axial gradient of the thermal neutron fluence could be determined. Figure 6.5 shows the interpolation
curve that was obtained by plotting the \( \rho_d \)-values in function of the relative distance of the glass monitors. The slope of this curve is a measure of the axial thermal neutron fluence. By means of linear regression, an equation can be calculate that expresses the \( \rho_d \)-value as a function of relative position of the glass dosimeters within the irradiation package. Hence, this curve can also be used to calculate the \( \rho_d \) of any sample or standard within the same irradiation package. These values are necessary in order to determine the \( \zeta \)-calibration factors and \( \zeta \)-ages (see equation 1.13 and 1.14).

Table 6.6: The results of the counts of the induced tracks in ED irradiated against an IRMM-540 glass monitors that were co-irradiated with the samples and standards. FCT refers to glass dosimeter and an age standards, GL refers to a separate glass dosimeter.

<table>
<thead>
<tr>
<th>Glass monitor</th>
<th>Position within radiation package (^a)</th>
<th>( N_d ) (^b)</th>
<th>( \rho_d (\pm 1\sigma) ) (^c)</th>
</tr>
</thead>
<tbody>
<tr>
<td>FCT2-M7</td>
<td>0.00</td>
<td>1983</td>
<td>3.162 ± 0.071</td>
</tr>
<tr>
<td>GL-7</td>
<td>16.4</td>
<td>2045</td>
<td>3.249 ± 0.072</td>
</tr>
<tr>
<td>GL-8</td>
<td>17.37</td>
<td>1908</td>
<td>3.030 ± 0.069</td>
</tr>
<tr>
<td>GL-6</td>
<td>30.85</td>
<td>2020</td>
<td>3.218 ± 0.072</td>
</tr>
<tr>
<td>GL-5</td>
<td>43.25</td>
<td>1955</td>
<td>3.123 ± 0.071</td>
</tr>
<tr>
<td>FCT1-M7</td>
<td>51.12</td>
<td>1937</td>
<td>3.079 ± 0.070</td>
</tr>
</tbody>
</table>

\(^a\) the relative position the glasses dosimeters expressed in mm, calculated from an arbitrary starting point, in this case FCT2-M7.

\(^b\) is the number of induced tracks counted in an ED irradiated against a IRMM-540 glass monitor.

\(^c\) is the density of the induced tracks (with standard error) expressed in tracks in an ED irradiated against a IRMM-540 glass monitor, expressed as \( 10^5 \) tracks/cm.

Figure 6.5: The interpolation curve obtained from plotting the induced track density of the glass monitors against the position of these glass monitors in the irradiation package. The FCT2-M7 glass monitor was arbitrarily chosen as distance zero. The slope of this curve is measure of the axial gradient of the thermal neutron fluence. The \( R^2 \) value indicated in the figure represents the multiple determination coefficient.

6.3.1.2 \( \zeta \)-Calibration

The theoretical concept of the \( \zeta \)-calibration method has already been outlined in section 1.5. In summary, the formula of this user-specific empirical calibration factor is given by the following equation (equation 1.13):
\[
\zeta = \frac{\rho_{s}/\rho_i}{\lambda_d(\rho_d/\rho_i)} \sigma(\rho_d)
\]

With fixed values:

\[\lambda_d = 1.55125 \times 10^{-10} \text{ a}^{-1} \text{ (Jaffey et al., 1971; Steiger and Jäger, 1997).}\]
\[G = 0.5 \text{ (ED-method)}\]
\[t_s = 27.9 \pm 0.5 \text{ (Hurford and Hammerschmidt, 1985) Ma for the Fish Canyon Tuff (FCT) standards and 31.4 \pm 0.5 \text{ Ma (McDowell and Keizer, 1997) for the Durango apatite age standards (DUR/DUS).}\]

This leaves \((\rho_s/\rho_i)\) and \(\rho_d\) to be determined. The \(\rho_s\) and \(\rho_i\)-values were derived from the number of tracks counted in respectively the age standards and the associated ED. In total six mounts of FCT and DUR/DUS apatite age standard were analysed in this work. The \(\rho_d\)-values were determined in one of two ways. For the mounts containing age standards embedded together with the glass monitors (FTC2-M7 and FCT1-M7), the already and directly determined \(\rho_d\)-values were used. For the others, \(\rho_d\) was calculated from the previously determined regression equation. Hence, this is an interpolated value.

Table 6.7 lists the resulting \(\zeta\)-factors of the apatite age standard (mounts).

<table>
<thead>
<tr>
<th>Standard</th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>FCT2-M7</td>
<td>100</td>
<td>843</td>
<td>1.482</td>
<td>(0.051)</td>
<td>1332</td>
<td>2.312</td>
<td>0.737</td>
<td>1995</td>
</tr>
<tr>
<td>FCT1-M7</td>
<td>107</td>
<td>1144</td>
<td>1.962</td>
<td>(0.058)</td>
<td>1348</td>
<td>2.555</td>
<td>0.952</td>
<td>1956</td>
</tr>
<tr>
<td>DUS-16</td>
<td>100</td>
<td>901</td>
<td>1.390</td>
<td>(0.046)</td>
<td>1663</td>
<td>2.566</td>
<td>0.570</td>
<td>1983</td>
</tr>
<tr>
<td>DUR 31</td>
<td>100</td>
<td>950</td>
<td>1.474</td>
<td>(0.048)</td>
<td>1524</td>
<td>2.367</td>
<td>0.652</td>
<td>1961</td>
</tr>
<tr>
<td>DUR 34</td>
<td>94</td>
<td>816</td>
<td>1.381</td>
<td>(0.048)</td>
<td>1434</td>
<td>2.412</td>
<td>0.598</td>
<td>1963</td>
</tr>
<tr>
<td>DUR 35</td>
<td>100</td>
<td>871</td>
<td>1.374</td>
<td>(0.047)</td>
<td>1542</td>
<td>2.445</td>
<td>0.587</td>
<td>1973</td>
</tr>
</tbody>
</table>

a The number of analysed grains.
b \(N_{N,s,i,d}\) are the number of counted spontaneous, induced tracks and induced tracks in the ED irradiated against an IRMM-540 glass monitor. Mind that \(N_d\) is an interpolated value.
c \(\rho_{s,i,d}\) are the densities of respectively the spontaneous (s), induced (i) and tracks and the tracks in an ED irradiated against an IRMM-540 glass monitors. Mind that \(\rho_d\) is an interpolated value. \(\rho_{s,i,d}\) are expressed as \(10^5\) tracks/cm², with the standard error between brackets.
d the \(\zeta\)-factor expressed in \(\text{a.cm}^2\).
e the standard error on the \(\zeta\)-factor.

In Table 6.7 the standard error or statistical uncertainty (\(\sigma(\zeta)\)) on the zeta-factor is also indicated. This uncertainty was calculated according to equation 6.2. This approach is known as the conventional model for uncertainty calculation (Green, 1981) and it assumes that a stochastic fission process can be entirely described by means of Poissonian statistics.
\[
\sigma_t = \sqrt{\left(\frac{\sigma_t}{t_s}\right)^2 + \frac{1}{N_s} + \frac{1}{N_i} + \frac{1}{N_d}}
\]

With:

- \(t_s\) = the reference age of the age standards (see above)
- \(\sigma_t\) = the error of the reference age
- \(N_s\) = the number of spontaneous tracks, counted in the age standard
- \(N_i\) = the number of induced tracks, counted in the ED irradiated against the age standard
- \(N_d\) = the number of induced tracks in the ED irradiated against an IRMM-540 glass monitor.

Similar to \(\rho_d\), this is an interpolated value calculated from the regression equation.

As mentioned in section 1.5.1, it is preferred to work with a SWMZ (sample weighted mean zeta) and an OWMZ (overall weighted mean zeta) instead of the individual \(\zeta\)-factors listed in table 6.7 (Hurford and Green, 1983). The SWMZ values have to be calculated separately for the FCT and DUR/DUS age standards, based on the data presented in table 6.7. Subsequently, the OWMZ can be calculated from these two SWMZ-values. The idea behind these weighted zeta’s is to give the individual \(\zeta\)-factors with a smaller statistical error a larger weight when calculating the mean value in order to enhance the reproducibility. Therefore, the reciprocal value of \(\sigma_t^2\) is used a weighing factor in the calculation of SWMZ:

\[
\text{SWMZ} = \frac{\sum \frac{1}{\sigma_j^2}}{\sum \frac{1}{\sigma_j}}
\]

The SWMZ has a weighted uncertainty, given by:

\[
\sigma_{\text{SWMZ}} = \sqrt{\frac{1}{\sum \frac{1}{\sigma_j}}}
\]

Inserting the data from table 6.7, yields the following SMWZ-values (expressed in a.cm²):

- \(\text{SWMZ}_{\text{DUR/DUS}} = 330.09 \pm 9.62 \text{ a.cm}^2\)
- \(\text{SWMZ}_{\text{FCT}} = 208.44 \pm 7.56 \text{ a.cm}^2\)

OWMZ can be calculated from these SWMZ-values in a similar manner:
The weighted uncertainty of the OWMZ is given by:

\[
\text{OWMZ} = \frac{\xi_{\text{UR}}, \xi_{\text{FCT}}}{\sqrt{\frac{1}{\sigma_{\text{UR}}^2} + \frac{1}{\sigma_{\text{FCT}}^2}}}
\]  

(6.5)

Based on the SWMZ values and their weighted uncertainties, the following calibration factor (also expressed in a.cm²) could be calculated:

\[
\text{OWMZ} = 254.87 \pm 5.94
\]

The obtained value corresponds well to those obtained by other researchers in the MINPET-group for a similar microscope set-up, similar age standards and IRM-540 glass monitors: 253.1 ± 2.4 (Prof. Dr. Johan De Grave) and 259.1 ± 3.3 (Dr. Stijn Glorie). This OWMZ calibration factor will be used to calculate the AFT ζ-ages. It is important to bear in mind that this is a user-specific value, which strongly depends on laboratory and observation conditions. It also depends on the type of glass dosimeter (in this case the IRM-540) and the type of material that were used, in this case apatite (De Grave, 2003).

