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SUBSIDENCE ET RÉGIME THERMIQUE DES BASSINS INTRACRATONIQUES ET DES MARGES CONTINENTALES PASSIVES

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Cette thèse a été réalisée
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RÉSUMÉ

L'objectif de cette thèse est d'étudier les effets thermiques et les mécanismes de subsidence des bassins sédimentaires. Deux problèmes ont été étudiés: 1) l'effet d'une combinaison de contraction thermique et de changements de phase sur la subsidence des bassins intracratoniques, 2) l'étude du paléo-régime thermique des marges continentales passives de l'est du Canada.

Récemment on a accepté l'hypothèse, que la contraction thermique de la lithosphère inialement chaude est la cause principale de la subsidence tectonique des bassins sédimentaires. Le problème majeur du mécanisme de la contraction thermique est qu'il prédit que la subsidence commence rapidement et ralentit avec le temps (c.a.d. en fonction $t^{1/2}$). Ce modèle ne satisfait pas les données de la subsidence des bassins sédimentaires du Michigan, d'Illinois, de la Baie d'Hudson et du Williston.

Un mécanisme de subsidence qui combine l'effet de la contraction thermique et des changements de phase est proposé dans cette thèse comme hypothèse de l'évolution des bassins sédimentaires intracratoniques. Dans ce modèle, la contraction thermique et les changements de phase sont la conséquence du changement de la condition thermique à la base de la lithosphère.

La subsidence tectonique est déterminée pour deux conditions aux limites différentes: 1) diminution brusque de la température à la base de la lithosphère et 2) diminution brusque du flux de chaleur à la base de lithosphère. La subsidence tectonique calculée est obtenue par superposition de la subsidence thermique et de la subsidence due aux changements de phase. Elle est amplifiée par: 1) les réajustements isostasiques, et 2) la migration additionnelle du changement de phase causée par la sédimentation. Le modèle tient également compte des variations du niveau de la mer.

Les calculs ont démontré que:

1) La subsidence due aux changements de phase est retardée par rapport à la subsidence thermique. Ce retard explique une accélération de subsidence qui s'était produite dans le stade initial d'évolution des bassins du Michigan et du Willison.

2) La durée de la subsidence dépend des conditions aux limites à la base de la lithosphère. Pour une variation du flux de chaleur à la base de la lithosphère, le retour à l'équilibre thermique est 4 fois plus lent que dans le cas du changement de température à la base de la lithosphère.

3) La différence de durée de subsidence entre les bassins du Michigan et du Williston pourrait s'expliquer par différentes conditions à la base de la lithosphère qui reflètent différents mécanismes d'interaction entre les plumes mantéliques et la lithosphère.

Si la subsidence tectonique est causée seulement par la contraction thermique, le flux de chaleur en excès (par rapport au flux en équilibre) peut être directement déterminé à partir du taux de la subsidence tectonique. Pour les marges de Nouvelle-Écosse et de la mer du Labrador, l'excès du flux estimé était de l'ordre de 28-56 $mW.m^{-2}$ au début du "drifting" et sa valeur actuelle est de 7-14 $mW.m^{-2}$ (en fonction des conditions aux limites). L'analyse des résultats montre une évolution distincte des marges de la mer du Labrador et celles du nord-est de Terre-Neuve. L'excès du flux de chaleur des marges du nord-est de Terre-Neuve a été extrêmement élevé, de l'ordre de 100-200 $mW.m^{-2}$ (en fonction des conditions aux limites) au début du "drifting". Ce phénomène est causé, soit par l'extension qui a continué après la séparation des continents, soit par la déformation ductile de la croûte inférieure et/ou du manteau supérieur.

ABSTRACT

The purpose of the study is to examine thermal effects on the mechanism of subsidence of sedimentary basins and continental margins. Two main problems were addressed: 1) the combined effects of the thermal contraction and phase changes on the subsidence in intracratonic sedimentary bassins, and 2) the determination of the paleo heat-flow in excess of the steady state along eastern Canada's passive margins.

Thermal contraction of an initially hot lithosphere is widely accepted as the main cause of tectonic subsidence in sedimentary basins. One of the problems with the thermal contraction mechanism is that it predicts that subsidence begins rapidly and slows down with time (i.e. $t^{1/2}$ behavior). This does not account well for the subsidence data from the Michigan, the Illinois, the Hudson's Bay, and the Williston basins.

A mechanism of tectonic subsidence is investigated which combines the thermal contraction and the cooling of the lithosphere with the effect of a phase transformation, following a change in thermal conditions at the lithosphere-asthenosphere boundary (LAB). The tectonic subsidence is determined for two different boundary conditions: 1) a sudden drop in temperature at the LAB, or 2) a sudden heat flow drop at the LAB. The calculated tectonic subsidence is obtained by the superposition of the thermal and phase change subsidence. It is amplified by: 1) isostatic adjustments, and 2) additional migration of the phase boundary due to the weight of the sediment load. The effect of sea-level variations is also included in the model.

The calculations show that:

1) The subsidence induced by the phase change lags behing the thermal subsidence. This delay of the phase change subsidence may explain some acceleration during the early stage of evolution of the Michigan and the Williston basins.

2) The duration of the subsidence depends very strongly on the boundary condition at the LAB. For a sudden drop in heat flow at the lithosphere-asthenosphere boundary, the return to equilibrium is 4 times slower than for a sudden drop in temperature.

3) The difference in duration of the subsidence between the Michigan and the Williston basins may be related to the different boundary conditions at the LAB (temperature drop for the Michigan and heat flow drop for the Williston) which reflect the different mechanisms of thermal interaction between the plumes and the lithosphere.

If tectonic subsidence is caused by the thermal contraction of the lithosphere only, the heat-flow in excess of background heat-flow can be estimated directly from tectonic subsidence rate.

For the Nova-Scotia and the Labrador Sea margins, the estimation shows that the heat-flow in excess decreased markedly from 28-56 $mW.m^{-2}$, immediately after rifting, to a present value between 7 and 14 $mW.m^{-2}$ (depending on boundary) conditions. The analysis suggests a distinctive evolution of the Labrador Sea and of the northeastern Newfoundland margins. The excess of heat-flow in northeastern Newfoundland margins immediately after rifting was on the order 100-200 $mW.m^{-2}$ (depending on boundary conditions). This suggests that the subsidence in these margins is also affected by continuing extension after break-up of the continents and/or ductile deformation in the lower crust or upper mantle.

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INTRODUCTION

L'étude des bassins sédimentaires représente une partie importante dans l'études de l'évolution des continents. L'objectif principal des recherches actuelles est de déterminer les processus qui ont causé la formation des bassins sédimentaires. Il est important de souligner qu'un intérêt particulier s'attache aux bassins sédimentaires à cause de leur richesse en hydrocarbures.

Les hydrocarbures sont formés par une maturation thermique de sédiments riches en matière organique durant l'enfouissement. Plusieurs facteurs contribuent au métamorphisme organique, mais le processus dépend d'abord du régime thermique pendant l'enfouissement (Tissot *et al.* 1974). Lopatin (1971) a conclu expérimentalement que le taux de réaction d'altération thermique de la matière organique double pour chaque 10°C d'augmentation de température. Le degré du métamorphisme organique est un indicateur de l'histoire thermique. Par conséquent, la connaissance de l'histoire thermique permet d'estimer la possibilité de la conversion des kérogènes en hydrocarbures.

Un bassin sédimentaire a été défini par Bally (1982) comme une région qui a été affectée par la subsidence et contient plus de 1 km de sédiments préservés sous forme plus ou moins cohérente. Différents types de bassins sédimentaires existent dans les continents. Les bassins se distinguent par leur morphologie, leur structure, la durée de leur évolution, l'épaisseur de sédiments et leur position à l'intérieur des plaques tectoniques . Cette diversité suggère que différents phénomènes tectoniques ont causé la formation des bassins sédimentaires.

De façon générale, on peut distinguer quatre principaux types de bassins sédimentaires continentaux:

1) Les bassins intracratoniques.

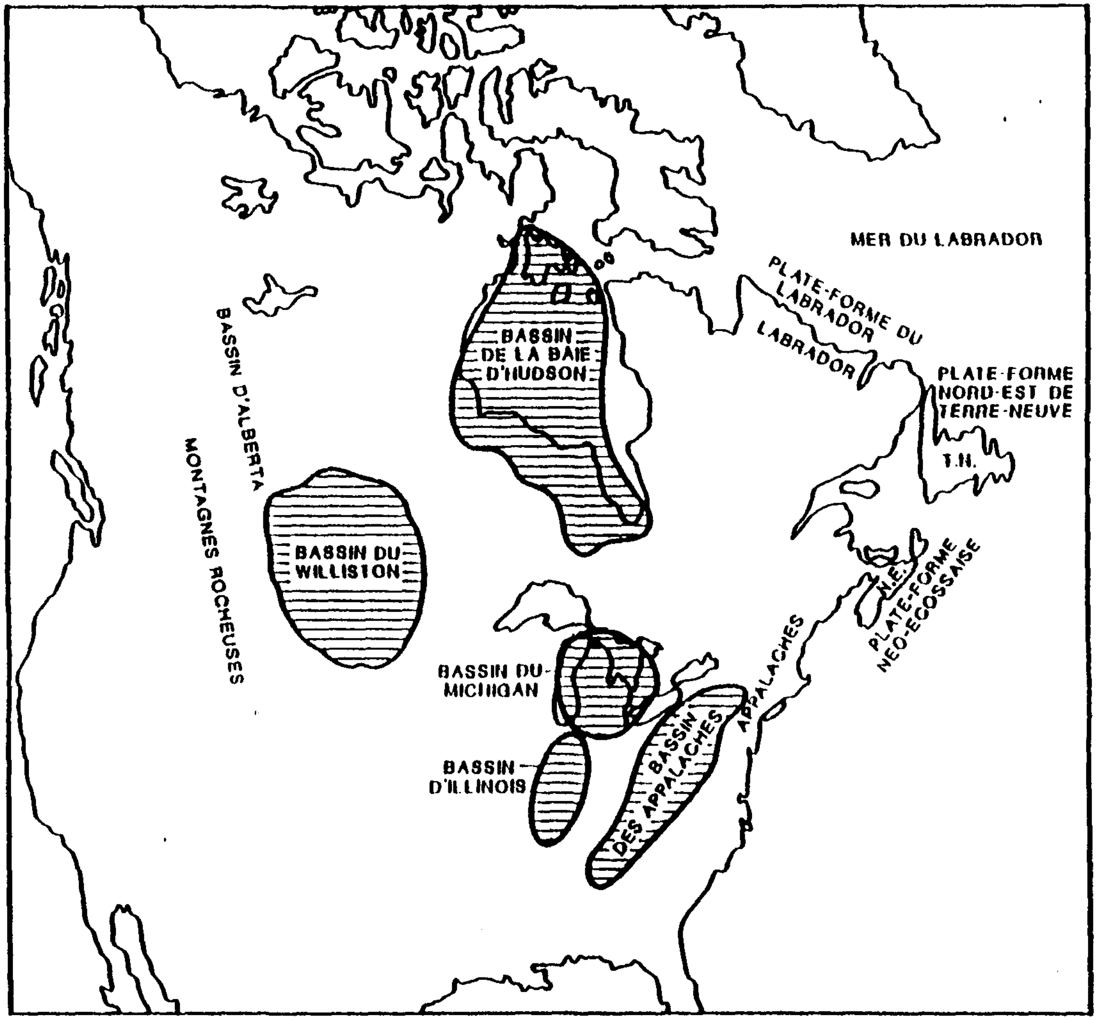
2) Les marges continentales passives.

