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Two-dimensional thermal modelling of the early tectonometamorphic evolution in central Himalaya

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Abstract

The Higher Himalayan Crystallines is a major structural feature of the Himalayan belt characterized by a typical collisional and metamorphic evolution from Eocene to Miocene. We use two-dimensional thermal modelling to examine the velocity and the duration of the two main phases of metamorphism in the Higher Himalayan Crystallines. Our calculations suggest that the Eohimalayan metamorphism is not a simple consequence of the burial of the Indian plate below the upper Himalayan wedge. The M1 metamorphic conditions (700 °C, 1 GPa) recorded in the central Himalaya between 37 and 32 Ma corresponds to a period of thermal relaxation of the Higher Himalayan Crystallines. During this period, part of the Himalayan convergence had taken place in the Himalayan wedge along south-vergent thrusting structures, located north and south of the future Main Central Thrust (MCT) zone. Relatively high temperature (~600 °C) recorded in the MCT between 23 and 18 Ma (Neohimalayan metamorphism) may be explained by high heat production rates, ~3 μW/m^3, in the upper Himalayan crust and high exhumation rates, greater than 3 mm/year, of the Higher Himalayan Crystalline rocks. By analogy with modern accretionary wedges, we propose that the Eohimalayan phase corresponds to the period of the formation of the Himalayan wedge through several internal and external thrusting. During the subsequent Neohimalayan phase, the Himalayan wedge reached a thermal equilibrium, leading to the localisation of the Himalayan convergence only along the MCT zone. © 2002 Elsevier Science Ltd. All rights reserved.

1. Introduction

The thermal evolution of an orogenic zone is dependent on five factors: thermal conductivity, radiogenic heat production, mantle heat flow, tectonics and erosion. The Himalayan belt prob-

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ably illustrates the relationships between these factors than any other collisional zones (England et al., 1992). The effects of these factors have been discussed in detail in several papers (England and Thompson, 1984; Mercier et al., 1991; England et al., 1992; Ruppel and Hodges, 1994; Huerta et al., 1998; Jamieson et al., 1996; Henry et al., 1997; Harrison et al., 1998). Although one-dimensional (1-D) thermal models of a collision zone reproduce the heating recorded by metamorphic units during their exhumation (England and Thompson, 1984; Mercier et al., 1991), they fail to explain inverted metamorphism and the effect of thrusting and underthrusting during the metamorphic evolution as observed in the Himalayan belt. In contrast, 2-D thermal conductive models do not reproduce the high temperatures (700 °C) along deep-seated thrust faults (e.g. Royden, 1993; Henry et al., 1997). This problem may be partially explained by the high temperature before movement along a major thrust fault (England et al., 1992; Harrison et al., 1999), or significant heat production by radionuclides in the crust, and frictional heating along the fault (Ruppel and Hodges, 1994; Jamieson et al., 1996; Harrison et al., 1998).

A variety of models are proposed to explain the relationships of inverted metamorphism and/or anatexis to large-scale faulting during the Himalayan metamorphism (e.g. England et al., 1992; Harrison et al., 1999; Hodges, 2000 for reviews). However, the precise timing and duration of thermal processes have never been thoroughly modelled in Himalaya. In this paper we discuss the cause of the episodic tectonic and metamorphic events during continuous subduction of the Indian plate since the onset of India–Asia collision. Our main purpose is to explain the pressure and temperature of the regional metamorphism using simple 2-D thermal simulation. This work focuses on the central part of Himalaya where metamorphism, geometry and the timing of the tectonometamorphic events are well constrained during the Oligocene and Miocene (Caby et al., 1983; Le Fort et al., 1986; 1997; Pêcher, 1989; Inger and Harris, 1992; MacFarlane, 1995; 1999; Vannay and Hodges, 1996; Coleman and Hodges, 1998; Guillot, 1999; Hodges, 2000; Catlos et al., 2001).

2. Geological background

On the Himalayan side of the India-Asia collision zone, the orogenic evolution is divided into three major periods (Le Fort, 1975; Burg et al., 1987; de Sigoyer et al., 2000; Hodges, 2000). The Protohimalayan phase (Cretaceous-Early Eocene) is defined to be the tectonometamorphic evolution prior to the India-Asia collision. This phase is preserved only in the Transhimalaya, Indus-Tsangpo suture zone and partly in the Tethyan sedimentary cover (Fig. 1). It is not recorded in central Himalaya because it is located far from the suture zone (Fig. 1). This event ended with the subduction of the Indian margin below the south Asian continent between 55 and 50 Ma (de Sigoyer et al., 2000).

The Eohimalayan phase (Middle Eocene–Late Oligocene) represents the main India–Asia collision event and subsequent thickening of the Indian plate prior to the initiation of north–south extension (Treloar, 1997; Guillot et al., 1997; de Sigoyer et al., 2000; Hodges, 2000). Prior to the collision, the northern Indian margin was composed of a continent crust (25 km in thickness) covered by a 10 km thick Cambrian to Eocene Tethyan shelf sediments (Le Fort, 1989). The internal Himalayan wedge was progressively thickened by tectonic imbrication of the upper Indian crust along the Main Himalayan Thrust (MHT) (Figs. 2 and 3). The tectonic thickening resulted in the burial of the Middle Proterozoic to Paleozoic clastic rocks (Zhao et al., 1993; Le

Fort, 1989; Vannay and Hodges, 1996; Parrish and Hodges, 1996), and the development of the Eohimalayan metamorphism (M1), which is characterised by high-pressure granulitic to upper amphibolitic facies conditions. M1 metamorphic rocks are preserved all along the Himalayan belt in the North Himalayan Crystalline massifs and in the Higher Himalayan Crystallines (HHC) (e.g. Guillot et al., 1999 and references therein). Pêcher and Le Fort et al. (1986) noticed in central Himalaya that this prograde metamorphism is associated with penetrative compressive shear structures, preserved only in the HHC. In the central part of the belt, from Zanskar to the Everest region, Eohimalayan metamorphism started at about 40 Ma with garnet growth and ended before 25 Ma (e.g. Guillot et al., 1999 and references therein)

The Neohimalayan phase (Early Miocene to Present) is found throughout the Himalayan belt and is characterised by south-vergent thrusting structures that separate the HHC, the Lesser Himalaya and the Subhimalayan zones. The structurally highest and oldest of the major faults is the 10 km-thick MCT which separates the 10-km thick HHC from the Lesser Himalaya from Bhutan to Pakistan (Fig. 1). The oldest MCT structures are amphibolite-facies shear zones that developed at about 23 Ma in central Himalaya (Hubbard and Harrison, 1989; Catlos et al., 2001). At the top of the HHC, the South Tibetan detachment system (STDS) separates weakly metamorphosed Tethyan sedimentary cover of the hanging wall from gneisses of the HHC (Fig. 3). Geochronological data suggest that the STDS was active simultaneously with the MCT during the Miocene (Hodges et al., 1992; Guillot et al., 1994; Vannay and Hodges, 1996). Geological relationships and geochronological data suggest that during the Neohimalayan phase, the HHC was a wedge-shaped body bounded by the underlying MCT and the overlying STDS (Hodges et al., 1992) (Fig. 3). Synchronous activity of these fault systems expelled the HHC southward on the Lesser Himalaya (Hodges et al., 1992; Grujic et al., 1996; Grasemann et al., 1999). During this period, the MCT is considered to be the equivalent of the present day Main Himalayan