6.3.1.3: Calibration of the length measurements

As discussed in section 2.2, the length distributions of confined tracks are an important tool for reconstructing the geochronological history of a sample. There is a slight variation in length measurements between individual researchers. Hence, the confined length distributions of two reference samples (Hurford 2 and Hurford 3) were measured here as a measure of internal control and reference, with Hurford 2 being an induced AFT-distribution. Their length distributions are depicted in figure 6.6. Despite a limited tendency towards lower values, the results were in close comparison with the values found in the literature and those obtained from the researchers in the MIPNPET research group: for Hurford 2 the values of 16.38 ± 0.08 µm (Prof. Dr. Johan De Grave) and 16.23 ± 0.09 µm (Dr. Stijn Glorie) were found and for Hurford 3, the values of 10.45 ± 0.13 µm (Prof. Dr. Johan De Grave) and 10.69 ± 0.15 µm (Dr. Stijn Glorie) were obtained.
6.3.2 Results of the AFT-analysis on the TN and ST-samples

6.3.2.1 The calculated ζ-ages of the TN and ST-samples

The results of the counting procedures as outlined in section 5.2.2 are shown in table 6.8. Based on these results and the previously calculated OWMZ, the ζ-age ($t(\zeta)$) (table 6.8) was calculated according to equation 1.14. Similar to the uncertainty of the individual ζ-factors (see above), this uncertainty was also calculated by means of the conventional model for uncertainty calculation (Green, 1981) (see above), given by:

$$RE(t) = \sqrt{\left(\frac{\sigma(t)}{\zeta}\right)^2 + \frac{1}{N_s} + \frac{1}{N_i} + \frac{1}{N_d}}$$

6.7

$$total~uncertainty = t \cdot RE(t)$$

6.8

With:

- $RE = \text{the relative error}$
- $\zeta = \text{OWMZ of 254.87 (a.cm}^2\text{)}$
- $N_{s,i,d} = \text{the number of spontaneous, induced tracks and induced tracks in the ED irradiated against a IRMM-540 glass monitor. } N_d \text{ is an interpolated value (see above).}$

Overall, the ζ-ages of both the TN and ST-samples are very consistent. The TN-samples mainly yielded Early Jurassic to Early Cretaceous age, with the exception of sample TN-36. Therefore this is considered as an outlier. The ST-samples all yielded an Eocene to Miocene ages.
## Table 6.8: Apatite fission track results (AFT) for the TN and ST-samples.

<table>
<thead>
<tr>
<th>Sample</th>
<th>n</th>
<th>$\rho_s (\pm \sigma)^\text{h}$</th>
<th>$N_\text{s}$</th>
<th>$\rho_i (\pm \sigma)^\text{h}$</th>
<th>$N_\text{i}$</th>
<th>$\rho_d (\pm \sigma)^\text{h}$</th>
<th>$N_\text{d}$</th>
<th>$\rho_s / \rho_i$</th>
<th>$P (\chi^2)^\text{df}$</th>
<th>$t(\Omega)^\text{f}$</th>
<th>$l_m$</th>
<th>$n_l$</th>
<th>$\sigma$</th>
</tr>
</thead>
<tbody>
<tr>
<td>TN-35</td>
<td>20</td>
<td>9.937 (0.284)</td>
<td>1228</td>
<td>3.760 (0.174)</td>
<td>465</td>
<td>3.166 (0.071)</td>
<td>1989</td>
<td>2.797</td>
<td>0.89</td>
<td>111.9 ± 7.1</td>
<td>12.0</td>
<td>102</td>
<td>1.6</td>
</tr>
<tr>
<td>TN-36</td>
<td>20</td>
<td>4.463 (0.199)</td>
<td>502</td>
<td>0.799 (0.083)</td>
<td>93</td>
<td>3.168 (0.071)</td>
<td>1991</td>
<td>6.593</td>
<td>0.79</td>
<td>260.9 ± 30.6</td>
<td>11.4</td>
<td>98</td>
<td>1.7</td>
</tr>
<tr>
<td>TN-37</td>
<td>20</td>
<td>20.420 (0.423)</td>
<td>2335</td>
<td>4.526 (1.990)</td>
<td>517</td>
<td>3.170 (0.071)</td>
<td>1992</td>
<td>4.683</td>
<td>0.81</td>
<td>186.5 ± 8.5</td>
<td>12.0</td>
<td>100</td>
<td>1.3</td>
</tr>
<tr>
<td>TN-38</td>
<td>20</td>
<td>11.111 (0.360)</td>
<td>1315</td>
<td>2.718 (1.517)</td>
<td>321</td>
<td>3.171 (0.071)</td>
<td>1993</td>
<td>4.132</td>
<td>0.95</td>
<td>164.9 ± 11.6</td>
<td>12.0</td>
<td>100</td>
<td>1.6</td>
</tr>
<tr>
<td>ST-33</td>
<td>46</td>
<td>2.842 (0.139)</td>
<td>419</td>
<td>4.106 (0.152)</td>
<td>734</td>
<td>3.127 (0.071)</td>
<td>1964</td>
<td>0.758</td>
<td>0.13</td>
<td>30.1 ± 2.1</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-35</td>
<td>8</td>
<td>3.761 (0.495)</td>
<td>70</td>
<td>6.842 (0.572)</td>
<td>143</td>
<td>3.129 (0.071)</td>
<td>1965</td>
<td>0.552</td>
<td>0.10</td>
<td>22.0 ± 3.3</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-37</td>
<td>71</td>
<td>1.600 (0.727)</td>
<td>484</td>
<td>4.054 (0.115)</td>
<td>1248</td>
<td>3.131 (0.071)</td>
<td>1966</td>
<td>0.422</td>
<td>0.66</td>
<td>16.8 ± 1.1</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-38</td>
<td>24</td>
<td>3.338 (0.176)</td>
<td>359</td>
<td>8.962 (0.295)</td>
<td>923</td>
<td>3.132 (0.071)</td>
<td>1968</td>
<td>0.405</td>
<td>0.10</td>
<td>16.2 ± 1.1</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-39</td>
<td>34</td>
<td>2.954 (0.152)</td>
<td>378</td>
<td>7.987 (0.249)</td>
<td>1027</td>
<td>3.135 (0.071)</td>
<td>1969</td>
<td>0.443</td>
<td>0.28</td>
<td>17.7 ± 1.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-40</td>
<td>4</td>
<td>1.334 (0.236)</td>
<td>32</td>
<td>2.562 (0.334)</td>
<td>59</td>
<td>3.137 (0.071)</td>
<td>1970</td>
<td>0.560</td>
<td>0.04</td>
<td>22.4 ± 5.0</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-41</td>
<td>36</td>
<td>2.061 (0.104)</td>
<td>393</td>
<td>5.567 (0.175)</td>
<td>1010</td>
<td>3.143 (0.071)</td>
<td>1974</td>
<td>0.404</td>
<td>0.58</td>
<td>16.2 ± 1.1</td>
<td>12.4</td>
<td>24</td>
<td>1.9</td>
</tr>
<tr>
<td>ST-42</td>
<td>19</td>
<td>1.282 (0.114)</td>
<td>126</td>
<td>3.185 (0.180)</td>
<td>315</td>
<td>3.145 (0.071)</td>
<td>1976</td>
<td>0.471</td>
<td>0.14</td>
<td>18.9 ± 2.1</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-43</td>
<td>43</td>
<td>3.596 (0.146)</td>
<td>608</td>
<td>7.227 (0.205)</td>
<td>1263</td>
<td>3.146 (0.071)</td>
<td>1977</td>
<td>0.506</td>
<td>0.48</td>
<td>20.3 ± 1.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-44</td>
<td>56</td>
<td>2.446 (0.923)</td>
<td>703</td>
<td>5.813 (0.143)</td>
<td>1653</td>
<td>3.148 (0.071)</td>
<td>1978</td>
<td>0.452</td>
<td>0.14</td>
<td>18.1 ± 1.0</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-45</td>
<td>16</td>
<td>1.268 (0.121)</td>
<td>110</td>
<td>1.585 (0.139)</td>
<td>130</td>
<td>3.151 (0.071)</td>
<td>1979</td>
<td>1.221</td>
<td>0.26</td>
<td>48.8 ± 6.5</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>ST-54</td>
<td>11</td>
<td>2.308 (0.224)</td>
<td>106</td>
<td>4.365 (0.292)</td>
<td>224</td>
<td>3.162 (0.071)</td>
<td>1986</td>
<td>0.621</td>
<td>0.40</td>
<td>28.3 ± 3.0</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

1 The number of analysed grains.

2 $\rho_s$, $\rho_i$, and $\rho_d$ are the densities of respectively the spontaneous (s), induced (i) and tracks and the tracks in an ED irradiated against a IRMM-540 glass monitor (with an U-content of 19.5 ± 0.5 ppm), mind that $\rho_d$ is an interpolated value. $\rho_s$, $\rho_i$, and $\rho_d$ are expressed as $10^3$ tracks/cm², with the standard error between brackets.