3) Les bassins de rifts avortés et les sillons aulacogènes.

4) Les bassins d'avant-pays ("foreland basins").

Un bassin intracratonique est un bassin sédimentaire situé dans lithosphère continentale rigide pré-Mésozoïque et dont la formation n'est pas associée au développement de méga-sutures (Bally et Snelson, 1980). Ce sont des bassins situés à l'intérieur des continents dont la formation n'est pas toujours liée aux phénomènes associés à la tectonique de plaques. L'épaisseur moyenne de sédiments des bassins intracratoniques est de l'ordre de 4-5 km. Dans le continent Nord-Américain, les bassins sédimentaires du Michigan, du Williston, de la Baie d'Hudson et d'Illinois (voir fig.I-1 pour la localisation) appartiennent à ce type de bassin.

Les bassins de marges continentales passives sont situés sur la lithosphère continentale et leur développement est relié à la cassure d'un continent et à la formation d'un bassin océanique. Les marges continentales passives bordent les océans Atlantique, Indien et Arctique. Elles se caractérisent par une épaisseur substantielle de sédiments excédant pour certaines marges anciennes 15 km. Les marges continentales passives se sont formées sur les bordures des continents actuels après la séparation des plaques. Leur développement commence avec la naissance d'un rift continental et continue durant les mouvements divergents des plaques. Par exemple, les marges continentales passives de l'est du Canada (voir la fig.I-1 pour la localisation) se sont formées après la fragmentation de la Pangée et la dérive des continents entourant l'Atlantique.



Localisation des bassins et des marges étudiés

Un rift intracontinental avorté avant rupture du continent donnera naissance à un bassin extensionnel (bassin de "type stretching"). Par exemple, la formation du bassin de la mer du Nord a été causée par un rift avorté lié au développement de l'Atlantique Nord (Sclater et Christie, 1980)

Les rifts intracontinentaux contiennent souvent des branches multiples (points triples) dont certaines ne s'ouvrent pas, donnant naissance à des sillons subsidents au milieu du continent (Burke, 1977). Ces bassins, ayant comme structure des grands grabens allongés, forment des sillons aulacogènes. L'épaisseur des sédiments dans ce type de bassin excède largement 10 km. Les aulocogènes de part et d'autre de l'Atlantique sud, se sont formés pendant la séparation de l'Amérique du Sud et de l'Afrique; le sillon aulacogène de Parentis dans le bassin d'Aquitaine date de l'époque de l'ouverture de l'Atlantique Nord. Au Québec, les sillons aulacogènes de l'Outaouais et du Saguenay se sont formés durant l'ouverture de l'Océan Iapetus.

Les bassins d'avant-pays sont formés dans un milieu soumis à des forces de compression durant une orogénèse causée par le convergence de plaques. Leur formation est le résultat de la flexure de la lithosphère sous la charge des nappes de charriage à la surface. Les bassins d'Alberta (au Canada) et celui des Appalaches sont des exemples de ce type de bassin. Le bassin d'Alberta dont la formation a affecté la fin de l'évolution du bassin du Williston, un des objets de cette recherche, s'est formé suite à l'enfoncement de la lithosphère sous le poids des Montagnes Rocheuses (Beaumont, 1981).

D'une manière générale, la formation des bassins sédimentaires, à l'exception des bassins intracratoniques, est étroitement liée aux phénomènes associés à la tectonique des plaques. Les mécanismes de subsidence des bassins sédimentaires intracratoniques sont mal connus et soulèvent des débats considérables.

Le terme "subsidence tectonique" englobe la partie de la subsidence causée par des phénomènes tectoniques actifs qui forment la structure des bassins et des marges continentales passives. Les sédiments s'accumulent dans la dépression formée tectoniquement et leur poids constitue une charge locale mise en place sur la lithosphère. L'ajustement isostatique (ou la flexure) dû à cette charge amplifiera la subsidence. Cette amplification est la réponse passive de la lithosphère à une charge sédimentaire à la surface. Mais la sédimentation n'est pas la cause principale de la subsidence, car elle s'arrête sitôt le bassin rempli.

La subsidence tectonique peut être déterminée à partir de l'épaisseur et de l'âge des sédiments à l'aide de la procédure de "backstripping" qui enlève les effets de l'amplification isostatique (ou de la flexure), des variations du niveau de la mer et de la compaction des sédiments.

Les mécanismes de subsidence tectonique suggérés traditionnellement pour les bassins intracratoniques peuvent être classés en deux catégories majeures:
1) La contraction thermique d'une lithosphère initialement chaude (Sleep, 1971; Sleep et Snell, 1976; McKenzie, 1978).
2) Les changements de phase causés par la transformation métamorphique d'un faciès moins dense en un faciès ayant une densité plus élevée (Kennedy, 1959; Lovering, 1958; Middleton, 1980; Favley, 1974; Spohn et Neugenbauer, 1978; Neugenbauer and Spohn, 1978, 1982).

L'hypothèse suivant laquelle la contraction thermique de la lithosphère initialement chaude est la cause principale de la subsidence tectonique est acceptée par une majorité de chercheurs. La contraction thermique nécessite une lithosphère initialement chaude. Différentes causes de réchauffement initial de la lithosphère ont été suggérées par différents auteurs: soit une intrusion convective de l'asthénosphère, soit l'ascension passive de l'asthénosphère durant une striction de la lithosphère, soit la perturbation thermique à la base de la lithosphère (Sleep, 1971; Ahern et Mrkvicka, 1984; McKenzie, 1978; Detrick et Crough, 1978). Le réchauffement de la lithosphère produit l'expansion thermique et le soulèvement de la surface suivis par l'érosion subaérienne. Par la suite, la lithosphère revient à l'équilibre thermique, se refroidit et se contracte. Le modèle de refroidissement est compatible avec le fait que l'épaisseur de la couche élastique de la lithosphère semble augmenter au cours de l'évolution des bassins du Michigan et du Williston (Haxby *et al.* 1976; Ahern and Ditmars, 1985).

McKenzie (1978) a proposé un modèle où la condition initiale est le résultat de la striction rapide de la lithosphère. La striction, est une amincissement par déformation plastique de la lithosphère avec un fort étirement dans la direction horizontale. L'amincissement de la lithosphère cause une remontée de l'asthénosphère chaude. Le fossé localisé formé par la striction est limité par des failles normales et/ou grabens. La subsidence tectonique dans ce type de bassins est la somme de deux composantes: la subsidence initiale causée par l'amincissement de la croûte et de la lithosphère et la subsidence thermique due à la contraction. Le modèle de striction a été appliqué à de nombreux bassins tels que la mer du Nord (Sclater et Christie, 1980), le bassin Pannonien (Royden *et al.* 1983a, 1983b) et les marges de l'Atlantique de l'Amérique du Nord (Royden et Keen, 1980; Keen et Beaumont, 1990).

Le modèle de striction implique un amincissement de la croûte et ne peut pas être appliqué aux bassins intracratoniques qui ont une croûte épaisse tels que les bassins du Michigan, du Williston, et d'Illinois.

Les changements de phase dans la croûte inférieure et/ou dans le manteau supérieur pourraient affecter la subsidence des bassins intracratoniques. La transformation de la phase moins dense en phase plus dense implique une diminution du volume et a pour conséquence la subsidence. La transformation du gabbro ayant une densité de 3.0 $Mg.m^{-3}$ en éclogite avec une densité 3.5 $Mg.m^{-3}$ a souvent été considérée pour les bassins sédimentaires. Kennedy (1959) et Lovering (1958) ont expliqué les mouvements tectoniques verticaux par un changement de phase (du gabbro en éclogite) au niveau de la discontinuité de Mohorovicic. D'après Haxby *et al.* (1976) cette transformation a causé la subsidence du bassin du Michigan et d'après Fowler et Nisbet (1985) elle a influencé l'évolution du bassin du Williston. Toutefois, la présence d'un tel changement de phase au niveau du Moho a soulevé des débats considérables. Green et Ringwood (1967), Ringwood (1972) affirment que la transformation gabbro-éclogite ne se produit pas dans la croûte continentale à cause de la faible vitesse de la transformation. Au contraire, les travaux expérimentaux de Ito et Kennedy (1970, 1971) suggèrent qu'une telle transformation est possible, et qu'elle peut se produire en un temps relativement court (inférieur à 10^6 ans). Dans les diagrammes de phases obtenus par ces auteurs, la transformation du gabbro en éclogite s'effectue en trois étapes: du gabbro en granulite à grenat, de la granulite à grenat en éclogite à plagioclase et de l'éclogite à plagioclase en éclogite.

Un changement de phase est compatible avec les données géophysiques telles que: 1) une vitesse élevée des ondes P dans la croûte inférieure et dans le manteau supérieur sous certains bassins intracratoniques (Hajnal *et al.* 1984) et 2) une anomalie de Bouguer positive après la correction pour l'effet des sédiments qui est interprétée en terme d'excès de masse dans la croûte inférieure ou dans le manteau supérieur (Datonji, 1981; Haxby et al. 1976).

Le problème majeur du modèle de contraction thermique est qu'il n'est pas en mesure d'expliquer les épisodes d'accélération du taux de subsidence que l'on observe dans la plupart des bassins intracratoniques. Il a été proposé que cette accélération puisse s'expliquer par plusieurs événements thermiques, ou par les variations eustatiques du niveau de la mer, ou par les variations des contraintes dans la plaque tectonique (Sleep, 1976; Cloetingh, 1988; DeRito, *et al.* 1983). De fait, certains épisodes d'accélération peuvent être corrélés avec une remontée du paléo-niveau de la mer. Cependant, l'accélération de la subsidence durant le stade initial de la formation des bassins du Michigan et du Willison (c.a.d. environ 20-40 Ma après le début de leur subsidence) ne semble pas être corrélée avec une remontée du niveau de la mer. Ce phénomème est peut-être dû au fait que des mécanismes autres que la contraction thermique ont affecté l'évolution de ces bassins.

Les changements de phase pourraient être à l'origine de cette croissance du taux de subsidence du stade initial de développement des bassins intracratoniques. La subsidence due aux changements de phase débute des que le refroidissement atteint les changements de phase. A partir de ce temps, la subsidence tectonique devient la somme de deux composantes: la subsidence due à la contraction thermique et la subsidence produite par les changements de phase.

Un mécanisme de subsidence tectonique combinant la contraction thermique et le changement de phase dans la croûte inférieure ou dans le manteau supérieur a été proposé et étudié pour expliquer la formation des bassins intracratoniques. Dans ce mécanisme, les phénomènes mentionnés sont engendrés par le changement de la condition thermique à la base de la lithosphère. Les calculs simulent deux formes de perturbation thermique avec différentes conditions aux limites: 1) une diminution brusque de la température; 2) une diminution du flux de chaleur à la base de la lithosphère.