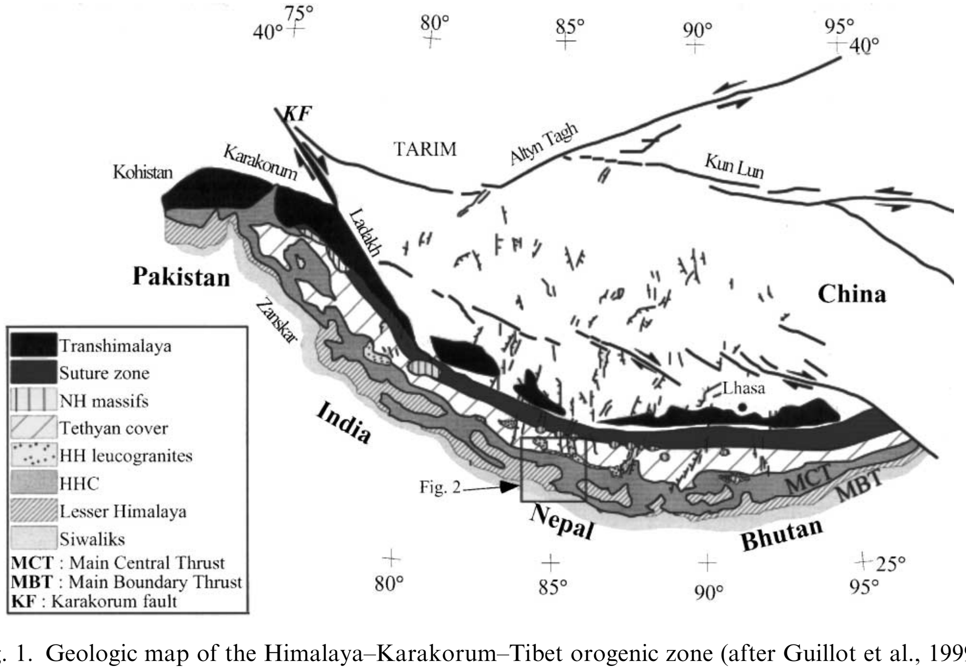

Fig. 1. Geologic map of the Himalaya–Karakorum–Tibet orogenic zone (after Guillot et al., 1999).

Thrust (Yin et al., 1999). Late development of ramp geometry is responsible for the steepening of the MCT close to the surface (MacFarlane, 1993; Catlos et al., 2001) (Fig. 3).

3. P–T-t evolution in central Himalaya

The tectonic and thermal evolutions of the HHC in central Himalaya have been recently reviewed (Pêcher, 1989; England et al., 1992; Vannay and Hodges, 1996; Guillot, 1999; MacFarlane, 1999). During Eocene (onset of Eohimalayan event) the HHC was buried at a depth of c. 35 km. by continental subduction and early thrusting on the STDS. This phase was responsible for the Eohimalayan metamorphism (M1) above the MCT zone and characterised by a progressive temperature increase from about 550 °C to locally greater than 700 °C and a pressure range between 1000 and 800 MPa (Pêcher, 1989; Vannay and Hodges, 1996; Guillot, 1999) (Fig. 4). Further to the West, in Zanskar, Vance and Harris (1999) observed a more complete P-T evolution characterised by a burial of 500 MPa without significant temperature increase, followed by

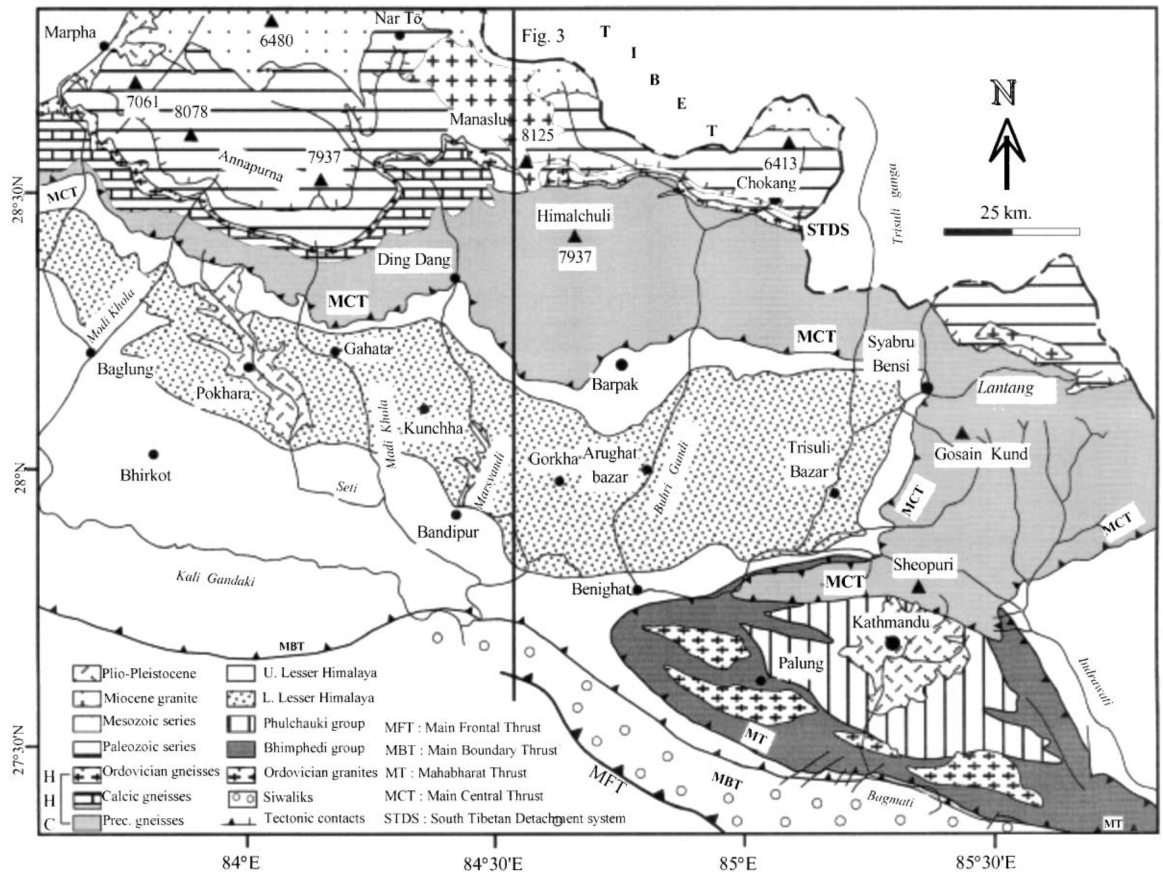

Fig. 2. Geologic map of the central Himalaya showing the location of the main structural features and the Higher Himalayan Crystallines (after Guillot, 1999).

isobaric heating of 100 °C at a constant pressure of about 1000 MPa. They estimated that the isobaric heating lasted 5 million years, between 33 and 28 Ma, before the decompression of the HHC. In the central Himalaya, the age of the Eohimalayan metamorphism is now well constrained, between 37 and 32 Ma (Inger and Harris, 1992; Mezger et al., 1992; Parrish and Hodges, 1996; Hodges et al., 1996; Vannay and Hodges, 1996; Coleman and Hodges, 1998; Godin et al., 2001; Catlos et al., 2001). There is no evidence of tectonic activity during the following period between 32 Ma and 23 Ma within the HHC or along the MCT zone.