3 $N_{i,d}$ are the number of counted spontaneous, induced tracks and induced tracks in the ED ($N_s$ is an interpolated value).

4 $P (\chi^2)$ is the chi-squared probability that the dated grains have a constant $\rho_s / \rho_i$-ratio (the count is only acceptable if this probability is higher than 0.05)

5 $t(\Omega)$ is the AFT-age, expressed in Ma.

6 AFT-length data, with $l_m$, the mean track length, $\sigma$ the standard deviation (in µm) and $n_l$ the number of natural, horizontal confined track.
6.3.3 Length distribution of the TN and ST-samples

Table 6.8 and figure 6.7 represent the results of the confined track measurements, following the procedure mentioned in section 5.2.3.1. In every of the older TN-samples a sufficient number of confined tracks could be measured. As can be seen in figure 6.7, these distributions are quite consistent with a mean track length between 11.4-12.0 µm and a σ between 1.3-1.7 µm. The four distributions are also slightly positively skewed.

For the ST-samples, on the other hand, only sample ST-41 had a “sufficient” number of confined tracks. (This is due to the young age, which in general results in low track densities and very low confined track densities in particular). Even this number of counted confined tracks (24) is barely statistically acceptable. The resulting length distribution exhibits a mean track length of 12.4 µm with a standard error of 1.9 µm.

6.3.4 AFT thermochronology and thermal history modelling

As previously mentioned, the ζ-age, the mean track length and the length distribution can be used to model t-T paths. In this thesis, the HeFTy modelling programme in combination with the Ketcham et al. (2007) annealing equations was used (see sections 2.1.2.3 and 5.2.3.2). The path envelopes of the resulting t-T paths are displayed in figure 6.8. The purple envelope contains the statistically ”goof fit” paths and the green envelope is the envelope of the paths with a statistically “acceptable fit”.

The t-T-path of sample ST-41 (figure 6.8) shows a three-stage cooling throughout the Cenozoic: (1) a first major (rapid) cooling phase which lasted until approximately 29-26 Ma; (2) from 26 Ma onwards this cooling decelerated and continued at a steady pace until around 13-12 Ma when there was a sharp increase in cooling rate which marked a third (3) and final phase encompassing renewed rapid cooling which lasted until the present.

Despite individual differences, the t-T paths of the TN-samples (figure 6.8) show the same general trend: (1) a major slow cooling throughout most of the Palaeozoic and Mesozoic (until about 200-120 Ma); followed by (2) a thermally more stable phase during the rest of the Mesozoic and a part of the Palaeogene (up to 45-20 Ma) and (3) an accelerated cooling phase that continued up to the present. Note that the difference between TN-35, 38 on the one hand and TN-36, 37 on the other could be, in part, due to the extra constraint needed to model the t-T-path of the latter two samples.
Figure 6.7: Confined track length distributions for the apatite samples from India (TN) and Thailand (ST). The total number of measured tracks (n), the mean track length (\( l_m \)) and the standard deviation (\( \sigma \)) of the distribution are also shown.
Figure 6.8: The modelled t-T-paths of samples ST-41 obtained by using the HeFTy-modelling programme and the Ketcham et al. (2007) annealing equation. Both the “good-fit” envelope (purple) and “acceptable fit” (green) are shown.
Figure 6.8 continued: The modelled thermal history of sample TN-35 and TN-36, TN-37, TN-38. Note that the slightly deviating shape of the t-T-path of sample TN-36 is probably due to the extra constraint, which wasn’t used on TN-35 and TN-36.
Chapter 7: Interpretation and discussion

In the first part of this chapter, the results of the thin section and chemical analysis will be discussed, compared to previously published findings and related to their geodynamic context. The second part will focus on the results of the AFT-analysis.

7.1. Minerals and textures

7.1.1. Textures

7.1.1.1 General characteristics

Altogether, the obtained results were in-line with the published results of Cobbing et al. (1986); Cobbing (2011) (see also table 4.1) and the unpublished results of Soens (2012-2013) and Yao (2012-2013). For example, Cobbing (2011) also classified the granitoids from southern Thailand as granodiorites, monzogranites or syenogranites. This author also described a predominantly porphyritic texture with megacrysts of pale K-feldspars. Soens (2011) and Yao (2011) described a porphyritic texture, instead they only mentioned a seriate texture with alkali-feldspar megacrysts. This difference could be due to an inter-batholith difference (see below).

As previously mentioned, all granites had a holocrystalline, phaneritic texture. This is a typical texture for granites, as these plutonic rocks typically crystallise from a slowly cooling melt, giving the crystals time to grow. Moreover, in rocks with a holocrystalline texture, the development of crystal faces is generally hindered due to lack of open space in which crystals can grow. This explains the predominance of subhedral to anhedral crystals (section 6.1.1). Besides this overall texture, orthoclase megacrysts were often observed (figure 6.2c). According to Vernon (2004), the presence of the megacrysts can be explained by means of a low N/G-ratio. This N/G-ratio is the ratio of the nucleation rate (N) over the growth rate (G). This ratio strongly depends on the temperature of the melt (see below) and has a controlling effect on crystal growth and hence on crystal size (figure 7.1). A low N/G-ratio implies that crystal growth will dominate over nucleation. As a result, a limited number of large euhedral crystals can develop, in our case these are the orthoclase megacrysts. Although the reason for this low N/G-ratio is still unclear, Vernon (2004) suggests that the low nucleation rate could be related to the local abundance of water, since the presence of water could depolymerise the Si-O and Al-O bonds in potential nuclei. A rather popular alternative suggests that these megacrysts simply follow the Oswald ripening rule, which states that small crystals will recrystallise into larger ones in an attempt to reduce the surface free energy. However, experiments
have shown that this is an unlikely mechanism for megacrysts or even for crystals observable under an optical microscope (Vernon, 2004 and references herein).

So to conclude, the general texture of these granites suggests an emplacement at relative deep crustal levels. This also explains the presence of microcline, which is the low temperature polymorph of alkali-feldspars and hence can only form during slow cooling. This slow cooling also explains the microperthitic exsolution or unmixing, which is characterised by the occurrence of albite lamellae in orthoclase. At high temperatures, albite and orthoclase are completely mixable and form a solid solution. If cooling is slow enough, this mixability will no longer be complete once the temperature falls below the solvus temperature\(^6\). As a result, the pre-existing homogenous feldspar minerals will break down into two components: an albite-rich (the lamellae) and an orthoclase (or microcline)-rich component (Vernon, 2004).

![Figure 7.1: Growth and nucleation rate in function of temperature. At high temperature (\(T_a\)), crystal growth dominates (low N/G-ratio), resulting in a few euhedral crystals. At low temperature (\(T_b\)), nucleation dominated (high N/G-ratio), leading to a large number of small euhedral crystals (Vernon, 2004).](image)

Based on the observed overgrowth relations, the granites crystallised according to the Bowen’s reaction series. This reaction series explains how felsic magmas (and hence granites) can form as a differentiation product from a primary mafic melt by means of e.g. fractional crystallisation (figure 7.2). This observation corresponds to the previous findings of e.g. Cobbing (2011). According to this author, the progressive growth of space-filling crystals started with mafic minerals such as biotite and

\(^6\) The solvus temperature is the temperature at which the solvus is reached. The solvus is a line or surface in a (respectively binary or ternary phase diagram) that separates a field of homogeneous solid solution from a field of unmixing in two or more phases (Bates and Jackson, 1984).
hornblende, followed by the crystallisation of feldspars. These feldspars form an open framework in which quartz can grow. They also follow the Bowen reaction series, as the crystallisation of Ca-Na-plagioclase preceded that of the K-feldspars (Cobbing 2011 and references herein). Note that sometimes the relation between the minerals could be misleading, e.g. the large orthoclase minerals sometimes appeared to be overgrown by plagioclase (figure 6.2c), which would be in disagreement with the Bowen reaction series. On closer examination, these “inclusions” of plagioclase were oriented parallel to the boundaries of the orthoclase mineral. Therefore, the most likely explanation is that the large orthoclase minerals simply absorbed the pre-existing plagioclase. During further growth of the orthoclase, these pre-existing crystals were oriented according to the crystal lattice of orthoclase.

Cobbing (2011) stressed that although the aforementioned porphyritic texture was generally observed, the different plutons tend to have slight textural differences, making it able to distinguish them from one another. By comparing table 6.1 and 6.2 with figure 4.11b these differences become clear. The difference is especially striking when comparing the mineralogy of sample ST-33 (from the Khao Phanom Bencha Batholith) with the others (from the Phuket-Takua Pa Batholith). The differences between the individual plutons in the Phuket-Tauka Pa Batholith are less pronounced. For example, ST-36, 37, 38 and 54 from the Ban Lam Ru Batholith differ from the others as they had a seriate texture instead of an inequigranular porphyritic texture. A seriate texture is characterised by a continuous variation in grain size (figure 6.2d). This texture can also be explained based on the N/G-ratio (Vernon, 2004), more specifically based on its dependence on temperature. Due to density
contrasts with the surrounding rock, felsic melts (from which the granite crystallises) will migrate upwards. During upward migration, the melt will cool down allowing crystals to nucleate. For example, at temperature $T_a$ (figure 7.1), the growth rate strongly exceeds the nucleation rate, resulting in a limited number of large crystals. With continued cooling (figure 7.1), the nucleation rate will eventually exceed the growth rate (at temperature $t_b$), which causes the formation of a large number of small crystals. This reasoning could explain a continuous varying grain size if the cooling and upward migration occurred gradually, and is yet again an indication for an emplacement at relative deep crustal levels.