Un des objectifs de cette recherche est de modéliser l'évolution des bassins du Michigan et du Williston suivant ces différentes conditions aux limites. Ces conditions peuvent être interprétées comme la conséquence de différents mécanismes d'interaction thermique entre les plumes mantéliques et la lithosphère.

Pour le bassin du Michigan, la durée du refroidissement et de la transformation de la granulite à grenat en éclogite suite à une diminution soudaine de la température à la base de la lithosphère est compatible avec l'histoire du bassin.

Toutefois, ce modèle ne peut s'appliquer au bassin du Williston qui se distingue des autres bassins de l'Amérique du Nord (Michigan, Baie d'Hudson, Illinois) par la longue durée de sa subsidence. Il faut noter que l'évolution du bassin du Williston fut complexe, et qu'une partie de sa subsidence est reliée au développement du bassin d'avant-pays d'Alberta. La subsidence locale dura environ 370 Ma, mais l'épaisseur totale de sédiments est seulement de 2.8 km; ceci est très modeste par rapport au bassin du Michigan qui contient 3.6 km de sédiments déposés en 160 Ma. Les causes de cette longue évolution du bassin du Williston restent énigmatiques. Plusieurs auteurs ont évoqué les changements de phase (Fowler et Nisbet, 1985; Quinlan, 1987), mais aucun modèle n'a été développé pour tester cette hypothèse.

Le mécanisme de subsidence tectonique développé restera valide pour expliquer l'évolution du bassin du Williston, si on suppose que la perturbation thermique est causée par le changement brusque du flux de chaleur à la base de la lithosphère. Dans ce cas, les calculs ont démontré que, pour cette condition aux limites, le retour de la lithosphère à l'équilibre est 4 fois plus lent. Le changement de phase pourrait être la transformation du gabbro en granulite à grenat dans la croûte supérieure.

L'autre objectif est de caractériser le paléo-régime thermique des marges continentales passives de l'est du Canada (de Nouvelle-Écosse, de la mer du Labrador et du nord-est de Terre Neuve).

L'étude du paléo-régime thermique des marges continentales passives est importante, d'une part parce que le mécanisme de subsidence est lié au régime thermique qui reflète donc l'évolution tectonique, et d'autre part parce que la température dans les sédiments affecte la maturation des hydrocarbures.

L'estimation du régime thermique permet donc d'approfondir la connaissance de l'évolution tectonique des marges et complète d'autres données géologiques et géophysiques lors de l'évaluation du potentiel en hydrocarbures des marges passives.

Les marges passives sont tectoniquement actives pendant la phase de "rifting" (de cassure). Cette phase se caractérise par un amincissement substantiel de la croûte, le développement de failles normales, le soulèvement thermique de la surface, le volcanisme, l'érosion subaérienne et un flux de chaleur élevé. Durant la phase "drifting" (de dérive) qui succède à celle de "rifting", le refroidissement de la lithosphère cause la subsidence et le dépôt de sédiments le long des marges.

Les changements de phase ne semblent pas jouer de rôle dans la subsidence tectonique des marges continentales passives. La subsidence tectonique de ces marges obéit à la relation théorique en $t^{1/2}$ du mécanisme de contraction thermique, où t est le temps écoulé depuis le début de la subsidence (Sleep, 1971; Sleep, et Snell, 1976). Cette relation reste valable pour les marges continentales passives de l'est du Canada, à l'exception des marges du nord-est de Terre-Neuve (Keen, 1979). La subsidence anomalique de cette région, où la subsidence initiale fut très rapide et suivie d'une période de subsidence lente, n'est pas expliquée jusqu'à présent.

La subsidence due à la contraction thermique est proportionnelle à la quantité de la chaleur perdue par la lithosphère durant son refroidissement. L'excès du paléo-flux de chaleur peut être directement estimé à partir du taux de subsidence tectonique en utilisant des relations entre le taux de la subsidence tectonique et le flux de chaleur. Les relations entre le taux de subsidence tectonique et le flux en excès dépendent des conditions initiales et des conditions aux limites (Mareschal, 1987, 1991). Étant donné l'incertitude sur les conditions aux limites, l'excès du flux de chaleur a été calculé avec deux conditions: 1) flux constant à la base de la lithosphère et 2) température constante à la base de la lithosphère.

Le flux de chaleur observé des marges continentales passives est la somme de trois composantes: le flux de chaleur réduit qui tient compte du flux provenant du manteau et des sources radioactives profondes, le flux causé par la production de chaleur de la couche radioactive de la croûte supérieure et l'effet transitoire dû au réchauffement de la lithosphère durant l'ouverture des continents (c.a.d. le flux de chaleur en excès). La valeur du flux réduit est estimée à 27 $mW.m^{-2}$ et le flux causé par la production de chaleur peut être estimé en tenant compte de la production de chaleur dans les sédiments et de celle de la croûte supérieure amincie par l'extension. Le flux actuel des marges étudiées fournit donc une contrainte sur le flux en excès.

En résumé, les objectifs principaux de cette recherche sont les suivants: 1) développer les modèles de la subsidence tectonique en combinant l'effet de la contraction thermique et des changements de phase; 2) montrer que ce mécanisme fournit une explication à l'augmentation du taux de la subsidence tectonique dans le stade initial de la formation des bassins sédimentaires; 3) montrer que les différences entre les bassins du Michigan et du Williston pourraient s'expliquer par différentes conditions à la base de la lithosphère; 4) estimer le paléo régime thermique des marges continentales passives de l'est du Canada dans le but d'approfondir la connaissance sur leur évolution tectonique.

La thèse a été accomplie par cumul de publications soumises à des revues scientifiques. Le premier article intitulé " Phase changes and thermal subsidence in intracontinental sedimentary basins" a été publié en 1991, dans *Geophysical Journal International,* **106,** pp. 657-665, le deuxième article " Paleo heat flow in eastern Canada's passive margins" a été accepté et sera publié dans un volume spécial de *Tectonophysics* consacré à la structure thermique de la lithosphère et le troisième, intitulé "Phase change and thermal subsidence of the Williston bassin" est soumis au *Journal canadien des Sciences de la Terre.* Chacun des articles mentionnés est présenté sous la forme d'un chapitre distinct.

La conclusion de cette thèse résume les principaux résultats qui découlent de ces trois articles. Elle met en évidence les problèmes en suspens et la contribution de cette recherche aux études des bassins sédimentaires intracratoniques et des marges continentales passives. Elle définit également quelques directions pour de futures recherches.

CHAPITRE I

PHASE CHANGES AND THERMAL SUBSIDENCE IN INTRACONTINENTAL SEDIMENTARY BASINS

Yvette Hamdani, Jean-Claude Mareschal and Jafar Arkani-Hamed

Summary.

A model of tectonic subsidence is developed to explain the late acceleration of subsidence observed in some intracratonic sedimentary basins. The proposed mechanism combines the effect of thermal contraction of an initially hot lithosphere with the effect of a subcrustal phase transformation that moves under changing pressure and temperature conditions.

The subsidence following a sudden change in temperature at the base of the lithosphere is calculated. The calculations show that: (1) phase changes, if present and activated, contribute substantially to the subsidence of sedimentary basins; (2) because the effect of phase change is delayed, subsidence accelerates after a time on the order of 20 Myr, and (3) the duration of the subsidence is on the order of 100 to 150 Myr. During the late stages of subsidence, the phase change is the dominating mechanism.

An application to the Michigan Basin is presented. The calculated sediment accumulation history fits well the record when the effect of sea-level changes is included in the model.

Key words: intracratonic sedimentary basins, Michigan basin, phase changes, tectonic subsidence

1. Introduction.

Thermal contraction of an initially hot lithosphere has been widely accepted as the mechanism of subsidence of sedimentary basins. Several causes have been suggested for the prior heating of the lithosphere, either some form of convective intrusion of the asthenosphere (Sleep 1971; Sleep & Snell, 1976; Sleep, Nunn & Chou 1980; Ahern & Mrkvicka, 1984; Nunn, Sleep & Moore 1984. 1984; Ahern & Ditmars, 1985), or passive upwelling of the asthenosphere during lithospheric stretching (McKenzie 1978; Jarvis & McKenzie, 1980). Sleep & Snell (1976) considered that thermal expansion of the lithosphere causes uplift and erosion; subsequently, the lithosphere returns to thermal equilibrium, cools and subsides by thermal contraction.

Alternatively, phase changes had been considered as a potential subsidence mechanism for intracratonic basins. Lovering (1958) and Kennedy (1959) suggested that uplift and subsidence could be explained by a phase transition at Moho depth. They proposed that pressure increase due to the weight of sediment causes the metamorphic transformation of gabbro into garnet-granulite and garnet-granulite into eclogite, and consequently subsidence. Numerical and analytical studies showed that phase changes are indeed a feasible mechanism of uplift and subsidence and cause cycles of sediment deposition followed by uplift and erosion (McDonald & Ness, 1960; O'Connell & Wasserburg 1967, 1972; Mareschal & Gangi 1977a). However, the hypothesis has been widely discounted because of experimental data indicating that the transformation does not take place in normal conditions because of the low rate of the reactions (Green & Ringwood 1967; Ringwood 1972). Green & Ringwood (1967) extrapolated the phase diagrams to surface pressure conditions; they found that eclogite would be stable for temperatures below 200°C and garnet-granulite between 200°C and 500°C; gabbro is thus metastable in the crust for temperatures below 500°C. This conclusion was challenged by Ito & Kennedy (1970, 1971) who studied the transformation in the 800-1250°C temperature range and obtained phase diagrams distinctly different from Ringwood & Green's. They showed three stages in the gabbro-eclogite transformation: gabbro to granet-granulite, garnet-granulite to plagioclase-eclogite, and plagioclase-eclogite to eclogite. The lines separating the fields of stability of these phases have slopes different from those of Ringwood & Green; extrapolation to lower pressures and temperatures shows that gabbro and garnet-granulite are stable for normal lower crustal conditions. The reaction rate depends strongly on temperature: in their laboratory studies, Ito and Kennedy obtained equilibrium in 5 minutes at 1200°C and 1 day at 800°C. They concluded that the reaction would take place in 1 Myr or less for temperatures above 400°C. Gabbro remains metastable in the crust only for temperatures lower than 400°C. The reaction rate was also studied by Ahrens & Schubert (1975) who concluded that the transition takes place in a geologically short time above 600°C (they computed that at 627°C, equilibrium is reached in 100 000 years).

Haxby, Turcotte & Bird (1976) proposed an alternative process for the phase change to trigger subsidence in sedimentary basins: they assumed that garnet-granulite is metastable in the lower crust, and that the intrusion of a mantle diapir, heats the lower crust and activates the transformation of garnet-granulite into eclogite. They applied this model to the Michigan basin.

The sedimentary basin phase change hypothesis seems to be supported by the observation of higher P wave velocity in the mantle below some intracratonic basins. For instance, Hajnal *et al.* (1984) have determined P_n velocity on the order of 8.3-8.5 $km.s^{-1}$ below the central part of the Williston basin. This would be characteristic of eclogite at 1,500 to 1,800 MPa pressure (Ito & Kennedy 1970). The P wave velocity in the lower crust, is on the order of 7.0 to 7.6 $km.s^{-1}$, thus typical of garnet-granulite.