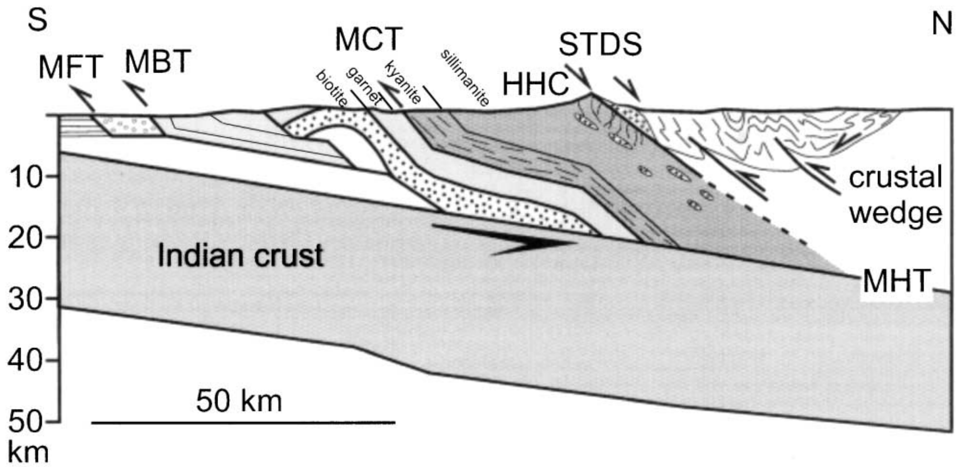

Fig. 3. Simplified geological section across the central Himalaya incorporating the seimic studies of INDEPTH I and INDEPTH II (Zhao et al., 1993; Makovsky et al., 1996). MBT, Main boundary thrust; MCT, Main central thrust; STD, South Tibetan detachment; KD, Kangmar dome.

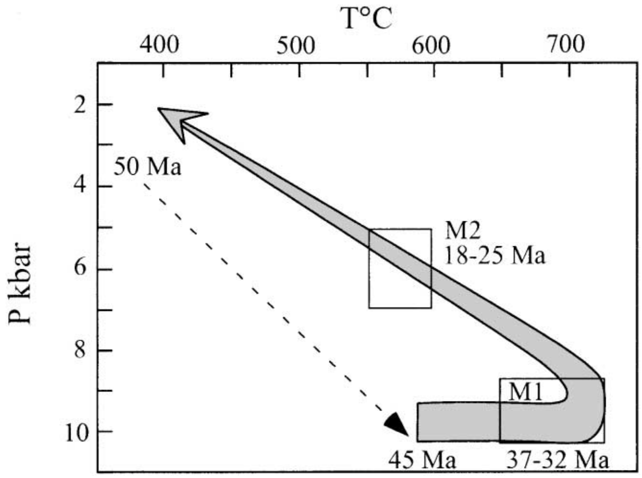

Fig. 4. P–T-t evolution of the HHC starting with its burial between 50 and 45 Ma. Thermal relaxation between 45 and 37–32 Ma leading to the Eohimalayan metamorphic conditions M1. Then, between 25 and 18 Ma, exhumation of the HHC, leading to the Neohimalayan metamorphic conditions M2.

The thrusting of the HHC over the Lesser Himalaya along the MCT zone is mainly recorded during Miocene, between 23 and 18 Ma in central Himalaya (Hubbard and Harrison, 1989; Hodges et al., 1996; Catlos et al., 2001) and is coeval with normal faulting along the STDS (Hodges et al., 1992; Guillot et al., 1994; Hodges et al., 1996; Coleman and Hodges, 1998). This phase is responsible for the development of the Neohimalayan metamorphism phase M2 (Fig. 4), well preserved in the MCT zone, and for the generation of leucogranitic plutons by widespread decompressional-anatexis (Harris and Massey, 1994; Guillot and Le Fort, 1995). P–T data suggest that M2 metamorphism is characterised by a pressure drop of about 300 MPa while temperature decreases, leading to average P–T conditions of 550–600 °C and 500–700 MPa in the MCT zone (e.g. Hodges et al., 1988; Vannay and Hodges, 1996; MacFarlane, 1995; Guillot, 1999; Kohn et al., 2001). Thrusting of the HHC over the Lesser Himalaya is associated in the footwall of the MCT zone with an upward increase in metamorphic grade (Le Fort, 1975; Pêcher and Le Fort, 1986). Numerous models have been developed to explained the relationship between this inverted metamorphism and anatexis within the Himalaya. They have been recently reviewed (Guillot, 1999; Harrison et al., 1999; Hodges, 2000) and will be not reinvestigated in this paper.

4. Questions posed by the metamorphic evolution of the HHC

Long-standing questions in Himalayan geology include the origin of the Eohimalayan metamorphic phase and the cause and timing of the Neohimalayan metamorphic phase in the HHC (e.g. Le Fort, 1989; Harrison et al., 1999; Hodges, 2000). It is widely considered that the Eohimalayan metamorphism was caused by burial of HHC beneath the northern margin of India and southernmost Tibet (Caby et al., 1983; Hodges et al., 1988; Pêcher, 1989; England et al., 1992; Coleman and Hodges, 1998; Hodges, 2000). The model, however, fails to explain high temperatures, up to 700 °C, during the burial and the occurrence of two distinct metamorphic events; Eohimalayan phase at 37–32 Ma and Neohimalayan metamorphic phase at 23–18 Ma.

Other questions include the exhumation rate of the HHC during the Neohimalayan metamorphic phase, geometrical and temporal relationships between the continuous convergence in the Himalayan belt and the discontinous record of the tectonometamorphic events during the Late Eocene–Miocene period.

5. Numerical model and boundary conditions

A 2-D numerical thermal model has been developed to study the P–T-t path registered in the HHC. It is assumed that the geometric and metamorphic conditions in the HHC during its earlier evolution were similar to present-day Himalayan evolution. This assumption is in agreement with the general opinion that the MCT is an Eocene–Miocene equivalent of the present day MHT (Nelson et al., 1996; Yin et al., 1999). For the model, we selected the geometry of the Himalayan belt, 400 and 95 km thick (compare Figs. 3 and 5). It comprises an upper fixed crustal wedge from the middle of the box to the right end. The 35 km-thick Indian crust, below the continental wedge, consists of two layers. The upper layer (10 km-thick) represents the HHC and the lower layer (25 km-thick) the middle and lower crust of the Indian craton. The crust is underthrusted below the crustal wedge along a 30°-ramp to a depth of 25 km and continues to greater depths at an angle of 10°. This geometry is similar to that observed along the INDEPTH seismic profile (Zhao et al., 1993; Makowsky et al., 1996).