The texture of sample ST-44 is another example of this inter-pluton difference. ST-44 was the only sample from the Ban Khata Pluton and it was also the only sample with a more or less equigranular texture. It was also one of the few samples in which quartz boundaries were straight and triple junctions between neighbouring quartz grains were present. (Note that this distinction is based on only one sample and therefore this may not be representative for the entire pluton). Due the proximity of this pluton to the KMF zone (e.g. figure 7.5), its texture is probably related to fault activity along the KMF zone and will be discussed in the following section.

Samples ST-34 and ST-53 were quartzites, so they had a sedimentary instead of igneous protolith. A quartzite is a granoblastic metamorphic rock, mainly consisting of quartz, with a sandstone protolith. They are considered to have formed through recrystallization by regional or thermal metamorphism (Bates and Jackson, 1984). According to Cobbing (2011) (and as indicated in table 4.1), these sandstones are a part of the Kaeng Krachan Group. In the literature (Fujikawa et al., 2005; Ridd 2009a,b) this group is described as sedimentary in origin without having been affected by a major metamorphic event. Thus, the metamorphism is considered to be local and related to the thermal effects of granite emplacement. This notion could be confirmed by a more extensive sampling of this host rock. This could also explain the quartz vein crosscutting the quartzite (sample ST-53), as hydrothermal fluids can dissolve and precipitate quartz. Fault activity could also result in vein formation, however seeing as this sample is not located in the immediate proximity of a fault zone, this is a rather unlikely scenario. After metamorphism, static recrystallization (see next section) probably created the typical granoblastic texture of the quartzites.

### 7.1.1.2 Possible indications of deformation

The most obvious indications of deformation in the thin sections are the undulose extinction of quartz grains and sutured quartz boundaries. Undulose (or undulatory) extinction is considered to be caused by a bending of the grains and can already occur at very limited temperatures (<300 °C) or pressures (<4-5 kbar) (Vernon, 2004; Passchier and Trouw, 2005). As this was observed in most
samples, and not just the samples located close to a mapped major fault zone (e.g. sample ST-37), this is probably related to burial and not (solely) to fault activity. Burial can explain the presence of sutured contacts through grain boundary migrations (Passchier and Trouw, 2005) and is consistent with the notion the granites were emplaced at relatively deep crustal levels. However, as indicated on the geological map, most samples were located close to small undifferentiated faults (figure 7.4), so a possible contribution of this fault activity to deformation cannot be excluded. Both scenarios fail to explain why the micas were not aligned.

As previously mentioned, sample ST-44 was characterised by straight grain boundaries of and triple junctions between quartz grains. Moreover, these quartz grains did not exhibit undulose extinction. These three characteristics suggest that static recrystallisation has transpired (Passchier and Trouw, 2005). Static recrystallisation is a process in which the grains strive towards a strain-free or low energetic stable state, through recrystallization and grain boundary migration (Passchier and Trouw, 2005). As grain boundaries migrate towards their centres of curvature, this will result in a disappearance of the smaller grains and a growth of the larger grains. Hence, this can also explain why the grains in this sample were more or less equigranular. (This process is also responsible for the formation of the granoblastic texture of the more or less monocrystalline quartzites). According to Vernon (2004) and Passhier and Trouw (2005), static recrystallisation only occurs after deformation. Seeing the proximity of sample ST-44 to the KMF, this could be related to a major shear phase along this fault. Sample ST-40 had a similar texture, although some quartz grains still showed an undulose extinction. Contrary to sample ST-44, this sample was not located close to a major fault zone. For an explanation for the occurrence of this energy-free configuration a more thorough sampling would be required.

Symplectic intergrowth structures can also indicate deformation. As previously mentioned a symplectite is a vermicular intergrowth of simultaneously formed minerals. They can either be the result of an the eutectic crystallization of two minerals from a melt or of an eutectoid (or solid-state) transformation (Vernon, 2004). The former mechanism is typically invoked to explain the granophyric symplectites. Hence, this texture is not related to deformation. Myrmekitic symplectites, on the other hand, typically occur in deformed granites, even if this deformation is limited to e.g. undulose extinction of quartz (Vernon, 2004). They are considered to be the result of solid state transformation of K-feldspars that accompanies P-T changes during deformation (Vernon, 2004; Passchier and Trouw, 2005). This explains why the observed myrmikitic symplectites (ST-42, 43, 44, 45) tended to overgrow alkali-feldspars (figure 6.1b) and why this intergrowth was only observed in
the samples (ST-42, 43, 44, 45), as these samples were located quite close to the fault zones (figure 4.3 and 7.5).

7.1.2 Mineral alteration

The most striking alteration was that of feldspars to sericite (a fine grained variety of muscovite) or to kaolinite. Sericitization can result from either pneumatolytic or hydrothermal alteration (Pichler and Schmitt-Riegraf, 1997). The latter of these being the most likely as kaolinisation and chloritisation (the alteration of biotite to chlorite) were also observed (Pichler and Schmitt-Riegraf, 1997). Moreover, some authors (e.g. Imai et al., 2013) have described the granites of the Phuket area as hydrothermally altered. Besides these two common types of weathering, plagioclase was also altered to calcite in the Ca-rich I-type granite (sample ST-33) (see below).

Note that since Thailand has a tropical climate, climate-controlled weathering, especially for kaolinisation, cannot be excluded. For example, some samples also showed oxidation along fractures, indicating this was the result of percolating water and hence indicating a climate-controlled weathering. So the observed weathering was probably a combination of both in situ weathering and hydrothermal weathering.

7.2 Granite classification

7.2.1 I-versus S-type granites and implications for the geodynamic context

As already discussed in section 4.3.2.2, I- and S-type granites are important indicators of the geodynamic context. Currently, the use of the term I and S-type granites is in decline and the terms metaluminous and peraluminous respectively are preferred. However, since the literature concerning the Western Granites exclusively uses the older terminology, this will be maintained in this work. A distinction between the two can be made based on the so-called Aluminum Saturation Index or ASI, which is the molar ratio of \( \frac{Al_2O_3}{(CaO+Na_2O+K_2O)} \). If this ratio is smaller than one, then the granites are classified as I-types (or metaluminous) (figure 7.3a). If this ratio is greater than one and if \( Na_2O+K_2O > Al_2O_3 \), granites are classified as S-type (or peraluminous) (figure 7.3a). If ASI> 1 but \( Na_2O+ K_2O < Al_2O_3 \) the granite is classified as peralkaline (Frost et al., 2001). Table 7.1 shows the classification of the studied samples based on the ASI. With one exception, the resulting classification corresponds with the one Cobbbing (2011) had previously drafted for the different plutons.
Frost et al. (2001) suggested to take the amount of Ca incorporated in apatite into account by calculating the ASI as the molar ratio of $\frac{Al_2O_3}{(CaO+Na_2O+K_2O-1.67P_2O_5)}$. As can be derived from Table 7.1, the difference between both ASI-values is negligible.

Table 7.1: The I and S-type classification of the ST-samples, based on the ASI and on mineralogical characteristics. For more information see text. Note that the second row represents the classification drafted by Cobbing (2011).

<table>
<thead>
<tr>
<th>Sample</th>
<th>ST-33</th>
<th>ST-36</th>
<th>ST-37</th>
<th>ST-38</th>
<th>ST-40</th>
<th>ST-42</th>
<th>ST-43</th>
<th>ST-44</th>
<th>ST-45</th>
<th>ST-54</th>
</tr>
</thead>
<tbody>
<tr>
<td>$Al_2O_3-(Na_2O+K_2O)$</td>
<td>0,05</td>
<td>0,02</td>
<td>0,04</td>
<td>0,03</td>
<td>0,03</td>
<td>0,04</td>
<td>0,04</td>
<td>0,03</td>
<td>0,03</td>
<td>0,04</td>
</tr>
<tr>
<td>ASI (Frost et al., 2001)</td>
<td>0,91</td>
<td>1,12</td>
<td>1,02</td>
<td>1,16</td>
<td>1,23</td>
<td>1,20</td>
<td>1,15</td>
<td>1,05</td>
<td>1,27</td>
<td>1,41</td>
</tr>
<tr>
<td>Classification</td>
<td>I-type</td>
<td>S-type</td>
<td>?</td>
<td>S-type</td>
<td>S-type</td>
<td>S-type</td>
<td>S-type</td>
<td>S-type</td>
<td>S-type</td>
<td>S-type</td>
</tr>
<tr>
<td>Classification based on Mineralogy</td>
<td>I-type</td>
<td>?</td>
<td>S-type</td>
<td>S-type</td>
<td>S-type</td>
<td>S-type</td>
<td>S-type</td>
<td>?</td>
<td>?</td>
<td>?</td>
</tr>
</tbody>
</table>