Figure 1 compares the subsidence records of the Michigan, Williston, and Hudson's Bay basins (data compiled respectively by Nunn & Sleep 1984, Quinlan 1987, and Fowler & Nisbet 1985). In all these basins, the subsidence started at ca 500 Ma. The subsidence lasted about 160 Myr in the Michigan Basin and 100 Myr in Hudson's Bay Basin, but the total subsidence of the Michigan Basin is double that of Hudson's Bay. This may be because the initial thermal perturbation was larger in the Michigan Basin; alternatively, it may indicate that an additional mechanism, such as a phase change, operated in this basin. The Michigan Basin contains Cambrian to Lower Ordovician units; however these units do not show the present characteristic shape of the basin, but are part of an elongated trough (Catacosinos 1973). The subsidence of the proto-Michigan Basin has commonly been interpreted as part of the general subsidence of the Reelfoot rift and Illinois Basin (Sleep *et al.* 1980; Nunn *et al.* 1984). The distinctive shape of the Michigan Basin appears only in the Middle Ordovician strata (e.g. Nunn *et al.* 1984). The Michigan Basin contains several unconformities that are correlated with changes in sea-level (Sleep 1976). In contrast to the Michigan Basin, the Williston Basin has undergone a longer and more complex history of subsidence. Subsidence started at 525 Ma and ended at 75 Ma, with several episodes of deposition interupted by unconformities. The total thickness of sediments at the center of the Basin is relatively modest, less than 2500 m.

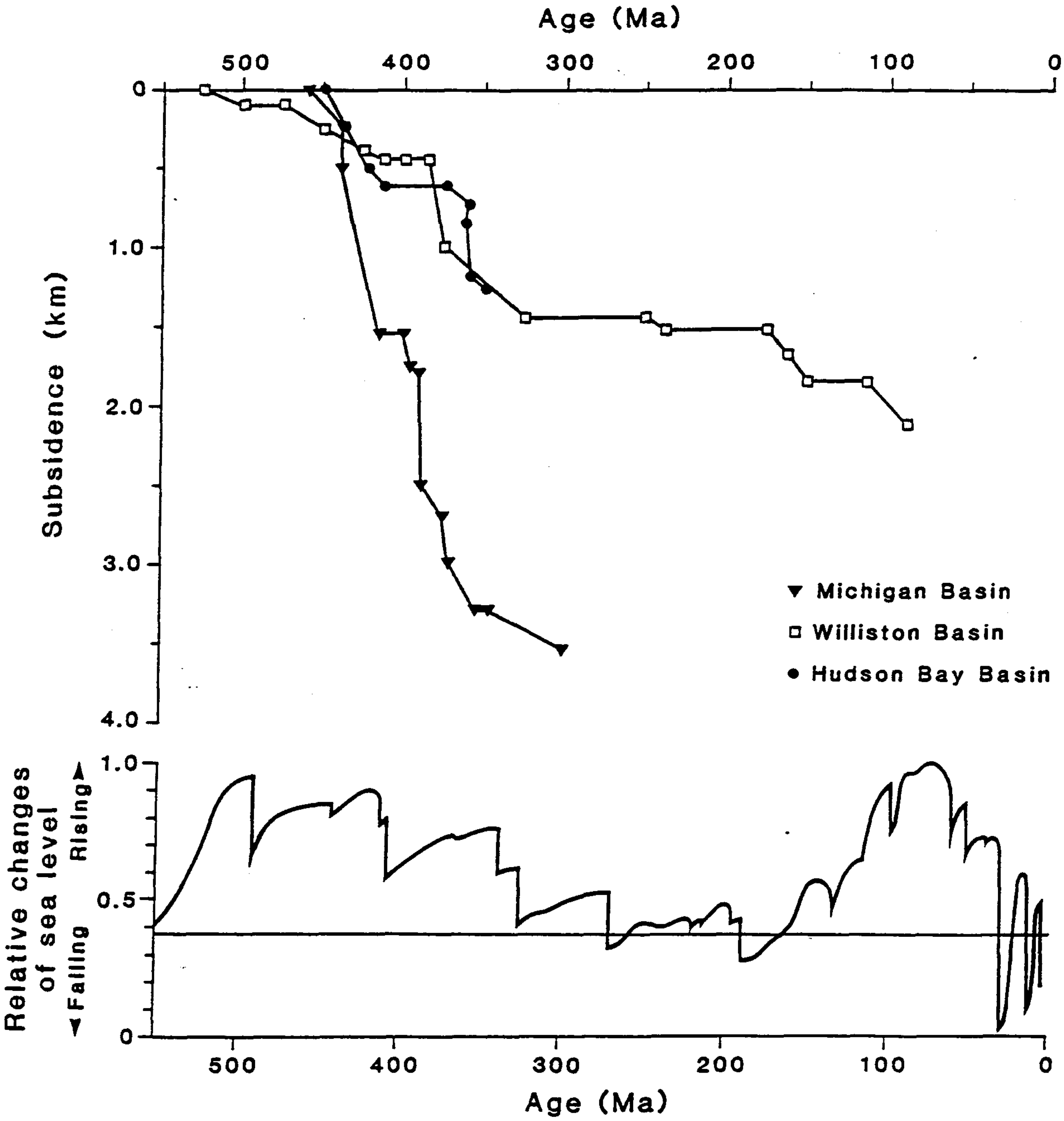


Figure 1: Comparison of the subsidence history in three north American sedimentary basins: Michigan, Hudson's Bay, and Williston, and sea level variations (Vail *et al.* 1977).

Multiple thermal events, a thicker lithosphere, or a different cooling mechanism are among tentative explanations for this very long subsidence record. Ahern & Ditmars (1985) have proposed that this complex history can be explained by two thermal events. Alternatively, the history of the Williston Basin could be explained by a single episode of cooling, with the appropriate boundary condition. The various unconformities between episodes of subsidence are then the result of eustatic changes. Fowler & Nisbet (1985) showed that the unconformities in the Williston basin are correlated with changes in sea-level and they suggested that the non exponential character of the subsidence is explained by the transformation of subcrustal materials into eclogite (see also Jerome 1988). Phase transitions could indeed explain the late acceleration of subsidence in some basins. The cooling of the lithosphere caused thermal subsidence; the phase change induced subsidence was delayed because some time is necessary for cooling to affect the phase change boundary in the lower crust.

Mathematically, the motion of a phase change boundary is determined by the solution to a Stefan-like problem (Carslaw & Jaeger 1959). This problem is difficult to solve analytically because it is non-linear. Analytical approximations to a linearized problem have been derived by O'Connell & Wasserburg (1967, 1972), Gjevik (1972, 1973), Mareschal & Gangi (1977a). The effect of the non-uniformity of surface loading was analyzed by Mareschal & Gangi (1977b). Gliko & Mareschal (1989) compared a non-linear asymptotic expansion to the linear approximation and they concluded that the linear approximation is valid as long as the phase boundary motion is smaller than the depth of the phase boundary.

The objective of this paper is to examine the combined effects of thermal contraction and phase changes on the subsidence of intracontinental sedimentary basins. The tectonic subsidence is determined for a stepwise change in the temperature boundary condition at the base of the lithosphere. The calculations show that the subsidence induced by the phase boundary is retarded by about 20 Myr. An example is presented to reconstruct the subsidence history of the Michigan Basin after a sudden change in temperature at the base of the lithosphere. The calculated subsidence, combined with the effect of sea level changes, fits well the sediment accumulation record for the Michigan Basin.

2. Formulation of the problem.

Thermal subsidence is determined by the decay of a transient thermal perturbation. This perturbation, Θ_1, is solution of the heat conduction equation (Carslaw & Jaeger, 1959):

$$\frac{\partial \Theta_1}{\partial t} = \kappa \frac{\partial^2 \Theta_1}{\partial z^2} \tag{1}$$

with appropriate initial and boundary conditions. κ is the thermal diffusivity, t is time, z is the vertical coordinate (defined positive downward), z = 0 is the surface, z = l is the lithosphere-asthenosphere boundary (LAB). A list of symbols and the values assumed for some important parameters is given in Table 1.

If thermoelastic effects are neglected (i.e. the thermal contraction is in the vertical direction), the thermal subsidence, S_o, is obtained as:

Table 1 : List of symbols and values of parameters.

SYMBOL	DEFINITION	VALUE
c	specific heat	700 $J.kg^{-1}.^{\circ}K^{-1}$
g	acceleration of gravity	9.8 $m.s^{-2}$
K	thermal conductivity	2 $W.m^{-1}.^{\circ}K^{-1}$
l	thickness of the lithosphere	100-150 km
L	latent heat	50 $J.g^{-1}$
P_o	surface pressure excitation	
P_m	pressure of thermodynamic equilibrium	
s	variable of Laplace transform	
S_o	thermal subsidence	400 m
S_l	phase change subsidence	1 km
T_o	surface temperature adjusted for crustal heat sources	
T_c	phase equilibrium temperature at the surface	
T_l	amplitude of temperature change at LAB	**150-250 °K**
T_m	temperature of thermodynamic equilibrium	
z_o	initial depth of the phase boundary	35-50 km
α	coefficient of thermal expansion	3 $10^{-5}\,^{\circ}K^{-1}$
β	geothermal gradient	6-8 $^{\circ}K.km^{-1}$
γ	inverse slope of the Clausius-Clapeyron line	**0.5-0.7 $^{\circ}K.MPa^{-1}$**
Δz_0	amplitude of phase boundary movement	10-20 km
Δz_m	displacement of the phase boundary	
$\Delta\rho$	density contrast between two phase	0.2 $Mg.m^{-3}$
$\Delta\rho/\rho$	relative density change between the two phases	0.07-0.08
Θ_1	thermal perturbation caused by the changing boundary condition	
Θ_2	thermal perturbation caused by latent heat	
κ	thermal diffusivity	$10^{-6}\ m^2.s^{-1}$
ρ	density of the phase being transformed	3.0-3.25 $Mg.m^{-3}$
ρ_m	mantle density	3.3 $Mg.m^{-3}$
ρ_s	sediments density	2.6 $Mg.m^{-3}$
τ	relaxation time for phase boundary motion	1 Myr
τ_o	z_0^2/κ	**40-80 Myr**
τ_l	l^2/κ	300-450 Myr

$$S_0(t) = \alpha \int_0^l \{\Theta_1(z, t=0) - \Theta_1(z,t)\}dz \tag{2}$$

where α is the thermal expansion coefficient.

The motion of the phase change is determined as follows. The Clausius-Clapeyron curve separates the regions of equilibrium of the two phases in a pressure temperature diagram; it is assumed to be linear (Figure 2). The point of intersection of the geotherm with the Clausius-Clapeyron line determines the depth of the phase transition. Initially, the geotherm intersects the Clausius-Clapeyron line at depth z_o; if the lithosphere cools, the new intersection lies at depth z_m. As the phase boundary moves up, transformation of light material into denser material is accompanied by subsidence. Conversely, heating of the lithosphere causes the phase boundary to move downward and results in uplift. Thus, the phase change induces additional subsidence, S_1:

$$S_1 = -\frac{\Delta\rho}{\rho}\Delta z_m \tag{3}$$

where $\Delta\rho$ is the density contrast between the two phases, ρ is the density of the phase being transformed, and Δz_m is the displacement of the interface between the two phases.