The 2-D heat diffusion equation

$$\frac{dT}{dt} = \kappa\left(\frac{d^2T}{dx^2} + \frac{d^2T}{dy^2}\right) - V(x, y)\left(\frac{dT}{dx} + \frac{dT}{dy}\right) + \frac{A(x, y)}{\rho C} \quad (1)$$

is solved using a finite element software PDEase 2-D^{TM} (1998). T is temperature in °C, $A(x,y)$ is heat production in μW/m^3 (the value of A depends the rock type in the model), ρ is density in kg/m^3 (2700 kg/m^3), C is specific heat in J/kg/°K (1000 J/kg/°K), κ is diffusivity in m^2/s (10^{-6} m^2/s), $V(x,y)$ is velocity in m/s which depends on the layer and time. As for the model of Henry et al. (1997), the movement of rocks appears as a transport term (Vx,y) in the heat equation. Boundary conditions are 0 °C at the surface and constant heat flow. The basal heat flow qm is 35 mW/m^2 according to Rao et al. (1976). A value of 35 mW/m^2 for the mantle contribution to surface heat flow is unusually high for an Archean craton but is in accordance with Early Cretaceous volcanism on the northern margin of Indian continent (e.g. Le Fort, 1989). The horizontal heat flow is 0

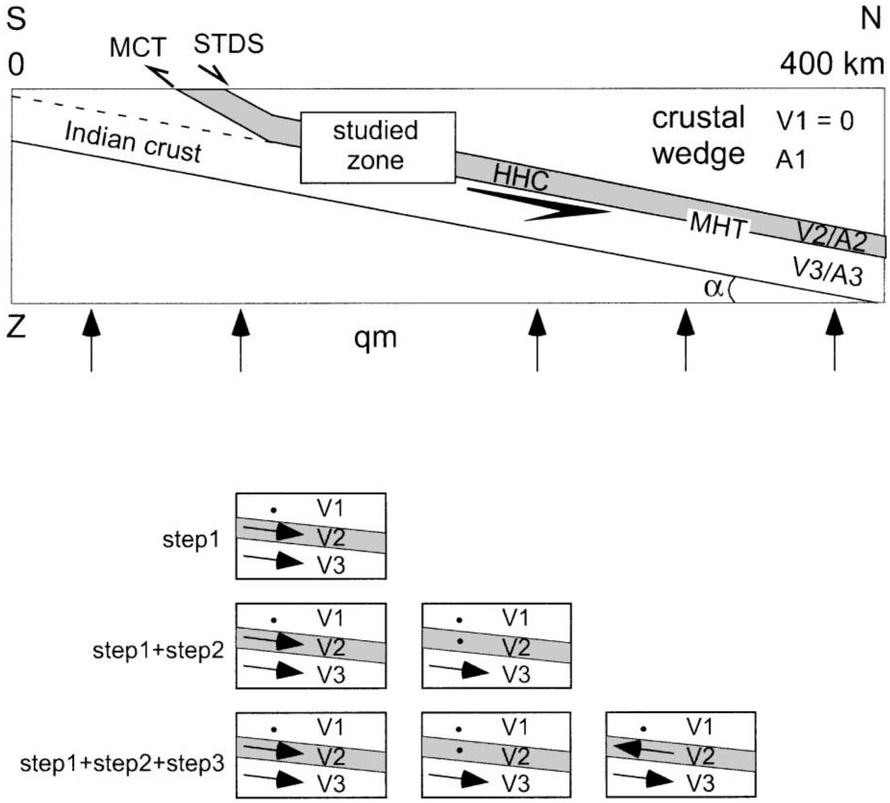

Fig. 5. Geometry, velocity field (V1, V2, V3) and thermal parameters (A1, A2, A3, qm) used for the finite element models. The insert corresponds to the studied zone. The values of each parameter are reported in the Table 1. The different steps (step 1, step 2, step 3) are reported with the velocities in the different units.

on the right-hand side of the model. The heat production A3 of the lower layer of the Indian crust (Fig. 5) is 1μW/m^3 . The crustal wedge and the upper crustal layer including the HHC are made of sediments and metamorphic rocks scraped off the Indian crust; the same value of heat production is assigned to them (A1 = A2) (Fig. 5). Three heat production values have been considered (1.5, 2.25 and 3 μW/m^3) which cover the range of expected values of sediments and lower continental crust (Van Schmus, 1989). All the values used in the experiments are in the Table 1. The geotherm in the stable Indian craton which is used as initial boundary condition on the left side of the model is given by the following equation (Turcotte and Schubert, 1982)

$$\begin{cases} T(y) = \dfrac{-A(y)}{2k} y^2 + \dfrac{qs}{k} y + T0 & \text{for } y \leqslant 35 \text{ km (in the crust)} \\ T(y) = \dfrac{qm}{k}(y - ymoho) + Tmoho & \text{for } y \geqslant 35 \text{ km (in the mantle)} \end{cases}$$

where *qm* and *qs* are the heat flow in the mantle and at the surface, respectively. Heat production values are reported in Table 1. For a thermal conductivity *k* of 3 Wm^{-1} °C^{-1}, the temperature at the Moho in the stable Indian lithosphere varies between 621 and 646 °C (Table 1). In order to test the accuracy of Pdease-2D computer programm to resolve by finite element, thermal earth science models, tests have been performed in 1-D. The differences between numerical and analytical results is less than 2 °C.

Three steps are considered successively (Fig. 5) with no internal deformation. Each step represents a step of the evolution of the Himalayan system. During the first step, the whole Indian lithosphere (upper and lower crustal layers and subcontinental lithospheric mantle) is underthrusted below the crustal wedge at a constant velocity. In this model, the STDS is thrusted southward (Vannay and Hodges, 1996; Coleman and Hodges, 1998; Godin et al., 2001). Two steady state thermal fields are computed with the Eq. (1), assuming velocities of underthrusting of 10 and 20 mm/year, similar to the present-day velocities in Himalaya (Lyon-Caen and Molnar,

Table 1
Thermal and velocity parameters used in this study. The different types refer to the experiments with different heat productions. The three steps (1, 2, 3) correspond to the burial, the thermal relaxation and the exhumation of the HHC respectively

Type	qm (mW/m^2)	qs (mW/m^2)	A1 (μW/m^3)	A2 (μW/m^3)	A3 (μW/m^3)	T stable Moho (°C)	V3 step 1–2-3 (mm/year)	V2 step 1 (mm/year)	V2 step 2 (mm/year)	V2 step 3 (mm/year)
Type 1	35	75	1.5	1.5	1	621	10	10	0	−6
Type 2	35	82.5	2.25	2.25	1	633	10	10	0	−6
Type 3.1	35	90	3.	3	1	646	10	10	0	−6
Type 3.2	35	90	3	3	1	646	10	10	0	−15
Type 4	35	70	1.5	1.5	1	621	20	20	0	−6
Type 5	35	82.5	2.25	2.25	1	633	20	20	0	−6
Type 6	35	90	3.	3	1	646	20	20	0	−6

1985; Jackson and Bilham, 1984; Pandey et al., 1995). These thermal fields are used as initial conditions for the second step.

During the second step, the HHC (upper crustal layer) is decoupled along the MHT from the rest of the Indian crust, in order to simulate the period of heating recorded by the HHC during the Eohimalayan phase. In this condition, the velocity of the HHC is set to 0. During this simulation, the rest of the Indian lithosphere below the HHC continues to be underthrusted at the same velocity as in the first step. The duration of the second step is 10 million years, corresponding to the time delay necessary for the heating of the HHC between 45 and 35 Ma. The resultant thermal field after 10 million years is used as the initial conditions for step 3.

During the third step, which simulates the exhumation of the HHC in order to reach the Neohimalayan conditions, the velocity below the HHC is constant. The HHC is exhumed from a depth of 35 to 25 km simultaneously along the MCT and the STDS. The duration of the third step is also 10 million years corresponding to the time difference between the Eohimalayan and Neohimalayan metamorphic events. It leads to a vertical velocity of 1 mm/year, and 6 mm/year. along the STDS. The parameters used in the different types, are summarized in the Table 1.

The geometry and the initial conditions for heat production employed in our modelling are very similar to the numerical experiments of Royden (1993), Henry et al. (1997), Harrison et al. (1998) and Huerta et al. (1998). Major differences include erosion parameter, underthrusting parameter, exhumation rate, and shear heating parameter.