Table 7.1 also show a classification based on mineralogy and on biotite colour. This difference in mineralogy (see also section 4.3) is simply an expression of the geochemical difference. I-type granites tend to have an excess of Ca with respect to Al. Meaning that although Al is fully accommodated by feldspars, Ca is only partly accommodated by feldspar. Therefore other Ca-rich minerals have to be present (Frost et al., 2001). For example, in sample ST-33 hornblende ($Ca_2(Mg,Fe)Al_2Si_2O_5(OH)_2$) and titanite ($CaTiSiO_5$) were observed together with (Al-poor) biotite ($K(Fe^{2+},Mg)$_2($Fe^{3+},Al,Ti)$_x(Al)_{1-x}Si_3O_{10}(OH)_2$). S-type granites, on the other hand, contain more Al than feldspar can incorporate and hence minerals rich in Al, but rather poor in Ca, Na and K. Some typical examples that were observed are Al and Ti-rich biotite and muscovite ($KAlSi_3O_{10}(OH)_2$) (Frost et al., 2001). Other typical minerals such as cordierite and garnet were not observed. As mentioned, another criteria for the distinction between I and S-types is the colour of biotite (figure 7.3b), which is related to the Ti- (Fe and Mg) content. The Al and Ti-poor biotite in I-types typically have a dark brown colour to brown green, whilst the Al and Ti-rich feldspar of the S-type granites show a strong pleochroism from yellow to reddish brown. Although these two characteristics conform the I-type character of ST-33, they do not provide any clarity on the classification of sample ST-45.
To summarise, the samples from the Phuket-Takua Pa Batholith are S-type granites, whilst the sample from the Khao Phanom Bencha Batholith is an I-type. Hence, this is another inter-batholith difference.

As mentioned in section 4.3.2.2, the occurrence of I- and S-type granites can be linked to specific geodynamic settings. Given the tectonic history of this area, the occurrence of I- and S-type granites in close proximity, probably indicates a continental volcanic arc setting. In this case, this arc would be related to an Andean-type subduction of the Neotethys underneath Thailand. This is in support of the model proposed by Cobbing (2011). Based on geochemical analysis throughout the entire Southeast Asian Tin Belt, this author proposed that the granites of the Western Belt are related to a Cretaceous arc that stretched from Myanmar all the way south to Phuket.

7.2.2 Classification based on QAPF-diagram
For this classification see section 6.1.2. This classification also reflects the difference between I- and S-type granites. Whilst the S-type granites are predominantly syenogranites, the I-type granite is a monzogranite (figure 6.4). This is in agreement with the findings of Cobbing (2011) that I-type granites range from gabbro to monzogranite, whilst S-type granites range from granodiorite to syenogranite.

7.3 AFT-thermochronology
7.3.1 The Cenozoic history of Peninsular Thailand
First of all it should be noted that the following discussion is based on a single t-T-path (of sample ST-41), which might not be representative for the entire Lower Peninsular. However, since the AFT-ages are so consistent, it is likely that the samples would have known a similar cooling history.

In figures 7.4 and 7.5 the AFT-ages for the ST-samples are presented on the geological map and on DTM (Digital Terrain Model) of the study area. Most AFT-ages are Late Oligocene to Middle Miocene and range between 30 and 16 Ma. Thus, the AFT-ages of most ST-samples reflect the tectonically relative stable phase in between the final faulting and exhumation along the KMF and RF zones (between 37-28 Ma) and the opening of the Andaman Sea (around 11 Ma). Hence, we interpret that these ages indicate that the granite bodies were brought up to depths within the APAZ due to fault-related exhumation and erosion. Further erosional denudation, related to the opening of the Andaman Sea and the associated base-level drop eventually brought the samples to the surface. This implies that the samples have resided in the APAZ for a non-negligible time span, which explains the shortened track length of 12.4 µm and a
somewhat broad distribution with a $\sigma$ of 1.9 µm. Again, it should be stressed that due to the young AFT-ages of the samples, track densities, especially confined track densities are very low and thus track length information for the ST-samples is very limited.

Sample ST-45 has a somewhat anomalous higher AFT-age of 49 ± 7 Ma. This might be due to several different causes. For example, it could simply be due to a compositional variation. Therefore the $D_{par}$ (section 2.1.2.1) of sample ST-45 (with a mean 1.6 ± 0.2 µm) was compared to that of the neighbouring samples (ST-41, ST-42, ST-43 and ST-44, with a mean of 1.5 ± 0.2 µm). As the resulting distribution do not differ significantly (all p-values > 0.1 for every combination, determined by means of the Tukey’s honestly significant difference test), this is probably not the case.

![Figure 7.4: The geological map of the study area southern Peninsular Thailand with indication of the AFT-ages (in Ma). For legend and the map of the entire area, see appendix C (from the department of mineralogical resources).](image)

The somewhat anomalous value could also be due to the location of the sample. As mentioned in section 2.2.3.2, a horizontal age difference can be due to a difference in denudation. In this case, the
denudation of sample ST-45 would have been slower than of its surroundings and hence differential erosion might have removed the sediments slower leading to slower cooling and hence an older AFT-age. Based on figure 7.5, this seems explanation seems plausible as the samples to the east of the KMF-zone (ST-33, ST-45) have somewhat higher ages than those to the west of the KMF zone, suggesting that the eastern area has known a slower erosion of the sedimentary cover. A more thorough sampling would be required to verify this hypothesis.

The stratigraphic position could also explain a difference in age, as the higher a sample is located in the stratigraphic column, the earlier it would have crossed the $T_c$-isotherm. Based on the age-elevation plot (figure 7.6) this does not seem to be the case, however this plot does not necessarily provide information on the stratigraphic position.

Figure 7.5: Digital terrain model of the study area in Peninsular Thailand with indication of the AFT-ages (in Ma). The position of the The Khlong Marui Fault (KMF) is based on Watkinson et al.(2008) and Morley et al.(2011). This map was plotted in by means of the Global Mapper software.

Figure 7.6 represents the age-elevation plot. Two possible regression lines with a more or less linear relation between age and elevation can be plotted, depending on whether or not the anomalous
value of ST-45 was taken into account. Note that although a similar relation was to be expected, the relief is relatively smooth and elevation differences are rather limited, so two possible regression lines could simply indicate scatter instead of a linear relation.

![Age-elevation plot]

**Figure 7.6:** The age-elevation plot of the ST-samples.

In short, the stepwise cooling shown in figure 6.8 can be interpreted as the result of (tectonically induced) erosional denudation. The initial rapid cooling (with a modelled cooling rate of 9.0 °C/Ma) indicates that the sample entered the AFT-registration thresholds when the region was already tectonically active. As mentioned in section 6.3.4, this initial cooling phase lasted until approximately 29-26 Ma and was followed by a more stable thermal period (figure 6.8). The ages of our ST-samples correspond remarkably well to the timing of fault activity and exhumation along the KMF and the RF zone. As discussed in section 4.1.3 these two faults, which bound the study area, have known two major periods of dextral shear followed by a final brittle strike-slip phase. The second shear phase transpired between ca. 48-40 Ma (cfr. age of ST-45) and was rapidly followed by the brittle phase around 37 Ma, this timing roughly corresponds to the moment in time when the granites entered the AFT-window of registration. As mentioned in section 4.1.3.2, the second shearing was already accompanied by exhumation. However, the exhumation was not completed until the end of the brittle faulting phase at approximately 28 Ma. As indicated on figure 6.8, the deceleration in cooling corresponds to the end of this exhumation phase. So this first rapid cooling phase can be interpreted in terms of denudation and exhumation, since exhumation of the ductile fault cores, and its associated erosion of the sedimentary cover, can cause denudation and cooling (see figure 2.7). Other contributing processes cannot be excluded. For example, the Mergui and Sumatra Basin opened around the same period (32 Ma). It is plausible that the base-level drop resulting from their opening, contributed to the denudation and hence to the cooling. Especially since the Oligocene
formation in the Mergui basin consists of (amongst others) conglomerate and fluviatile sandstone (Morley and Racey, 2011), indicating intense erosion.

From the Oligocene (28 Ma) to until the Middle Miocene (approximately 11 Ma), no major tectonic changes along the passive continental margin of Peninsular Thailand occurred. This corresponds to
Figure 7.7: a) Tectonic map of Peninsular Thailand showing the AFT-ages of Upton (1999) (after Morley, 2004), the red box indicates the current study area; b) comparison of the AFT-ages obtained by Upton (1999) for Peninsular Thailand and those from this thesis along a north-south transect.
the more or less thermal stable phase in the t-T-path (figure 6.8). Around 13-12 Ma, the t-T past shows an accelerated cooling (from 1.2 °C/Ma to 5.1 °C/Ma) which lasted until the present. This accelerated cooling can be linked to the opening of the Andaman Sea (which is summarised in figure 4.9). As mentioned in section 4.2, the opening of the Andaman Sea occurred in two phases, with the first one starting in the Middle Miocene. The opening of the Andaman Sea implies a strong base-level drop, which can cause (erosional) denudation of Peninsular Thailand, which in its turn resulted in an increased cooling. As the associated spreading is still active, the cooling is still on going. Evidence for this erosion has been found the Mergui Basin (see section 4.2.1.2).

We interpret that the bulk of our AFT-ages should be placed in this entire context (meaning from stability to cooling).