The phase transformation causes the release (or absorbtion) of latent heat and a transient thermal perturbation Θ_2, which retards the motion of the phase boundary. The phase boundary location is determined by the condition of thermodynamic equilibrium. The temperature, T_m, and pressure, P_m, corresponding to thermodynamic equilibrium between the two phases, are related by the integrated Clausius-Clapeyron equation:

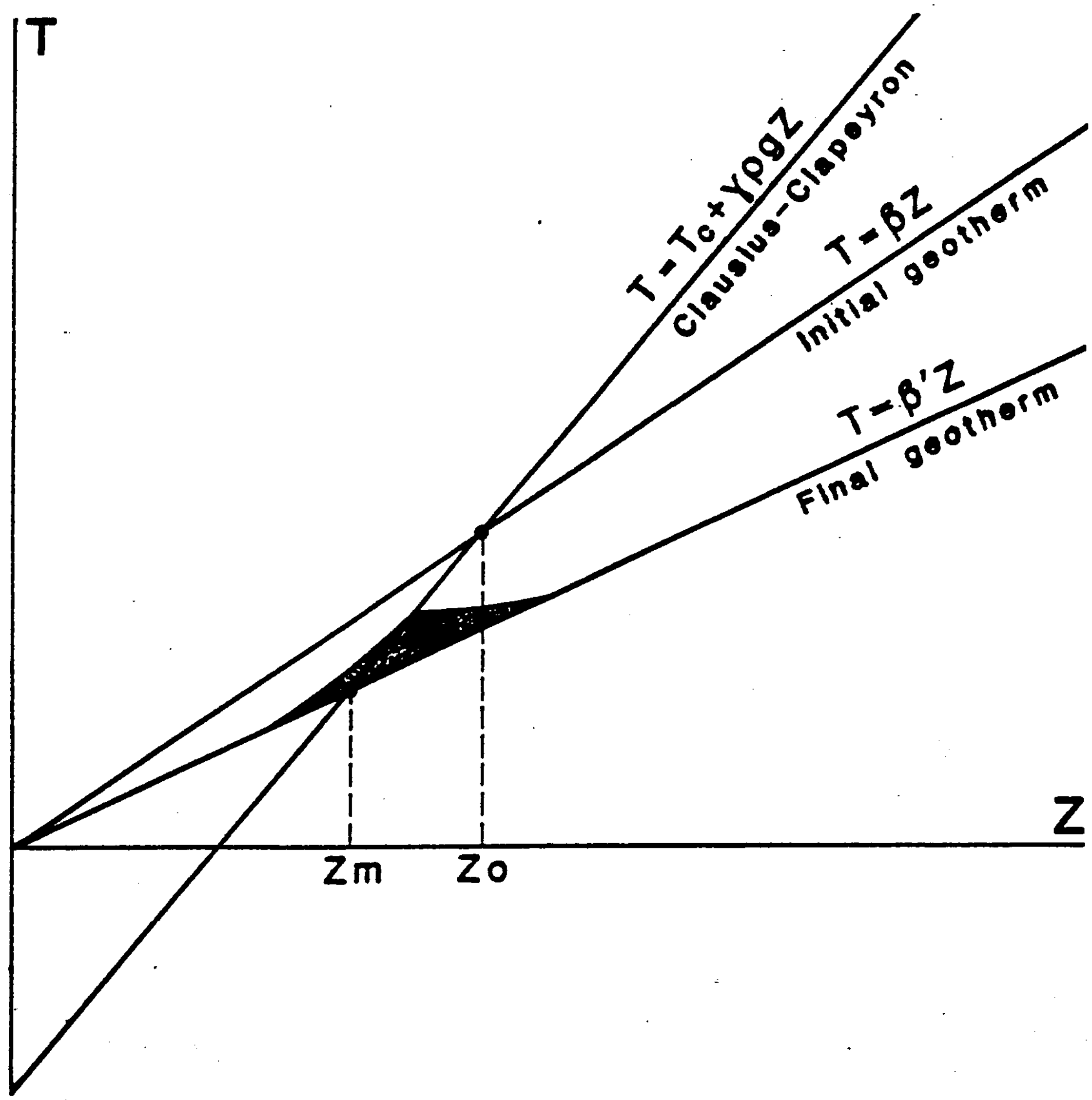


Figure 2: Relationship between Clausius-Clapeyron line, two equilibrium geotherms and the depth of the phase change.The intersection between the geotherm and the Clausius-Clapeyron determines the phase boundary; as the phase boundary moves from z_o to z_m, latent heat is released, raises the temperature (shaded region on the diagram),and retards the phase boundary motion.

$$T(z_m(t),t) = T_c + \gamma P(z_m(t),t) = T_c + \gamma\{g\rho z_m(t) + P_0(t)\} \tag{4}$$

where z_m is the depth of the phase boundary, T_c is the temperature of equilibrium at the surface, γ is the inverse slope of the Clausius-Clapeyron line, g is the acceleration of gravity, P_0 is the surface pressure change (i.e. due to sediment loading).

The temperature in the lithosphere is the sum of several components:

$$T(z,t) = T_0 + \beta z + \Theta_1(z,t) + \Theta_2(z,t) \tag{5}$$

where T_0 is the surface temperature corrected for radiogenic heat production in the crust (i.e. it is obtained by extrapolating the temperature gradient at depth to the surface), β is the geothermal gradient below the shallow crust where radiogenic heat sources are concentrated, Θ_1 is the thermal perturbation caused by the changing boundary condition, Θ_2 is the thermal perturbation induced by the release of latent heat by the moving boundary. Thermodynamic equilibrium thus implies:

$$T_m = T_0 + \beta z_m(t) + \Theta_1(z_m(t),t) + \Theta_2(z_m(t),t) = T_c + \gamma\{g\rho z_m(t) + P_0(t)\} \tag{6}$$

The initial depth of the interface is z_0:

$$T(z_0) = T_0 + \beta z_0 = T_c + \gamma g \beta\rho z_0 \tag{7}$$

Therefore, the phase boundary motion is given by:

$$\Delta z_m(t) = z_m(t) - z_0 = \frac{\Theta_1(z_m(t),t) + \Theta_2(z_m(t),t) - \gamma P_0(t)}{(\gamma g\rho - \beta)} \tag{8}$$

The equation (8) is a non-linear integral equation because the heat source and the point where temperature is defined are both moving. A linear approximation (Appendix A) was obtained by Mareschal & Gangi (1977a) for small displacement of the phase boundary

$z_m(t) \approx z_0$ (i.e. the phase boundary motion is small compared with the depth of the interface). They defined the characteristic time τ for the phase boundary motion as:

$$\tau = \left\{ \frac{L}{2\sqrt{\kappa}\, c\,(\gamma g \rho - \beta)} \right\}^2 \tag{9}$$

where L is the latent heat, c is the specific heat, τ represents the time required for the thermal perturbation induced by the release of latent heat to be diffused away. Mareschal & Gangi (1977a) estimated this time constant τ to be in the range of 0.1 to 1 Myr. The range of validity of the linear approximation was investigated by Gliko & Mareschal (1989) who demonstrated that it is always valid for times smaller than τ.

3. Tectonic subsidence caused by changing temperature at the base of the lithosphere.

The tectonic subsidence is obtained as the superposition of the thermal subsidence, caused by contraction of the cooling lithosphere, and of the subsidence induced by the phase transformation. Both processes are induced by changing thermal boundary conditions at the LAB.

The thermal subsidence for a sudden change in temperature, T_l, at the LAB is derived in Appendix B. It is given by:

$$S_o(t) = \frac{\alpha T_l l}{2} \left\{ 1 - \frac{8}{\pi^2} \sum_{n=0}^{\infty} \frac{1}{(2n+1)^2} \exp(-(2n+1)^2 \pi^2 t/\tau_l) \right\} \tag{10}$$

where $\tau_l = l^2/\kappa$ is the heat conduction time for the lithosphere.

The phase boundary motion following a stepwise change in temperature at the LAB is obtained in Appendix B. The leading term of the series is:

$$\frac{\Delta z_m(t)}{\Delta z_o} = \frac{l}{z_o}\left\{ \operatorname{erfc}\left(\frac{1}{2}\sqrt{\frac{\tau_o'}{t}}\right) - \exp\left(\sqrt{\frac{\tau_o'}{\tau}} + \frac{t}{\tau}\right) \operatorname{erfc}\left(\frac{1}{2}\sqrt{\frac{\tau_o'}{t}} + \sqrt{\frac{t}{\tau}}\right) \right\} \tag{11}$$

where $\Delta z_o = T_l z_o / l(\gamma g \rho - \beta)$, τ is the characteristic time of the phase boundary readjustment, and the time constants $\tau_o = z_o^2/\kappa$ and $\tau_o' = (l - z_o)^2/\kappa$.

The characteristic time τ (the time necessary for the latent heat to diffuse away from the phase boundary) determines the velocity of the phase boundary. Mareschal & Gangi (1977a) have estimated τ to be on the order of 1 Myr or less, which is much smaller than τ_o and τ_o'. This implies that it is the rate of temperature change and not the release of latent heat that controls the phase boundary motion. When $\tau \ll \tau_o'$, the phase boundary movement is simply given by:

$$\frac{\Delta z_m(t)}{\Delta z_0} = 1 + \frac{2l}{\pi z_0} \sum_{n=1}^{\infty} \frac{(-1)^n}{n} \exp(-n^2\pi^2 t/\tau_l) \sin(n\pi z_0/l) \tag{12}$$

The total tectonic subsidence after a stepwise change in temperature at the LAB is shown in Figure 3. Time is relative to τ_l and the total subsidence is normalized to the thermal subsidence (i.e. $\alpha T_l l/2$). On Figure 3a, the depth of the phase boundary is constant and the different curves correspond to different ratios of the subsidence induced by the phase change to the thermal subsidence. On Figure 3b, the phase change subsidence equal to the thermal subsidence and the different curves correspond to different depths of the phase boundary (i.e. τ_o/τ_l).