5.1. Erosion parameter

The models by Henry et al. (1997) and Huerta et al. (1998) suggest erosion of the crustal wedge during the burial of the HHC. In our models, the upper limit of the system remains horizontal, implying no erosion of the wedge. We assume no erosion during the underthrusting of the HHC between 50 and 35 Ma because of a slow denudation rate recorded before early Miocene and the absence of significant detrital accumulation in the Bengal fan before early Miocene (Copeland and Harrison, 1990). Our assumption is supported by the preservation of Tethyan sediments in the toe of the crustal wedge, and the uplift of the Himalayan belt after the Oligocene (France-Lanord et al., 1993). Therefore, erosion became important during the extrusion of the HHC along the MCT and STDS, which corresponds to the third step of our modelling.

5.2. Underthrusting parameter

Royden (1993) and Huerta et al. (1998) proposed that the crustal wedge formed by material transfer from the footwall of the MHT to the hanging wall. The large addition of materials to the hanging wall is not supported by geological and geochemical data in the area. The main Himalayan units, particularly the HHC and the Lesser Himalaya, are clearly different on the basis of detrital zircon ages (Parrish and Hodges, 1996) and isotopic composition (France-Lanord et al., 1993; Guillot and Le Fort, 1995). Moreover, the model by Huerta et al. (1998) predicts the temperature of 600 °C after 32 million years, whereas the temperature is reached within 15 million years in Himalaya. As in the case of Henry et al. (1997), our first step corresponds to the underthrusting of a rigid plate below a crustal wedge. In other simulations, for example, Harrison et al. (1998) the underthrusting processes are not investigated.

5.3. Exhumation parameter

Other models use simple ramp geometry (e.g. England et al., 1992; Henry et al., 1997). We choose a ramp-flat geometry, as Harrison et al. (1998), because it fits the present-day geometry and produces a rapid underthrusting of the HHC followed by its rapid exhumation.

5.4. Shear heating parameter

Previous models used shear heating as a possible source of heat to explain the high temperatures recorded above the MCT zone (England et al., 1992; Henry et al., 1997; Harrison et al., 1998). However, paleopiezometry experiments (Engleder, 1993), laboratory deformation experiments (Rutter, 1997) and tectonic modelling (Treloar, 1997; Leloup et al., 1998) argue that in the case of ductile shearing, shear stress is too low (10's of MPa) to produce significant shear heating. We have tested the possibility, that shear heating is not required to explain the metamorphic conditions recorded in the HHC. We will show that realistic temperatures are reached with no shear heating.

6. Results

6.1. Burial of the HHC: step 1

In step 1, we considered the burial of the HHC below the upper continental Himalayan wedge (Fig. 4). The temperature reached at a depth of 35 km at the base of the HHC ranges between 370 °C and 550 °C (Fig. 6a). Only when heat production is high in the crustal wedge and HHC ($3\mu W/m^3$) and the burial velocity is low (10 mm/year) (type 3) is a temperature of 550 °C, compatible with the minimal temperature recorded at the beginning of the Eohimalayan metamorphism, reached (Fig. 6a and b).

It is widely considered that metamorphic temperatures increased within the HHC due to burial during Eohimalayan metamorphism, preceding slip along the MCT (Pêcher, 1989; Vannay and Hodges, 1996; Guillot, 1999). The results of step 1 shows that it is impossible to reach a temperature of ~700 °C at a depth of 35 km during continuous burial of the whole Indian plate, even if the heat production is high. Although high heat production rate (type 3) can result in ~550 °C at the beginning of the Eohimalayan metamorphism. The temperature of 700 °C was reached in Zanskar only after a period of 5 million years of isobaric thermal relaxation of the HHC (Vance and Harris, 1999). The condition is probably very similar in central Himalaya. According to our model, the time necessary for the HHC to reach a depth of 35 km is estimated between 5 million years for a velocity of 20 mm/year to 10 million years for a velocity of 10 mm/year.

Considering that the burial of the HHC started with the collision at about 50 Ma (Fig. 4), this suggests that the depth of 35 km was reached between 45 and 40 Ma, i.e. 10 to 5 million years before the time of Eohimalayan metamorphism (37–32 Ma). This strongly suggests that the HHC records a period of isobaric heating of 5 to 10 million years after its burial (Fig. 4).

Taking into account that the northward motion of the Indian plate was continuous and did not stop between 45 and 35 Ma (Patriat and Achache, 1984; Klootwijk et al., 1992), we investigated an other step (step 2) to reconcile the necessity of heating of the HHC after its burial and the continuous burial of the Indian plate along the MHT.

6.2. Burial followed by decoupling: step 2

Davy and Gillet (1986) investigated the effect of a delay of the erosion after crustal thickening. The delay leads to an increase of the maximum temperature (isobaric heating) and the maximum temperature is experienced during uplift. 1-D thermal modelling of the Eohimalayan meta-

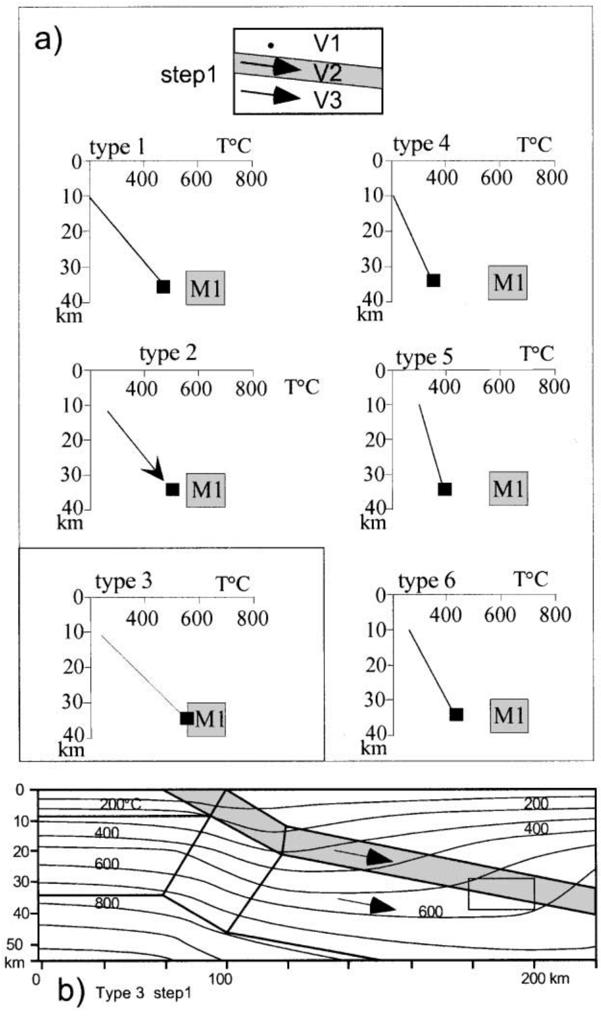

Fig. 6. (a) P–T results of the burial of the Indian plate, including the HHC, below the Himalayan orogenic wedge (step 1). The large box corresponds to the M1 metamorphic conditions recorded in the MCT zone, the black box corresponds to the calculated P–T conditions. See text for discussion. (b) Isotherms in °C computed for type 3-step1. See Table 1 for parameters. The grey area corresponds to the HHC. The rectangle box corresponds to the P–T area reported on Figs. 6–8. The type 3-step1 is characterized by the downward deflection of the isotherms related to the underthrusting of the HHC and the Indian crust below the crustal wedge.

morphism showed that a time interval of 5 to 10 million years is required to obtain a thermal relaxation of the thickened crust (e.g. England et al., 1992). These models, however, do not take into account the fact that during this interval, the northward convergence of the Indian plate still continues. In step 2, we first impose the burial of the HHC below the upper Himalayan wedge according to step 1 conditions, running to steady-state conditions. When the depth of 35 km is reached by the HHC, burial is followed by decoupling of the HHC from the Indian crust that continues to subduct below the HHC along the MHT (Fig. 5). We compute the transient thermal evolution of the HHC resting at constant depth over a period of 10 million years.