As previously mentioned, the area north of the this study area has been extensively dated by e.g. Upton (1999). The results for Peninsular Thailand are displayed in figure 7.7a. As can be derived from figure 7.7b, our results (and findings) correspond well. For example, this author also concluded a cooling through exhumation and erosion between 44 and 20 Ma for the fault zones in Peninsular Thailand.

7.3.2 The cooling history of the Madras block

![Figure 7.8](image-url)

Figure 7.8: The geological map of the study area in the Tamil Nadu State in southern India with indication of the AFT-ages (in Ma). For legend see figure 3.1.
In figures 7.8 and 7.9, the AFT-ages for the TN-samples are presented on the geological map and on the DTM (Digital Terrain Model) of the study area. Except for TN-36 (which has an age of 261 Ma), the AFT-ages of the TN samples seem to reflect the break-up of Gondwana (ranging between 112-187 Ma) (figure 3.2). These ages are interpreted as a cooling age through the APAZ temperatures, which most likely results from a slow denudation. This denudation can be related to (slow) exhumation and erosion and was probably enhanced by a base-level drop due to rifting of India during the break-up of Gondwana around 130 Ma (figure 3.6c). The AFT-track length distribution (figure 6.7) clearly indicates a significant partial annealing has occurred, as the tracks have been shortened and the distribution is unimodal with a σ between 1.3-1.7 µm. The mean track length (11.4-12.0 µm) is shorter than would be expected for an undisturbed basement type or slow cooling (12.5-13.5 µm), implying that the samples have resided in the APAZ for a non-negligible time. This indicates that a further (more recent) cooling and hence denudation and exhumation are needed in order for the samples to reach surface temperatures.

Figure 7.10 shows the age-elevation plot, the regression line also shows an expected linear relation between age and elevation. A similar relation had already been established for the Dharwar craton (Gunnel et al., 2003).
In short, the stepwise cooling shown in figure 6.8 can be explained through tectonic and erosional denudation. The discussion of the t-T-paths will focus on that of samples TN-35, 38 as these required the least constraints for modelling. Hence, artificially created effects are avoided. Despite their individual differences, all TN-samples indicate a slow cooling phase (with a cooling rate of around 1 °C/Ma) throughout the Palaeozoic and at least a part of the Mesozoic. This indicates that the samples entered the AFT-system during a phase of on going denudation, which probably started after the Malagasy Orogeny (section 3.3). Hence, this denudation can be related to slow exhumation and erosion.

Prior to modelling (and based on previous research such as Gunnell et al., 2003), the t-T-paths of the TN-samples were expected have been affected by the break-up of Gondwana. Especially erosion associated with the base-level drop resulting from the rifting of India from Antarctica-Gondwana and the opening of the Indian Ocean, was considered to have caused denudation. Based on figure 6.8, only sample TN-35 meets this expectation, as this t-T-paths only stabilises after the separation of India from Antarctica-Australia and the opening of the Indian Ocean (around 125-130 Ma) (section 3.4.3). In the t-T-path of samples TN-36, 37 the early thermal stabilisation could be related to the extra constraint that was used to model the thermochronologic history. However, the t-T-path of samples TN-38 also reaches stability between 200-180 Ma, which indicates that this sample was not affected by denudation as a result of the aforementioned opening of the Indian Ocean. Mind that this apparent stability could be due to the temperature boundaries of the APAZ. As indicated in figure 6.8, TN-38 reached the 60°C-isotherm (or upper APAZ- AFT TSZ boundary) at around 200 Ma, meaning that a further cooling would no longer have been recorded in the lengths of the AFT (section 2.2.1.2). Sample TN-35 on the other hand, reached the 60°C-isotherm later at around 70 Ma. Hence any additional cooling or denudation in between 200 Ma and 70 Ma would have been recorded in this sample. The same reasoning can also be applied for the other two samples.
So, the four TN-samples have known a continuous cooling from the Ordovician (which is probably a result from the Malagasy Orogeny and hence had already started in the Cambrian) onwards until around 200-120 Ma, when they reached thermal stability. This thermal stability corresponds to the development of the eastern Indian passive continental margin, which was established after the opening of the Indian Ocean.

After this phase of thermal stability, all four t-T-paths demonstrate a final Cenozoic cooling event (at a cooling rate of approximately 1.3 °C/Ma), which started around 45-20 Ma and brought the samples to the surface conditions. This cooling phase is considered to be related to the tectonic effects of the India-Asia collision. More specifically, this cooling phase can be related to crustal shortening and thrusting, which also contributed to the development of the Himalaya-Tibet Plateau and probably also caused accelerated denudation in continental India and its passive margins.

These results are in line with those of Biswal and Seward (2003) and Gunnell et al. (2003). For example, Biswal and Seward (2003) also described a slow cooling from the Ordovician until approximately 120 Ma. They also found a heating effect of the Kergeulen plume. A similar heating effect was not observed in our modelled t-T-paths. This difference can possibly be explained due to the distance from the Kergeulen plume, as the study area of Biswal and Seward (2003) was situated far more to the north and hence a lot closer to the Kergeulen plume (see figure 7.11).

Gunnell et al. (2003) performed an extensive AFT-study in order to reconstruct the denudation history of the passive continental western Indian margin. Even though they worked on the opposite margin of our study area (see figure 7.11), they still found an increase in denudation around 120 Ma, which they attributed to the of the separation of India from Antarctica-Australia. Depending on the model parameterisation, they also found an increase in denudation during the Cenozoic (since around 50 Ma), which they interpreted as a response to far-field lithospheric stresses. The India-Asia collision was proposed as a possible cause for these stresses. The authors do highlight that the precise timing and magnitude of denudation strongly depends on model parameterisation such as initial track length.

![Figure 7.11](image.png)

*Figure 7.11: Schematic illustration of India indicating the positions of the study area of 1) Biswal and Seward (2003); 2) this thesis and 3) Gunnell et al. (2003). The red dot indicates the position of the Kergeulen plume.*
Chapter 8: conclusion

Based on the results presented in chapter 6 and the interpretation and the discussion in chapter 7, the following conclusions can be drawn.

(1) Petrographically, the ST-samples can be classified as syeno- or monzogranites with a predominant holocrystalline, phaneritic, inequigranular texture. However, there are subtle differences between the individual batholiths and plutons. These findings are in-line with those of Cobbing (2011).

(2) These textural characteristics typically indicate an emplacement of these granites at relatively deep crustal levels, allowing a slow (pneumatolitic post-magmatic) cooling. The undulose extinction and sutured texture of quartz that were observed in most samples were also interpreted as a result of burial.

(3) The observed weathering was interpreted to be the result of hydrothermal activity, although the tropical climate of Thailand probably also caused, or at least contributed to, recent weathering of the crystalline basement.

(4) Based on mineralogical properties and the chemical analysis, most samples from southern Peninsular Thailand (ST-samples) could be classified as S-type (or peraluminous) granites, with sample ST-33 as the sole exception as this was classified as an I-type granite (metaluminous). Again, these findings are in agreement with those of Cobbing (2011). This mutual occurrence of I- and S-type granites has important implications for the geodynamic setting, as this indicates a continental volcanic arc setting related to the subduction of the Neotethys underneath Thailand during the Cretaceous.

(5) The major aim of this thesis was to fill the gap in the existing record and to determine the thermochronologic history of Lower Peninsular Thailand by means of the AFT-method. The AFT-ages (table 6.8) of most samples range between 30-16 Ma (or Oligocene to Middle Miocene). These ages corresponds to a phase of relative tectonic quiescence between the final reactivation and exhumation along the RF (Ranong Fault) and KMF (Khlong Marui Fault) zones and the opening of the Andaman Sea. These findings are in agreement with those of Upton (1999).

(6) By means of the thermal history modelling, the Cenozoic thermochronologic history of one of the Southern Peninsular Thailand samples was reconstructed (figure 6.8). The following three phases were recognised: (i) after emplacement of the granites in the Cretaceous at relatively deep crustal levels, the granites were rapidly exhumed along the KMF (and probably also the RF) zone in the Early Cenozoic. This exhumation (and the associated erosion) led to
denudation and rapid cooling of the basement, which lasted until ca. 29-26 Ma, at which time both the activity of the fault zones and the exhumation along the fault zones had ceased; (ii) from the Oligocene (ca. 29-26 Ma) until the Middle Miocene (ca. 13-12 Ma), no major tectonic changes occurred. During this period, the samples experience a phase of more or less thermal stability as they remained at approximately the same crustal levels reached at the cessation of the previous cooling phase. However, the samples were not yet fully exhumed to the surface, as the t-T-model indicates that they remained at a temperature of around 100 ± 10°C (or a depth of roughly 3-4 km considering a “normal geothermal” gradient); (iii) from the Middle Miocene onwards, the samples were further brought to the surface by means of erosion due to the base-level drop associated with opening of the Andaman Sea. This erosional denudation invoked a final phase of rapid cooling, which is currently still active.

Note that the somewhat older AFT-ages of samples ST-33 and ST-45 suggest a differential erosion along the KMF zone: the area located to the east of this fault could have known a slower denudation and erosion. However, further sampling is required to verify this hypothesis.

(7) The samples of the Tamil Nadu State (or TN-samples) all yielded an Early Jurassic to Early Cretaceous age (112-187 Ma) (table 6.8), with the exception of TN-36 (which yielded an age of 261 Ma). These ages were interpreted as cooling ages reflecting the break-up of Gondwana.