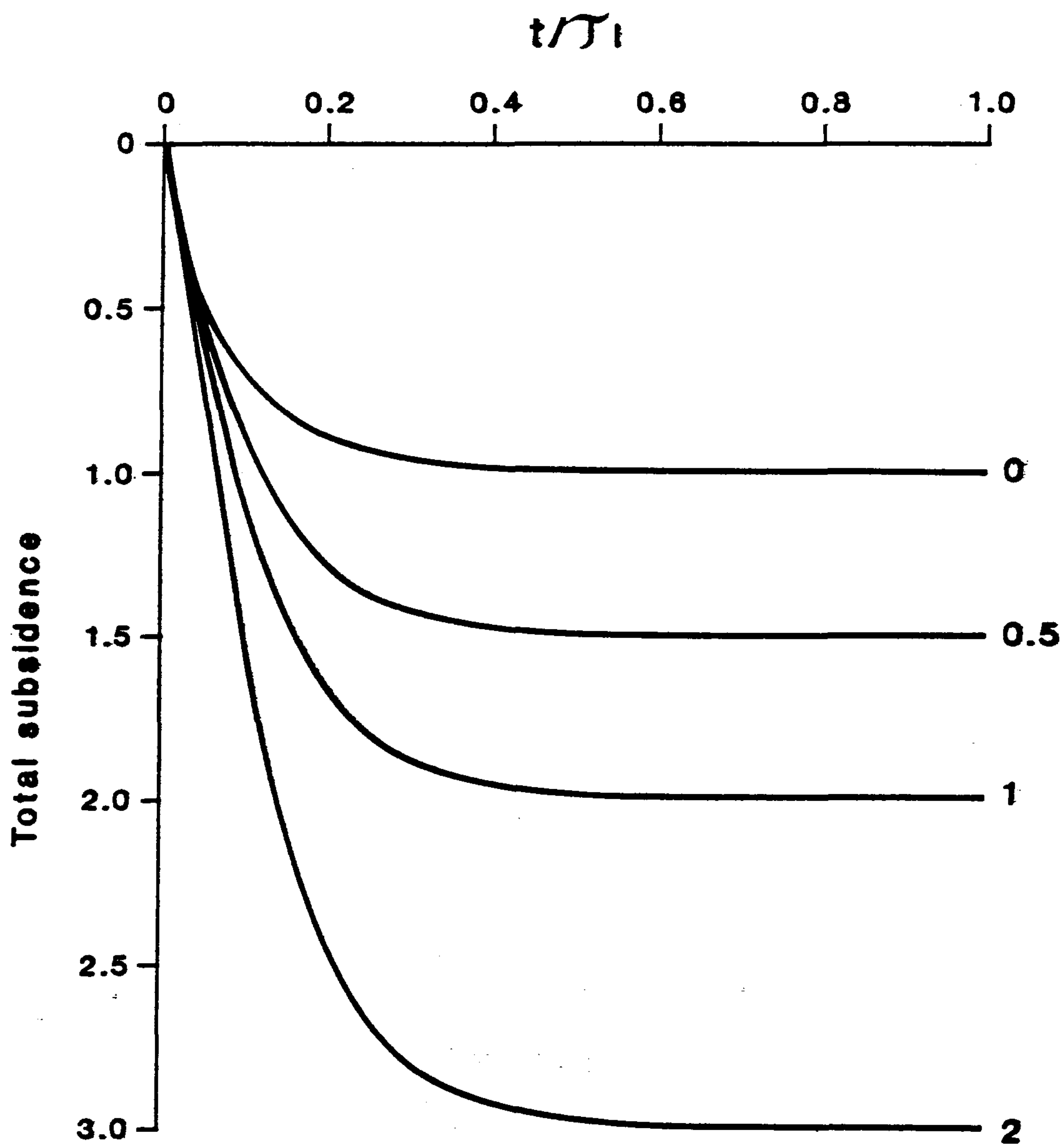


Figure 3a: Total subsidence after a change in temperature at the LAB. The total subsidence is normalized to the amplitude of thermal subsidence. Time is relative to l^2/κ. The different curves correspond to different ratios of phase change to thermal subsidence. $\tau_o/\tau_l = 0.1$.

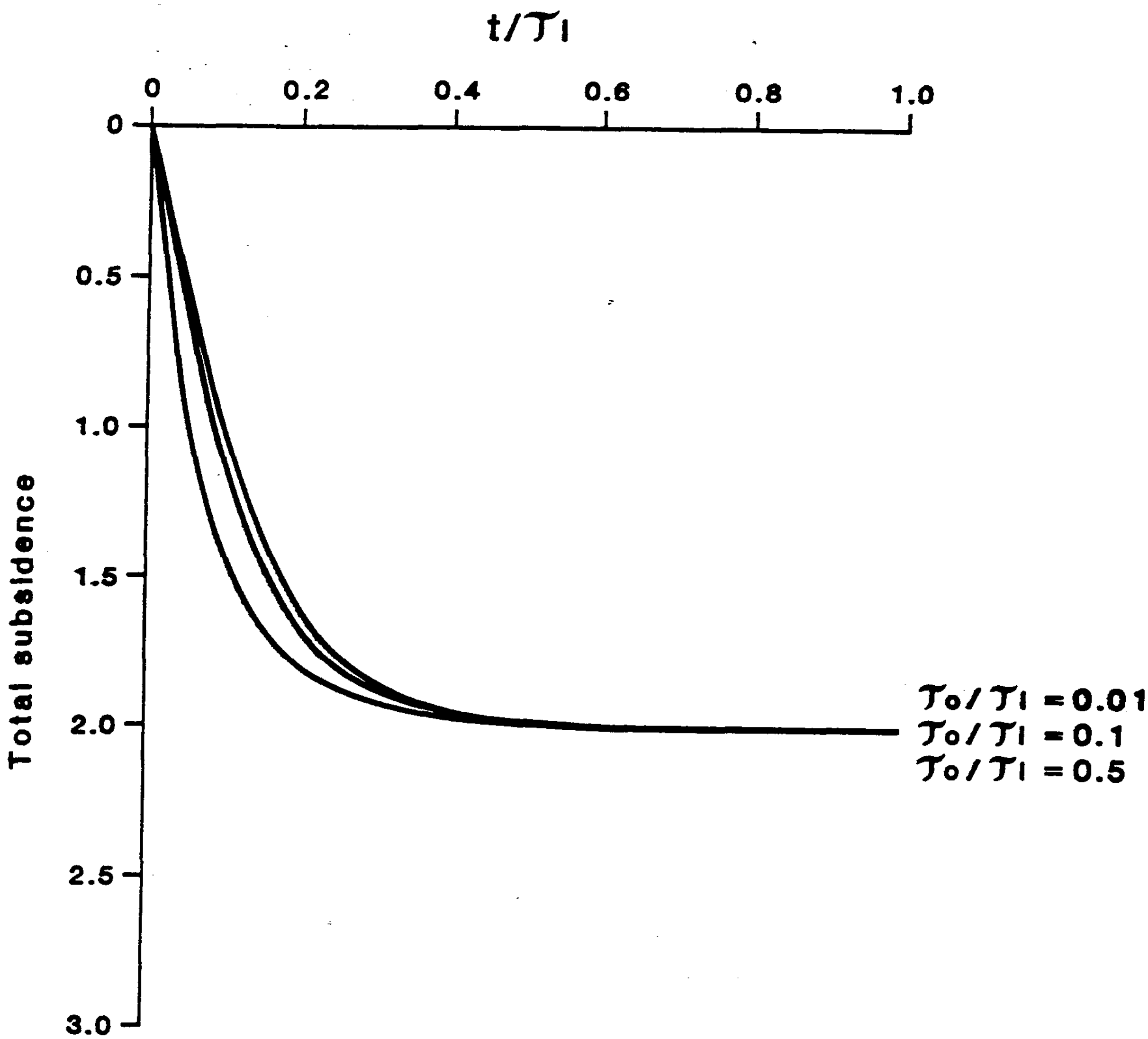


Figure 3b: Total subsidence after a change in temperature at the LAB. The total subsidence is normalized to the amplitude of thermal subsidence. Time is relative to l^2/κ. The different curves correspond to different values of initial depth to phase boundary (i.e. τ_o/τ_l).

The small difference between curves indicates that the initial depth of the phase boundary does not affect much the onset of phase change induced subsidence; however, it affects the amplitude of the phase boundary motion. The amplitude of subsidence is determined by the temperature change in the lithosphere. For a 100 km thick lithosphere, and with standard values of the thermal parameters (i.e. $K = 2$ W $m^{-1}\,{}^{\circ}K^{-1}$, and $\alpha = 3 \times 10^{-5}\,{}^{\circ}K^{-1}$) , a temperature change of 250°K at the LAB induces 375 m of thermal subsidence and 1 km of phase change induced subsidence. Estimates of the temperature difference between hot plumes and normal mantle material is on the order of 150 to 250°K (White & McKenzie, 1989).

The tectonic component of the subsidence is amplified by two processes: (1) isostatic adjustments, and (2) additional migration of the phase boundary due to the weight of the sediment load. For a one-dimensional model, flexural effects are neglected and isostatic adjustments are directly included in the calculations through an isostatic amplification coefficient. For typical values of sediments and mantle density, the isostatic amplification is about 3.3 for Airy's isostasy. The effect of pressure change on the phase boundary is calculated in Appendix C. If the inverse slope of the Clausius-Clapeyron γ is 0.6 °K.MPa^{-1}, and the difference in slopes between the Clausius-Clapeyron and the geotherm is 10 mK.m^{-1} (K/km), the effect of pressure is not very large: the phase boundary migrates by 2 m for 1 m of sediments deposited and it causes 20% additional subsidence.

4. Model of sediment accumulation for the Michigan Basin.

The subsidence history of the Michigan Basin is well documented (see for instance Sleep & Snell, 1976). It contains 3.5 km of sediments that accumulated between 460 and 300 Ma. The sediment accumulation from Cambrian to Lower Ordovician is part of a more general pattern of subsidence including the Reelfoot Rift and the Illinois Basin. The characteristic shape of the Michigan Basin is exhibited only in younger strata. Models of evolution are concerned with the later subsidence history. The record shows periods of marked subsidence followed by periods of low subsidence or erosion. Sleep & Snell (1976), and Nunn & Sleep (1984) explained the subsidence history by the cooling of an initially hot lithosphere with an average excess temperature of 250°K. The main difficulty with a pure cooling model is that it predicts that subsidence starts at a high rate and slows down (i.e. $t^{1/2}$ behavior). Figure 1 shows a pattern that is almost opposite to that prediction: subsidence is slow initially, it accelerates before slowing down again. Sleep (1976) showed that sea-level changes, during the subsidence episode, have affected the sediment accumulation record. The combination of cooling and phase changes could provide an explanation of the subsidence history, if the effect of sea-level changes is included. Figure 4 shows the result of the calculated subsidence of the Michigan Basin induced by a combination of phase change and thermal subsidence. The calculations assume that a sudden 240°K drop in temperature took place at the base of the lithosphere at 460 Ma. The initial lithospheric thickness is assumed to be 140 km, the initial depth of the phase boundary 45 km, the difference between Clausius-Clapeyron and geothermal gradients ($\gamma g \rho - \beta$) 7 mK.m^{-1}, and the relative density change between the two phases $8x10^{-2}$.