In all types, we observed a temperature increase of 70 to 180 °C in the HHC at a constant depth of 35 km (Fig. 7a). The temperature recorded during the burial, strongly controls the final tem-

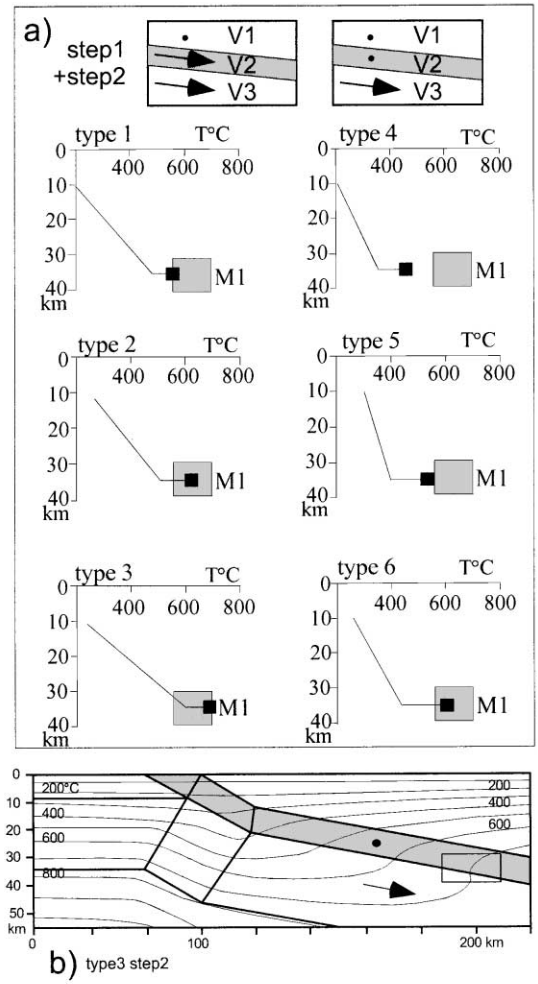

Fig. 7. (a) P–T results of the decoupling of the HHC from the continuous subducting Indian crust (step 1 + step 2). Same symbols as Fig. 6 (See text for discussion) (b) The type3-step2 corresponds to the period of resting in depth of the HHC. The continuous underthrusting of the Indian crust below the HHC is also marked by the downard deflection of the isotherm.

perature recorded during the thermal relaxation. The maximum temperature of 510 to 620 °C at the end of step 2, in types 1–2–4–5–6, is lower than the maximum temperature recorded during the Eohimalayan metamorphism (700 °C). In contrast, type 3 fits the highest temperature recorded by C at the end of the Eohimalayan metamorphism (Figs. 7a and b). In type 3, a temperature

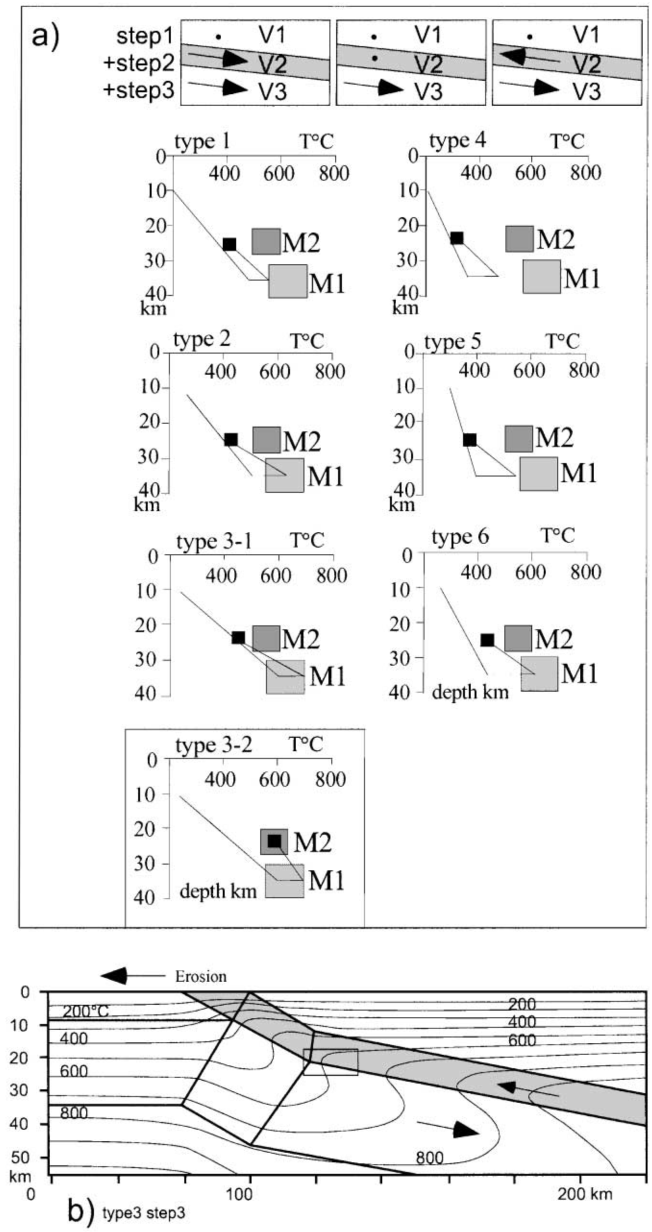

Fig. 8. (a) P–T results for the complete evolution of the HHC (step 1 + step 2 + step 3). Same symbols as Fig. 6. Only type 3–2 fit with the P–T-t evolution records by the HHC. (b) The type3-step3 corresponds to the exhumation of the HHC characterized by the strong thermal inversion of the isotherm throughout the MCT zone.

increase of 150 °C during the thermal relaxation yielded the temperature of 700 °C at the base of the HHC (Fig. 7b). The consequence of this model in term of tectonic activity throughout the Himalayan belt will be discussed further.

6.3. Exhumation of the HHC after burial and decoupling: step 3

Step 3 corresponds to the exhumation of the HHC after its burial (step 1) and its heating (step 2). We first consider that the extrusion of the HHC occurred between a shallow dip MCT (10°) at the bottom and the STDS at the top, similar to the present-day geometry. We first consider an average vertical exhumation rate of 1 mm/year for the HHC (equivalent to 6 mm/year along the STDS) deduced from the barometric (difference of 300 MPa between M1 and M2) and geochronological data (between 32–37 and 23–18 Ma). Below the MCT, the Indian plate still subducts at a rate of 10 mm/year relative to the upper Himalayan wedge, leading to a velocity of 16 mm/year accommodated along the MCT. After 13 million years of exhumation, the depth of 25 km, corresponding to the depth of the Neohimalayan event is reached at the base of the HHC.

Whatever the type considered (types 1 to 6), we observed a strong cooling of 150 to 250 °C and we cannot reproduce the temperature conditions of 550 to 600 °C recorded during M2 (Fig. 8a). This strong cooling is due to the continuous underthrusting of the Indian plate that cooled the HHC by conduction. In this condition, we have failed to reproduce the Neohimalayan P–T conditions M2.