(8) By means of thermal history modelling, the thermochronologic history of the TN-samples starting from the Palaeozoic could be reconstructed (figure 6.8): (i) after the Malagasy Orogeny, the TN-samples were subjected to a slow cooling throughout most of the Palaeozoic and a part of the Early-Middle Mesozoic (until about 200-120 Ma). This cooling is considered to be, at least in part, a result of denudation related to the break-up of Gondwana, more specifically to the separation of India (including the Tamil Nadu basement) from Gondwana around 130 Ma; (ii) after the rifting of India from Gondwana and after the associated opening of the Indian Ocean (around 130-125 Ma), the eastern Indian margin evolved into a passive margin. This period of tectonic quiescence translates to a phase of thermal stability in the t-T-path; (iii) at around 45-20 Ma, the samples experience another phase of denudation induced cooling, which led to a Cenozoic cooling event. This more recent denudation is interpreted as a distal effect of the India-Asia collision and bulging of the Indian plate.

This reconstruction underscores those of Biswal and Seward (2003) and Gunnell et al. (2003).
We propose two potential avenues of future research for Southern Peninsular Thailand. First of all, irradiating the samples from South Peninsular Thailand with a $^{252}$Cf-source or preferably by heavy ions (section 1.2.4) could increase the number of confined tracks and hence this could allow statistically more robust AFT-length data to be incorporated in thermal history modelling of the samples. An (U-Th)/He analysis of apatites (and zircons), which has a lower closing temperature than the AFT-system, would allow the determination of exhumation rates. Another topic for future research in Peninsular Thailand could be to further investigate the possibility of differential denudation along the KMF (and RF) zone by means of e.g. $^{40}\text{Ar}/^{39}\text{Ar}$-dating, which is used for dating fault movements.

For the Tamil Nadu samples we also suggest an (U-Th)/He analysis as this could help to refine the thermal histories.


References


References


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**Nederlandstalige samenvatting (Dutch Summary)**

**Doelstelling**
Deze thesis had als voornaamste doel om een beter inzicht te krijgen in de evolutie van de passieve rand het schiereiland van Thailand en ook om de bestaande gegevens in zake petrogenese en (thermo)chronologische evolutie van Thailand te vervolledigen. Hierdoor werden 12 granitoïden uit Zuid-Thailand langs een N-Z profiel geanalyseerd met de apatiet fissiesporen (AFT)-methode. Momenteel een veelgebruikte en krachtige analyse methode voor het reconstrueren van een (lage temperatuur) thermochronologische evolutie (Tagami and O’Sullivan, 2005). De AFT-methode werd ondersteund door zowel een petrografische als chemische analyse van de granitische gesteenten.

Hieraan werd ook een O-W profiel van kristallijne sokkelgesteenten uit de Tamil Nadu Staat in Zuid(oost)-India, gelegen aan de overzijde van de Indische Oceaan (en van de aangrenzende Andaman Zee), geanalyseerd met de AFT-methode. Het doel van deze analyse was van tweeërlei aard; ten eerste liet deze analyse toe een beter inzicht te verwerven in de evolutie van beide passieve randen van de Indische Oceaan (en van de Andaman Zee). Ten tweede liet deze analyse ook toe om de techniek van de thermochronologische analyse beter onder de knie te krijgen, gezien de stalen uit Thailand door een laag gehalte aan inwendige sporen zich niet echt leenden voor een thermochronologische modellering.

**Theoretische achtergrond**
De apatiet fissiesporen methode (AFT-methode) steunt op de spontane fissie van $^{238}$U-nuclei. Als een gevolg van deze nucleaire fissie, zullen fissiefragmenten zich aan hoge snelheid doorheen het kristalrooster bewegen. Hierdoor worden lijnvormige beschadigingssporen gecreëerd, de zogenaamde fissiesporen (FT). De densiteit van de geaccumuleerde spontane (oppervlakte) sporen gecalibreerd tegen de huidige concentratie van $^{238}$U is een maat voor de verstrekken tijd (Tagami and O’Sullivan, 2005). Deze laatste wordt bepaald door artificieel een fissie reactie te induceren door de stalen te bestralen in een nucleaire reactor met thermische neutronen en dan de oppervlakte dichteit van deze zogenaamde geïnduceerde sporen te bepalen.

De sluitingstemperatuur van het AFT-systeem bedraagt ongeveer 100 °C ± 20 °C, waardoor het systeem gevoelig is voor lage temperaturen. Bijgevolg stellen de AFT-ouderdommen meestal afkoelingsouderdommen i.p.v. vormingsouderdommen voor. In deze thesis werd er gebruik gemaakt van de externe detector (ED) methode in combinatie met de $\zeta$-kalibratietechniek. Deze combinatie laat toe de ouderdom te berekenen uit de verhouding van de oppervlakte dichtheid van spontane sporen t.o.v. geïnduceerde sporen. Hierbij worden de spontane sporen geteld op een gepolijst en
geëtst oppervlak van de apatiet korrel. De oppervlakte dichtheid van de geïnduceerde sporen wordt bepaald a.d.h.v. een telling op de ED (in dit geval muscoviet). Gezien een kristal de neiging heeft om zijn kristalrooster te herstellen bij hogere temperaturen, zullen de sporen bij dergelijke verhoogde temperaturen verkorten (een proces genaamd uitgloeiing) en uiteindelijk zullen ze volledig verdwijnen (of vervagen). Voor apatiet in het bijzonder begint deze uitgloeiing rond de 60 °C en de sporen zullen volledig vervagen vanaf 120 °C. Naast een verandering van de sporenlengte zorgt dit dus ook voor een verlaging van de sporendichtheid. Dit uitgloeingsproces heeft als groot voordeel dat de lengteverdeling en de gemiddelde sporenlengte van fissiesporen gebruikt kunnen worden voor het reconstrueren van de thermische geschiedenis. Er bestaan verscheidene modellen die de uitgloeiing van fissiesporen in apatiet beschrijven. In deze thesis werd er gebruik gemaakt van het Ketcham et al. (2007) model. Let wel dat lengtemetingen, in tegenstelling tot de dichtheidsbepaling, uitgevoerd worden op inwendige sporen (Engels: confined tracks) en niet op oppervlakte sporen. Voor de stalen van Zuid-Thailand bleek het meten van een statistische aanvaardbare hoeveelheid inwendige sporen problematisch te zijn, omdat deze jonge stalen een laag U-gehalte hadden. Bijgevolg hadden deze stalen een zeer lage sporendichtheid.

Geologische achtergrond.

Beide studyegebieden zijn gelegen in Zuid(-Oost) Azië en kenden een complexe geologische geschiedenis die teruggaat tot het Paleozoïcum.

In Zuid-India werkte deze vervorming in op een sokkel die opgebouwd is uit verschillende cratonische blokken uit het Archeaan en Proterozoïcum.

Figuur 3.2 geeft een overzicht van de tektonische geschiedenis van India, deze kan opgesplitst worden in drie belangrijke gebeurtenissen: (1) de Laat Proterozoïsche Malagasië orogenese die initiëerde als een Pacifisch-type orogenese in het Neoproterozoïcum en evolueerde naar een Himalaya-type tijdens het Cambrium. Deze orogenese zorgde voor de amalgamatie van de verschillende crustale blokken die Zuid-India opbouwen. Hiermee was ook ineens de amalgamatie van Gondwana voltooid (Santosh et al., 2009; Saitoh et al., 2011; Collins et al., 2007; 2013); (2) tijdens het Jura begon India zich af te splitsen van Gondwana. Deze fase werd gevolgd door een noordwaartse drift tijdens het Krijt, die zorgde voor de opening van de Indische Oceaan (rond 130-125 Ma) ten zuiden van het migrerend blok en de sluiting van het Tethys bekken ten noorden ervan (Chatterjee et al., 2013); (3) de laatste grote tektonische gebeurtenis was de botsing van India met (Eur)Azië en de hiermee gepaarde Himalaya orogenese. Momenteel heerst er nog discussie over het exacte tijdstip van de botsing: 55 Ma (vb. DeCelles et al., 2002) of 35 Ma (vb. Aitchison et al., 2007).
Thailand kende een volledig verschillende geodynamische geschiedenis, die sterk beïnvloed was door zijn specifieke tektonische opbouw. Thailand kan opgesplitst worden in twee continentale blokken; het Indochina Blok en het Sibumasu Blok. Enkel de laatste is van belang voor deze thesis. Verder kan Thailand onderverdeeld worden in drie zogenoemde Graniet Provincies: een Westelijke, een Oostelijk en een Centrale, waarvan enkel de Westelijke van belang is voor deze thesis.

De tektonische en magmatische geschiedenis van het Sibumasu Blok en de Westelijke Graniet Provincie van Thailand worden samengevat in figuur 4.1. Drie processen zijn van groot belang voor deze thesis: (1) het magmatisme tijdens het Krijt dat verantwoordelijk was voor de intrusie van de bestudeerde granietlichamen. Dit magmatisme wordt gezien als een gevolg van de subductie van de Neotethys Oceaan onder het Sibumasu Blok (Crow and Khin Zaw, 2011; Watkinson et al., 2011); (2) de activiteit van de Khlong Marui Breukzone (KMF) en de Ranong breukzone (RF). Deze breuken kenden twee fasen van dextrale shear, één tijdens het Laat-Krijt (> 80 Ma) en één tijdens het Midden-Mioceen (48-40 Ma). Deze laatste fase ging waarschijnlijk al gepaard met een significante exhumatie. Na een periode van rust, werden beide breuken gereactiveerd als sinistrale strike-slip breuken tijdens het Laat Eoceen (37-30 Ma) (Watkinson et al., 2008; 2011). Tijdens deze reactivatie ging de exhumatie ook verder; (3) de opening van het Andaman Zee. Deze opening verliep in twee fasen van extensie, een eerste tijdens het Midden-Mioceen (rond 11 Ma) en een tweede fase tijdens het Laat-Mioceen tot Vroeg-Plioceen (5-4 Ma) (Khan and Chakraborty, 2005).