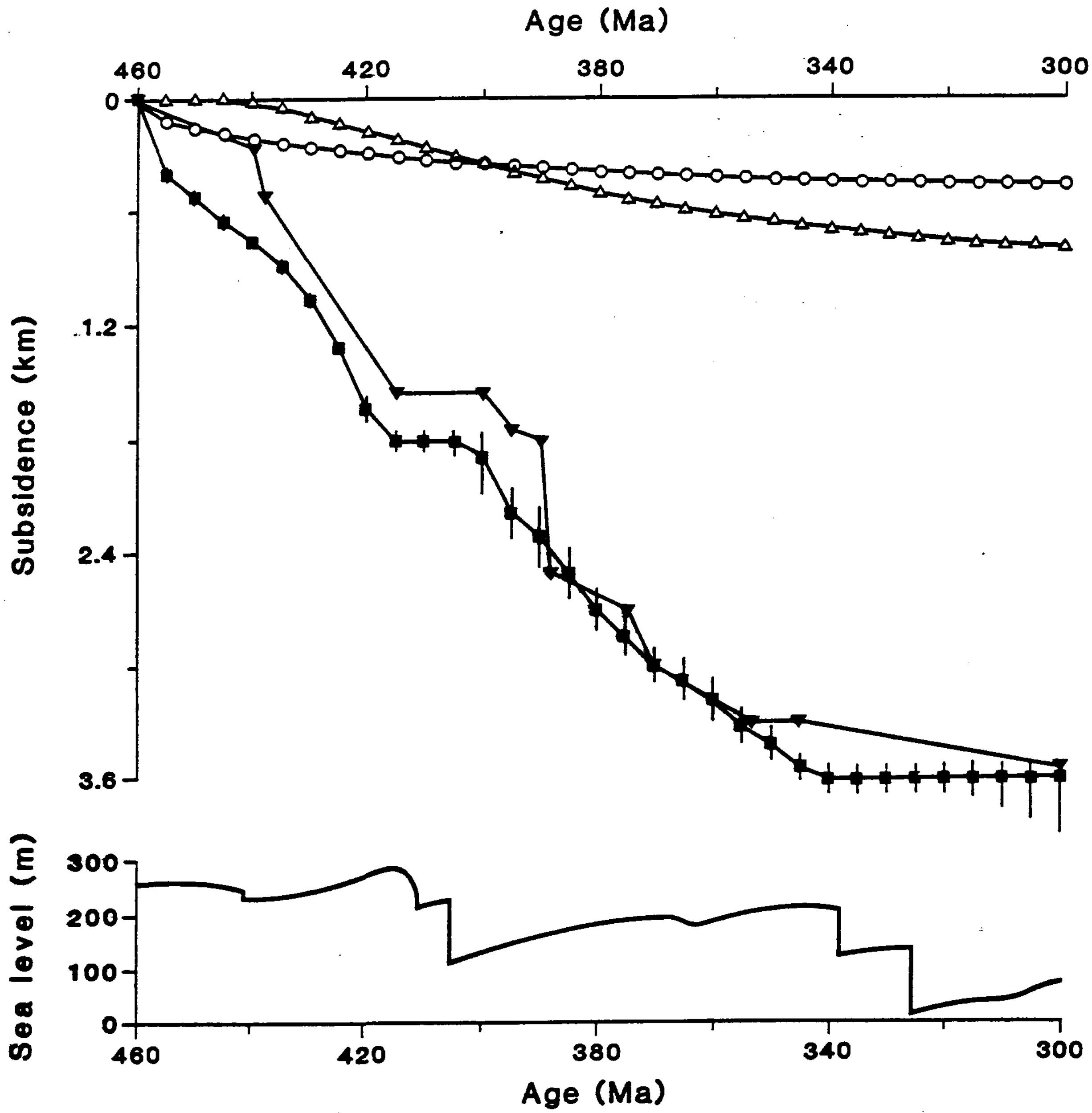


Figure 4: Comparison between calculated, recorded subsidence for the Michigan Basin, and changes in sea level. Open triangles represent the phase change induced subsidence (without isostatic amplification), open circles the thermal subsidence, plain squares the calculated total sediment accumulation including isostasy and sea level changes, and plain triangles the recorded subsidence. The error bars on the calculated subsidence indicate the effect of increasing or decreasing by 50 per cent the amplitude of the sea-level variations.

Airy isostasy is assumed, and the isostatic amplification is 3.3, corresponding to a sediment density of 2.6 $Mg.m^{-3}$ and to a mantle density of 3.3 $Mg.m^{-3}$. The sea level variations determined by Vail, Mitchum & Thompson (1977) were used to calculate the sediment accumulation; sediment deposition and isostatic amplification occur only as long as the basin is below sea-level; during periods where sea level drops, there is no accumulation and it is assumed that erosion is negligible. There is considerable debate about the accuracy of the sea-level curves. The question concerns the amplitude of the sea-level changes rather than the timing of these changes. The calculations were made for different amplitudes in sea-level changes: the "standard" amplitude, 1.5 times the standard amplitude, and 0.5 times the standard amplitude. The model shows that the subsidence induced by phase boundary motion is retarded by 20 Myr; this explains well the acceleration of subsidence around 440Ma. For a 140 km thick lithosphere, the thermal subsidence is completed in 50 Myr. After that time, the phase change induced subsidence dominates and causes most of the subsidence after 400 Ma. The unconformity between 420 and 400 Ma is correlated with a drop in sea-level. It is worth noting that changes in the amplitude of the sea-level variations do not affect much the subsidence history. In particular, the timing and the duration of the unconformities are not very sensitive to the amplitude of the sea-level changes. Only the finer details of the sedimentation history depend on the amplitude of sea-level curves; they are also affected by other effects such as sediment compaction and density. Regardless of these effects, the grosser feature of the subsidence history, in particular the timing of the different sedimentation episodes, are well explained.

5. Discussion and conclusions.

Thermal subsidence was determined for a sudden change in temperature at the LAB. The motion of a phase boundary within the lithosphere was calculated, and the subsidence induced was superposed to the thermal subsidence to obtain total tectonic subsidence. Sea-level variations were included to determine the sediment accumulation history. The calculated subsidence fits surprisingly well the gross features of the sedimentary record in the Michigan basin, considering the large uncertainty on the amplitude of sea level variations. For the parameters of the model, the final depth of the phase change boundary is 35 km, implying the presence of eclogite below the Moho. Haxby *et al.* (1976) suggested that diapiric intrusion had induced the transformation from gabbro to eclogite in the lower crust and caused the subsidence of the Michigan basin. This load would explain the positive gravity anomaly observed over the basin when the effect of the lower density of the sediments has been removed (McGinnis, 1970; Nunn & Sleep, 1984; Van Schmus & Hinze, 1985).

The calculated subsidence fits reasonably well the sediment record for the Michigan basin. This, however, does not demonstrate convincingly that phase changes did indeed affect the evolution of this and other basins.

1) It is not at all clear that the phase change is necessary to explain the subsidence history of the Michigan Basin. The phase change is introduced in part to explain the amplitude of subsidence and in part to explain its acceleration after 20 Myr. The presence of eclogite below the crust is compatible with the gravity anomaly. But a larger temperature change can also explain the amplitude of subsidence (about 1200 m before isostatic amplification).

Because the sea level changes are poorly determined, it is premature to draw definite conclusions from the apparent acceleration of subsidence.

2) Even if phase changes did affect the subsidence of the Michigan basin, the same model cannot be readily extended to explain the much longer subsidence of the Williston Basin. This subsidence history could be accounted for by a complicated thermal history with multiple heating events. Alternatively, different boundary conditions at the LAB could be invoked to explain the dissimilarity between the evolutions of the Williston and the Michigan basins. Calculations show that for a temperature drop at the base of the lithosphere, return to thermal equilibrium is completed in 100-150 Myr and that, for a drop in heat flow at the LAB, return to equilibrium takes four times as long and will require 400-600 Myr (see for instance, Lachenbruch & Sass 1978, Mareschal & Bergantz, 1990). These boundary conditions reflect the interactions between the lithosphere and the asthenosphere and must find a physical interpretation.

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Appendix A: The linear approximation for the phase boundary movement.

The movement of the phase boundary is determined by equation (8). The thermal perturbation, Θ_2, can be expressed as the "convolution" of the Green's function for a point source and the rate of latent heat release at the moving boundary (Carslaw & Jaeger, 1959):

$$\Theta_2(z_m,t) = \frac{L}{c}\int_0^t \frac{\dot{z}_m(t')}{2\sqrt{\pi\kappa(t-t')}} \quad \exp\left\{\frac{-(z_m(t)-z_m(t'))^2}{4\kappa(t-t')}\right\}dt' \qquad (A.1)$$

where c is the heat capacity, L is the latent heat. The Green's function used in (A1) neglects the effect of the surface boundary condition.

The linear approximation is obtained with the assumption that:

$$\left|\frac{(z_m(t)-z_m(t'))^2}{4\kappa(t-t')}\right| \ll 1 \qquad (A.2)$$

Then, the thermal perturbation, Θ_2, is given by:

$$\Theta_2(z_m,t) \approx \frac{L}{c}\int_0^t \frac{\dot{z}_m(t')dt'}{2\sqrt{\pi\kappa(t-t')}} \qquad (A.3)$$

and Θ_1 can be approximated as:

$$\Theta_1(z_m,t) \approx \Theta_1(z_0,t) \qquad (A.4)$$

The equation (8) is now reduced to a linear integral equation that is most conveniently solved in Laplace transform domain (see for instance, Doetsch 1963, Sneddon 1972). The Laplace transform of equation (8) is:

$$z_m(s) = \frac{z_o}{s} + \frac{\Theta_1(z_o,s)+\Theta_2(z_o,s)-\gamma P_o(s)}{(\gamma g\rho-\beta)} \qquad (A.5)$$

s denotes the transform variable; functions of s are Laplace transforms. The Laplace transform of equation (A.3) is given by:

$$z_m(s) = \frac{z_o}{s} + \frac{\Theta_1(z_o,s) + \Theta_2(z_o,s) - \gamma P_o(s)}{(\gamma g \rho - \beta)} \qquad (A.6)$$

and

$$\Theta_2(z_o,s) = \left\{ z_m(s) - \frac{z_o}{s} \right\} \sqrt{\tau s}$$

and the phase boundary is obtained as:

$$z_m(s) = \frac{z_o}{s} + \frac{\Theta_1(z_o,s) - \gamma P_o(s)}{(\sqrt{\tau s}+1)(\gamma g \rho - \beta)} \qquad (A.7)$$

where

$$\tau = \left\{ \frac{L}{2\sqrt{\kappa}\, c(\gamma g \rho - \beta)} \right\}^2 \qquad (A.8)$$

Appendix B: Sudden temperature change at the base of the lithosphere.

The Laplace transform of the heat equation is:

$$\frac{s}{\kappa}\Theta_1(s,z) = \frac{\partial^2 \Theta_1}{\partial z^2} - \frac{\Theta_0(z)}{\kappa} \qquad (B.1)$$

where Θ_0 is the initial condition, which vanishes.

This equation is solved with the following boundary conditions:

$$\Theta_1(z=0,t) = 0 \qquad (B.2a)$$

$$\Theta_1(z=l,t) = T_l \qquad (B.2b)$$

The Laplace transform of the temperature perturbation, $\Theta_1(z,s)$, is obtained as:

$$\Theta_1(z,s) = \frac{T_l}{s} \frac{\sinh\left(\sqrt{\frac{s}{\kappa}}\ z\right)}{\sinh\left(\sqrt{\frac{s}{\kappa}}\ l\right)} \qquad (B.3)$$

The Laplace transform of the thermal subsidence is given by:

$$S_o(s) = \frac{\alpha T_l}{s}\sqrt{\frac{\kappa}{s}}\;\frac{\cosh\left(\sqrt{\frac{s}{\kappa}}l\right)-1}{\sinh\left(\sqrt{\frac{s}{\kappa}}l\right)} \qquad (B.4)$$

which, in time domain, gives:

$$S_o(t) = \frac{\alpha T_l l}{2}\left\{1 - \frac{8}{\pi^2}\sum_{n=0}^{\infty}\frac{1}{(2n+1)^2}\;\exp(-(2n+1)^2\pi^2 t/\tau_l)\right\} \qquad (B.5)$$

where $\tau_l = l^2/\kappa$.