As discussed above, numerous processes have been proposed to explain the preservation of high temperature conditions during M2, but none of them satisfactorily explains all the geological constraints (e.g. Harrison et al., 1999; Hodges, 2000). We noticed previously that the MCT activity associated with the Neohimalayan metamorphism phase was short, between 23 and 18 Ma, suggesting a rapid exhumation of the HHC. The predominant heat transfer mechanism should be advection rather than conduction. Assuming that the exhumation rate during the MCT motion was similar to the present-day uplift rate recorded by geodetic data, i.e. >3 mm/year. (Jackson and Bilham, 1994; Pandey et al., 1995), it gives a minimum rate of exhumation along the STDS of 15 mm/year. Combined with the subduction rate of 10 mm/year, the relative velocity is calculated to be 25 mm/year along the MCT.

We tested this velocity using type 3–2. A temperature of about 600 °C is reached at a depth of 25 km after 3 million years (Fig. 8a and 8b). This is consistent with the temperature recorded in the MCT zone during M2, suggesting that a high exhumation rate contributed to the preservation of the peak metamorphic condition of the HHC. It is noted that a strong thermal inversion took place in the HHC as a consequence of underthrusting of the cold Indian plate beneath the MCT and heat advection within the HHC (Fig. 8b).

7. Discussion

7.1. Heat production in the Himalayan wedge

Rao et al. (1976) first suggested that the surface heat flow is unusually high in the Indian shield probably due to a high mantle contribution and a high upper crustal internal heat generation.

They have shown that the contribution of the upper crust is about 30 mW m^{-2}. Their results were supported by later workers. Henry et al. (1997) and Huerta et al. (1998) supported high heat production rates based on high temperatures, 600 °C, of metamorphism in the HHC. Jamieson et al. (1996) suggested high heat production (3 $\mu W/m^3$) of upper continental origin for the Barrovian metamorphism. Our modelling confirms these results, suggesting a high heat production of 3 $\mu W/m^3$ within the upper Indian crust. Such a high heat production combined with a low heat production of the middle and lower crust (1 $\mu W/m^3$) leads to an average heat production of 1.6 $\mu W/m^3$ for the Indian crust, very similar to the heat production of the average continental crust, estimated at 1.75 $\mu W/m^3$ (Van Schmus, 1989). This heterogeneous heat production within the Indian crust is compatible with the concentration of high heat production (>4 $\mu W/m^3$) rocks within the HHC (Vidal et al., 1982; Scaillet et al., 1990) and the Indian shield (Roy and Rao, 2000).

7.2. The shape of the P–T-t path and shear heating

The shapes of a P–T-t path vary depending on the methods used to obtain P and T. Diffusion studies of garnet yielded a continuous P–T path, but only in the stability field of garnet (e.g. Spear, 1993). In order to obtain a complete P–T path, we used discontinuous reactions and local successive parageneses, that record successive P–T equilibria. Consequently, successive P–T boxes are drawn on a pressure-temperature diagram and the resulting P–T path is obtained by joining the boxes and the P–T path are usually considered to show the continuous thermal history of the metamorphic rocks (e.g. Spear, 1993). In Himalaya, the Eohimalayan phase occurred during Eocene while the Neohimalayan phase occurred during Miocene (e.g. Guillot et al., 1999 and Hodges, 2000 for reviews). This discontinuous metamorphic events suggest that the observed P–T-t path in the HHC do not correspond to a continuous metamorphic evolution but to discrete tectono-metamorphic events. Our modelling (type 3) with three distinct successive steps (burial, heating and exhumation) are able to reproduce the observed P–T path and the timing of these events. A way to explain that the observed P–T-t path in central Himalaya do not corresponds to a continuous event, is to imagine a decoupling of the HHC from the rest of the underthrusting Indian plate, leading to the heating of the HHC over a period of 10 million years.

7.3. Comparison between the Himalayan belt and accretionary wedges

The northwards movement of the Indian plate is considered to be continuous since 55 Ma (Patriat and Achache, 1984; Klootwijk et al., 1992) and a minimum of 40% of the total India–Asia convergence (50 mm/year) is accommodated in the Himalaya (Lyon-Caen and Molnar, 1985; Jackson and Bilham, 1994; Guillot et al., 2000). However, a maximum velocity of underthrusting of 10 mm/year is required to reproduce Barrovian metamorphic conditions (Royden, 1993; Huerta et al., 1998; Jamieson et al., 1996; Harrison et al., 1998; this study). Since the Himalayan convergence rate is at least 20 mm/year, this suggests that a large part (50%) of the convergent velocity within the Himalayan belt is accommodated by internal deformation and/or activation of several thrusting structures at the same time. Evidence for tectonic activity along the main thrust zones north and south of the MCT during the Eohimalayan period is known in central Himalaya. South of the MCT, Rai et al. (1998) showed that the Mahabharat Thrust (MT)

(Fig. 2) was active before the MCT allowing the exhumation of the Kathmandu nappe. Similarly, Vannay and Hodges (1996) and Coleman and Hodges (1998) also showed that the STDS located north of the MCT was a thrusting structure during the Eohimalayan event. Guillot et al. (1995) also suggested that the 50 km wide Himalayan nappe preserved in the Indus suture zone, north of the central Himalaya, would have extended towards the south for an additional minimum of 70 km before 30 Ma, i.e., during the Eohimalayan phase. The Kangmar Thrust active before the MCT is a good candidate to explain the southward extension of the north Himalayan nappe (e.g. Burg et al., 1987). Another south-vergent thrust has been discoverved by Le Fort and Guillot (1998) within the Tethyan sedimentary cover also in central Nepal. Along the Ngaru Thrust, active before the STDS, the upper Liassic sediments thrust over the Permo–Triassic sediments. Finally, Yin et al. (1999) demonstrated that the Gangdese Thrust located in South Tibet was strongly active before the MCT, between 30 and 24 Ma.

In the absence of precise geochronological data on the different thrusting structures, except for the Gangdese Thrust, it is impossible to evaluate whether these thrusting structures were simultaneously active during the Eohimalayan phase. The existence of several active thrusting structures during the formation of the Himalayan wedge is very similar to the tectonic processes active in oceanic or continental accretionary wedges (Platt, 1986; Von Huene and Scholl, 1991; Chalaron et al., 1995; Cattin and Avouac, 2000). During the period of building of an accretionary wedge, the thickening rate is superior to the erosion rate which is the case for the Eohimalayan period (Copeland and Harrison, 1990; Stow et al., 1990). Then, when the accretionary wedge reaches the mechanical equilibrium, the thickening is compensated by erosion (Dahlen et al., 1984) which is also the case for the Himalaya during the Neohimalayan period during when the rate of exhumation of the HHC is compensated by the rate of erosion and sedimentation in the Bengal fan (Copeland and Harrison, 1990; France-Lanord et al., 1993).

According to our thermal simulations and the tectonic arguments presented above, we propose the following scenario for the early evolution of the Himalayan wedge. During the early Eocene, the progressive building of the internal Himalayan wedge by south-vergent thrusts leads to the burial of the HHC down to 35–40 km depth (Fig. 9a). From Upper Eocene to Late Oligocene, the HHC was tectonically inactive, resting at a depth of about 35 ± 5 km (Fig. 9b). The vertical accumulation of high heat producing upper crust in the internal Himalayan wedge allows the heating of the HHC (Eohimalayan metamorphism). During this time interval, the Himalayan convergence was accomodated by external thrusts such as the MT and internal thrusts (STDT, KT, GT) and by underthrusting of the Indian plate along the MHT (Fig. 9b). Then, during the Early Miocene (Neohimalayan phase), the HHC was sufficiently thermally weakened to be exhumed by ductile extrusion (Grujic et al., 1996; Grasemann et al., 1999) (Fig. 9c). During the Neohimalayan phase, the extrusion of the HHC was rapid with exhumation rate of a minimum of 3 mm/year and a velocity of a minimum velocity of 20 mm/year along the MCT.