**Stalen en Methodes**

De voorbereidingsfase bestond voornamelijk uit het separeren van apatiet. Dit kan worden samengevat in de volgende vijf stappen: (1) het verbrijzelen van de gesteentemonsters; (2) een droge en natte zeving; (3) een magnetische separatie; (4) een zware vloeistofscheiding en (5) het handmatig selecteren van een honderdtal apatietkorrels onder een Leica M16 FA stereo microscoop. Na het separatieproces werden de apatietkorrels ingebed in epoxy, gepolijst en geëtst gedurende 70 s in een 2.5 % HNO₃ oplossing bij een temperatuur van 25 °C. Na het etsen werden de monsters afgedekt met een Goodfellow Clear Ruby muscoviet als externe detector (ED). De etsing van apatiet was nodig om de spontane sporen waarneembaar te maken onder een microscoop. De monsters werden samen met apatiet ouderdomstandaarden (Durango Apatiet en Fish Canyon Tuff Apatiet) en IRMM-540 glas dosimeter geplaats in een bestralingspakket (figuur 5.5) om dan vervolgens samen bestraald te worden in een goed gethermaliseerd kanaal (X26) van de BR1-onderzoeksreactor van het Belgisch Nuclear Onderzoekscentrum (SCK-CEN) in Mol. Na bestraling werden de muscoviet-ED geëtst gedurende 40 minuten in een 40% HF oplossing bij 20 °C, teneinde de geïnduceerde sporen zichtbaar te maken. Vervolgens werd de oppervlakte dichtheid van de spontane en geïnduceerde sporen bepaald d.m.v. tellingen onder een Olympus BH-2 microscoop tegen een vergrooting van
1250X. Er werd telkens getracht om een duizendtal spontane sporen te tellen. Deze waarden werden gebruikt om de ζ-ouderdom te berekenen. Ten slotte werden de lengtes van ongeveer 100 inwendige sporen gemeten m.b.v. een KONTRON-MOP-AMO meetsysteem. Zoals eerder vermeld, was dit aantal niet haalbaar voor de jonge ST-stalen uit Zuid-Thailand. De AFT-ouderdommen, gemiddelde sporenlengtes en lengteverdeling konden vervolgens gebruikt worden om de t-T-paden van de stalen te reconstrueren m.b.v. het HeFTy-programma en het Ketcham et al. (2007) uitgloeingsmodel. De resultaten worden weergegeven in tabel 6.8, figuur 6.7, 6.8.

Naast deze AFT-analyse werd ook een petrografische en chemische analyse uitgevoerd op de ST-stalen. De petrografische analyse kwam neer op een slijpplaatanalyse m.b.v. een Olympus BH-2 microscoop.

De chemische analyse bestond uit een ICP-OES (Inductively Coupled Plasma-Optical Emission Spectroscopy) analyse van de hoofdelementen. Deze analyse steunt op het principe dat geëxiteerde atomen een elektromagnetische straling uitzenden als ze terugvallen naar hun grondtoestand. Gezien de golflengte van deze straling element-specifiek is en gezien de intensiteit van deze straling afhangt van de concentratie van de aanwezige elementen, laat deze analyse een kwalitatieve en kwantitatieve bepaling van o.a. de hoofdelementen toe (tabel 6.5).

Resultaten en interpretatie
Voor een uitgebreide beschrijving van de mineralogie en de texturen wordt de lezer verwezen naar tabel 6.1, 6.2, 6.3.

Uit de petrografische analyse van de ST-stalen konden we besluiten dat de stalen syeno- of monzogranieten zijn met over het algemeen een holokristallijn-e, faneritische, inequigranulaire textuur. Echter waren er kleine onderlinge verschillen tussen de verschillende batholieten en plutonen. Dit komt overeen met eerdere beschrijvingen van de Westelijke granietgordel (vb. Cobbing, 2011). Een dergelijke textuur wijst er op dat de granieten vrij diep in de korst intrudeerden en bijgevolg traag konden afkoelen. Dit werd verder bevestigd door de textuur van de kwarts korrels. Hun undueuze uitdoving en sutuur contacten wijzen immers ook op een begraving, wat zo zijn implicaties heeft voor de interpretatie van de AFT-ouderdom en de t-T-paden.

Het type graniet (I of S) geeft informatie over de geodynamische setting. Uit zowel de chemische analyses als mineralogische kenmerken volgde dat, met een enkele uitzondering (ST-33), alle stalen behoorden tot het S-type. Deze bevinding kwam opnieuw overeen met die van Cobbing (2011). Gezien de geologische geschiedenis van dit gebied, wijst het samen voorkomen van I- en S-typen granieten op een continentale vulkanische boog die gelinkt zou zijn de subductie van de Neotethys onder Thailand tijdens het Krijt.
De AFT-resultaten worden weergegeven in tabel 6.8 en de t-T-paden in figuur 6.8. De (meeste) ST-stalen leverden een Cenozoïsche AFT-ouderdom op. Deze ouderdom reflecteert de stabiele fase tussen enerzijds de laatste reactivatie (en de hiermee gepaard gaande exhumatie) langs de RF en KMF zone (rond 37-28 Ma) en anderzijds de opening van de Andaman Zee (rond 11 Ma). Vanwege een te lage dichtheid van intrinsieke sporen, kon slechts enkel voor monster ST-41 het t-T pad geclassificeerd worden. Het model suggereert een afkoeling doorheen het Cenozoicum en kan onderscheid worden in drie fasen: (1) een eerste snelle afkoeling tot ongeveer 29-26 Ma. Deze fase wordt gelinkt aan denudatie veroorzaakt door exhumatie en de hieraan geassocieerde erosie langs de breukzones. Waarschijnlijk zal denudatie t.g.v. het openen van het Mergui Bekken en het Sumatra Bekken ook bijdragen geleverd hebben aan deze snelle afkoeling; (2) een min of meer thermisch stabiele fase tot ongeveer 13-12 Ma. Deze wordt gerelateerd aan de periode tussen het einde van de exhumatie van de breukzone en het begin van de opening van de Andaman Zee. Tijdens deze periode deden zich immers geen sterke tektonische veranderingen voor; (3) een hernieuwde fase van versnelde afkoeling die nog steeds aan de gang is. Deze versnelde afkoeling is waarschijnlijk het gevolg van de opening van de Andaman Zee, gezien dit een base-level drop met zich meebrengt, wa op zijn beurt resulteert in een denudatie in het schiereiland van Thailand.

Behalve staal TN-36 (261 Ma), leverden de TN-stalen een Mesozoïsche ouderdom op (112-187 Ma) die het opbreken van Gondwana weerspiegelen, ook deze ouderdommen werden geïnterpreteerd als afkoelingsouderdommen. De afkoelingsgeschiedenis van de TN-stalen kan worden samengevat in de volgende 3 fasen: (1) een eerste afkoelingsfase doorheen het grootste deel van het Paleozoïcum en een deel van het Mesozoïcum (tot ongeveer 200-120 Ma). Deze afkoelingsfase wordt gezien als het gevolg van denudatie die waarschijnlijk het gecombineerde effect is van enerzijds een trage erosie en exhumatie na de Malagasië orogenese en anderzijds erosie geassocieerd aan de base-level drop die gepaard ging met de opening van de Indische Oceaan; (2) een fase van thermische stabiliteit doorheen de rest van het Mesozoïcum en het Paleogeen (tot ongeveer 45-20 Ma). Deze fase komt overeen met de ontwikkeling van de Oostelijk passieve continentale Indische rand na de opening van de Indische Oceaan; (3) een fase van versnelde afkoeling van 45-20 Ma tot het heden. Deze versnelde afkoeling kan geïnterpreteerd worden als een uplift fase, gerelateerd aan de tektonische gevolgen van de India-Eurazië botsing die waarschijnlijk resulteerde in een versnelde denudatie van de passieve Indische rand.
Appendix A: Photographs of the ST-samples
Appendix B: Geological map of the entire Tamil Nadu and Pondicherry (from the Geological Survey of India)
Appendix C: The southern sheet of the geological map of Thailand (from the department of mineralogical resources)
### Legend:

**Sedimentary and metamorphic rocks**

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<tr>
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</thead>
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<tr>
<td>Quaternary</td>
<td>Alluvial and coastal deposits: gravel, sand, silt, and clay</td>
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<td>Terrace deposits: gravel, sand, silt, and coastal sediments</td>
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<tr>
<td>Triassic</td>
<td>Red conglomerate, sandstone, siltstone, and mudstone: shale and coal beds: siltstone and red shale</td>
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<tr>
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<tr>
<td>Carboniferous</td>
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### Igneous rocks

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<tr>
<td>Triassic</td>
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<td>Permian</td>
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<tr>
<td>Carboniferous</td>
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