A high exhumation rate during a short period of about 3 million years in our model is compatible with the short duration of the MCT activity during the Miocene. This result suggests that the time necessary to reach the Neohimalayan metamorphic conditions after the Eohimalayan phase is a third of time recorded by geochronological data between M1 and M2. These results are strengthened by the absence of continuous geochronological recordings in central Himalaya during Oligocene (Guillot, 1999; Godin et al., 2001).

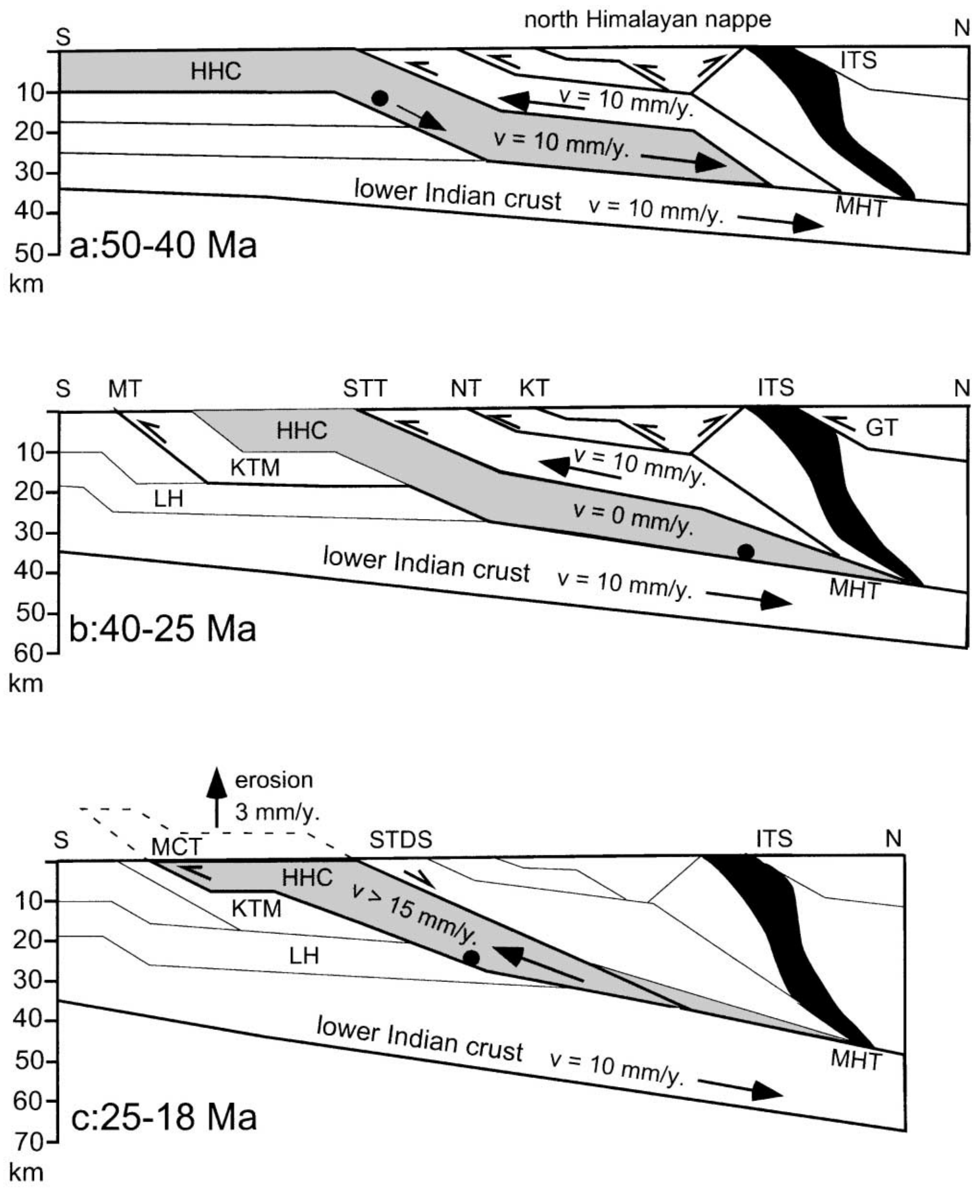

Fig. 9. Schematic evolution of the early evolution of the Himalayan thrust system. The active faults are shown as bold black lines. Step 1 (a): early Eocene building of the internal Himalayan wedge by underthrusting of the HHC at 10 mm/year and activation of south-vergent thrusting zones. Step 2 (b): upper Eocene to Oligocene period of activation of internal and external south-vergent thrusting zones leading to the exhumation of the frontal Kathmandu nappe (KTM) along the Mahabharat thrust (MT). In the internal zone, the Gangdese thrust (GT), Kangmar thrust (KT), the Ngaru Thrust and the South Tibetan Thrust lead to the thickening of the internal Himalayan wedge. During this period, the HHC is blocked at a maximum depth of 35–40 km and its thermally relaxed. Step 3 (c): early Miocene extrusion of the thermally weakened HHC between the STDS and the MCT. During this period the Himalayan convergence is completly accomodated along the MCT with a total velocity of 25 mm/year. All along the early Himalayan evolution, the lower Indian crust is continuously underthrusted below the MHT as observed in present-day accretionary wedges.

8. Conclusions

We have shown that because of the continuous subduction of the Indian plate since the beginning of the India–Asia collision, the best way to reproduce the P–T conditions recorded during the Eohimalayan metamorphism is to impose a decoupling of the HHC from the subducting Indian plate after its burial at a depth of about 35 ± 5 km. The decoupling allows the HHC both to remain at constant depth and to be heated up. From our calculations, 10 million years is necessary to reach a temperature of 700 °C in the HHC, while the underthrusted Indian plate remained at a relatively low temperature (<600 °C). During this period, the Himalayan convergence was probably accommodated along other thrusting structures located north and south of the High Himalayan Crystallines. By analogy with the functioning of accretionary wedge, the Eohimalayan phase corresponds to the period of building of the Himalayan wedge above the major intra-crustal decollement, the MHT.

To preserve the high temperature (600 °C) during the exhumation of the HHC along the MCT and the STDS a relatively high vertical rate (~3 mm/year), similar to the present-day uptift rate, is required. This suggests that the MCT and STDS were probably active over a short time (<10 million years) during the Earlier Miocene, compatible with the short period of leucogranite emplacement and geochronological records of MCT activity between 23 and 18 Ma (Hubbard and Harrison, 1989; Guillot et al., 1994; Hodges et al., 1996; Vannay and Hodges, 1996; Guillot, 1999; Godin et al., 2001). With these velocity conditions, shear heating or other heat supplies are not required to explain the observed P–T path.

We propose that since Miocene time, the Himalayan wedge reached mechanical and thermal equilibrium, where erosion, thickening and exhumation processes are balanced. The proposed analogy between the Himalayan belt and accretionary wedge explain the apparent contradiction between continuous convergent processes at the plate scale and discontinuous processes at the crustal scale. We propose that the underthrusting of the Indian crust could be a continuous processes with an average velocity of about 10 mm/year, while the tectonic and thermal activities of the metamorphic units such as the HHC was probably a discontinuous processes with long periods of no tectonic activity (>10 million years) followed by rapid (cm/year) and short tectonic activity (<5 million years).

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