The Laplace transform of the phase boundary motion is obtained from equation (A.5):

$$\Delta z_m(s) = z_m(s) - \frac{z_o}{s} = \frac{T_l}{s(\gamma g\rho - \beta)(\sqrt{s\tau}+1)}\;\frac{\sinh\left(\sqrt{\frac{s}{\kappa}}\;z_0\right)}{\sinh\left(\sqrt{\frac{s}{\kappa}}\;l\right)} \qquad (B.6)$$

which can be rearranged as:

$$\frac{\Delta z_m(s)}{\Delta z_o} = \frac{l}{z_o}\frac{1}{s}\frac{1}{\sqrt{\tau}(\sqrt{\tau s}+1)}\quad\sum_{n=0}^{\infty}\{\exp(-((2n+1)l - z_o)\sqrt{s/\kappa})\ldots$$

$$-\exp(-((2n+1)l + z_o))\sqrt{s/\kappa})\} \qquad (B.7)$$

The phase boundary motion is thus obtained by inverting each term of this series. It yields (Oberhettinger & Badii 1973):

$$\frac{\Delta z_m(t)}{\Delta z_0} = \frac{l}{z_0}\sum_{n=0}^{\infty}\left\{\mathrm{erfc}\left(\frac{1}{2}\sqrt{\frac{\tau_n'}{t}}\right) - \exp\left(-\frac{\tau_n'}{4t}\right)\; w\left(\frac{1}{2}\sqrt{\frac{\tau_n'}{t}}+\sqrt{\frac{t}{\tau}}\right)\ldots\right.$$

$$\left. + \exp\left(-\frac{\tau_n''}{4t}\right)\; w\left(\frac{1}{2}\sqrt{\frac{\tau_n''}{t}}+\sqrt{\frac{t}{\tau}}\right) - \mathrm{erfc}\left(\frac{1}{2}\sqrt{\frac{\tau_n''}{t}}\right)\right\} \qquad (B.8)$$

where $\tau_n' = ((2n+1)l - z_o)^2/\kappa$ and $\tau_n'' = ((2n+1)l + z_o)^2/\kappa$ and $w(x) = \exp(x^2)\,\mathrm{erfc}(x)$, and where Δz_0 is the amplitude of the phase boundary motion:

$$\Delta z_0 = -\frac{T_l z_o}{(\gamma g \rho - \beta) l} \qquad (B.9)$$

If the characteristic time for the phase boundary motion, τ, is neglected (i.e $\tau \ll \tau_0$ and $\tau \ll \tau_l$, equation (B.6) becomes:

$$\Delta z_m(s) = \frac{T_l}{(\gamma g \rho - \beta)} \frac{\sinh(\sqrt{\tau_0 s})}{s \sinh(\sqrt{\tau_l s})} \qquad (B.10)$$

where $\tau_0 = z_0^2/\kappa$ is the thermal diffusion time of the crust, and $\tau_l = l^2/\kappa$ is the thermal diffusion time for the lithosphere. This equation (B.10) could be obtained by assuming that the latent heat is zero (i.e. that the delay of the phase movement by the latent heat is negligible compared to the thermal conduction time of the lithosphere). The inverse of the Laplace transform (B.10) is obtained as:

$$\frac{\Delta z_m(t)}{\Delta z_0} = 1 + \frac{2l}{\pi z_0} \sum_{n=1}^{\infty} \frac{(-1)^n}{n} \exp(-n^2 \pi^2 t/\tau_l) \sin(n \pi z_0/l) \qquad (B.11)$$

Appendix C. Effect of sediment loading on the phase boundary.

In the linear approximation, different causes of the phase boundary motion can be superposed. For instance, the effect of surface pressure change due to sediment loading can superposed to the effect of cooling of the lithosphere. The motion of the phase boundary following a change in pressure at the surface is obtained from equation (A.6):

$$\Delta z_m(s) = -\frac{\gamma P_0(s)}{(\gamma g \rho - \beta)} \frac{1}{\{\sqrt{\tau s} + 1\}} \qquad (C.1)$$

where $P_0(s)$ is the Laplace transform of the surface pressure change. For a sudden change in pressure at the surface, $P_0(s) = \overline{P}_0/s$, the solution is given by (Gliko & Mareschal, 1989):

$$\Delta z_m(t) = \frac{-\gamma \overline{P}_o}{(\gamma g \rho - \beta)} \{1 - \exp(t/\tau)\,\mathrm{erfc}(\sqrt{t/\tau})\} \tag{C.2}$$

and for a constant rate of change of the pressure (i.e. constant rate of sediment deposition), the solution is obtained by integrating (C.2). It gives:

$$\Delta z_m(t) = \frac{-\gamma \overline{P}'_o \tau}{(\gamma g \rho - \beta)} \left\{ 1 - 2\sqrt{\frac{t}{\pi\tau}} + \frac{t}{\tau} - \exp(t/\tau)\,\mathrm{erfc}(\sqrt{t/\tau}) \right\} \tag{C.3}$$

CHAPITRE II

PHASE CHANGE AND THERMAL SUBSIDENCE OF THE WILLISTON BASIN

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Abstract.

A mechanism of subsidence for the Williston basin is investigated which combines cooling and thermal contraction of the lithosphere with a lower crustal phase transformation. The cooling and the phase boundary movement follow a sudden change in heat flow at the lithosphere-asthenosphere boundary. The calculations show that: 1) the heat flow boundary condition explains the long duration (more than 350 Myrs) of subsidence of the Williston basin, and 2) the delay of the phase change subsidence (about 40 Myrs) explains an acceleration of subsidence in the early stage of the basin's evolution. When isostatic adjustements and the effect of sea-level variations are included, the calculated sediment accumulation history fits well the observed record in the Williston basin.

1.Introduction.

Thermal contraction of an initially hot lithosphere is generally considered as the main cause of tectonic subsidence of sedimentary basins (Sleep, 1971; Sleep and Snell, 1976; Sleep *et al.* 1980; Ahern and Mrkvicka, 1984; Ahern and Ditmars, 1985; McKenzie, 1978; Jarvis and McKenzie, 1980). The thermal contraction mechanism behavior predicts that subsidence begins rapidly and slows down with time (i.e. $t^{1/2}$ behavior). This does not fit the subsidence data from several intracratonic basins. For example, the subsidence record of the Michigan, the Illinois, Hudson's Bay, and the Williston basins in North America are inconsistent with a simple thermal contraction model; subsidence accelerates some time after it is initiated (Nunn et Sleep, 1984; Quinlan, 1987; Haid, 1991; Middleton, 1980). This suggests that another mechanism also contributes to subsidence in these basins.

Among other mechanisms, the phase change had been considered as a potential cause of subsidence for intracratonic basins. Lovering (1958) and Kennedy (1959) suggested that uplift and subsidence could be explained by a phase transition from gabbro to eclogite at Moho depth. They proposed that increase in pressure due to the weight of sediment causes the metamorphic transformation of gabbro into garnet-granulite and garnet-granulite into eclogite, and consequently subsidence. Numerical and analytical studies showed that phase changes are indeed a feasible mechanism of uplift and subsidence and cause cycles of sediment deposition followed by uplift and erosion (McDonald and Ness, 1960; O'Connell and Wasserburg, 1967, 1972; Mareschal and Gangi, 1977a). However, there have been serious questions about this hypothesis. The experimental data obtained by Ringwood and Green (1966), Green and Ringwood (1967), and Ringwood (1972) indicate that the transformation from gabbro to eclogite does not take place in normal crustal conditions because of the low rate of the reaction. These results were challenged by Ito and Kennedy

(1970, 1971) who concluded that the transformation would take place in 1 Myr or less for temperature above 400°C. Ahrens and Schubert (1975) also showed that the reaction would occur in a geologically short-time for temperature above 600°C.

Mathematically, the motion of a phase change boundary is determined by the solution to a Stefan-like problem (Carslaw and Jaeger, 1959). This problem is difficult to solve analytically because it is non-linear. Analytical approximations to a linearized problem have been derived by O'Connell and Wasserburg (1967, 1972), Gjevik (1972, 1973), and Mareschal and Gangi (1977a). The effect of the non-uniformity of surface loading was analyzed by Mareschal and Gangi (1977b), and Mareschal and Lee (1983). Gliko and Mareschal (1989) compared a non-linear asymptotic expansion to the linear approximation and they concluded that the linear approximation is valid as long as the amplitude of phase boundary motion is smaller than the depth of the phase boundary.

Favley (1974), Spohn and Neugebauer (1978), Neugebauer and Spohn (1978,1982) and Middleton (1980) proposed that deep crustal metamorphism may contribute to subsidence in continental margins and intracratonic basins. Middleton (1980) suggested that the transformation of the greenschist facies to amphibolite facies produced the initial stage of subsidence in the Cooper and Eromanga basins (in Australia).

Hamdani *et al.* (1991) calculated the thermal and phase change subsidence following a sudden change in temperature at the lithosphere asthenosphere boundary (LAB). They concluded that phase transitions coupled with thermal contraction explain the gross features of the subsidence record, including the acceleration of the subsidence, in the Michigan basin. The cooling of the lithosphere produces thermal contraction and subsidence. The phase boundary is delayed because of the time for cooling at the LAB to reach the phase change boundary (i.e. only some time after the basin's initiation, the phase change motion produces additional subsidence). However, this model is difficult to directly apply to the Williston basin which exhibits a much longer (more than 350 Myrs) local subsidence history with a relatively modest sediments accumulation, because the return to thermal equilibrium following a sudden temperature change at LAB should take place in less than 200 Myrs, depending on lithospheric thickness. In order to explain the long subsidence record, some authors have called for multiple thermal events and several episodes of cooling and thermal contraction (DeRito *et al.* 1983; Komintz and Bond, 1991).

The purpose of this paper is to demonstrate that the Willison basin's subsidence can be explained by a single episode of cooling following a stepwise change in heat flow at LAB. Two models are investigated : 1) thermal contraction only caused by a substantial decrease (7.2 $mW.m^{-2}$) in heat flow at the LAB; 2) a combination of thermal contraction with a phase change following a relatively smaller heat flow (3.2 $mW.m^{-2}$) change at the LAB. In the latter case, a large fraction of the lower crust is transformed from gabbro into garnet-granulite. The long duration of the subsidence in the Williston basin is the result of the heat flow boundary condition. Some acceleration of subsidence in the early stage of the basin's evolution is accounted for by the phase change.

2.Williston basin: Geological and geophysical framework

2.1 Analysis of subsidence history

The Williston basin, one of the major intracratonic basins in North America, straddles the border between Canada and the United States. In its center, this nearly circular basin contains approximatively 5 km of sediments (fig.1) ranging in age from Cambrian to Tertiary (Porter *et al.* 1982; Ahern and Mrkvicka, 1984; Haid, 1991).

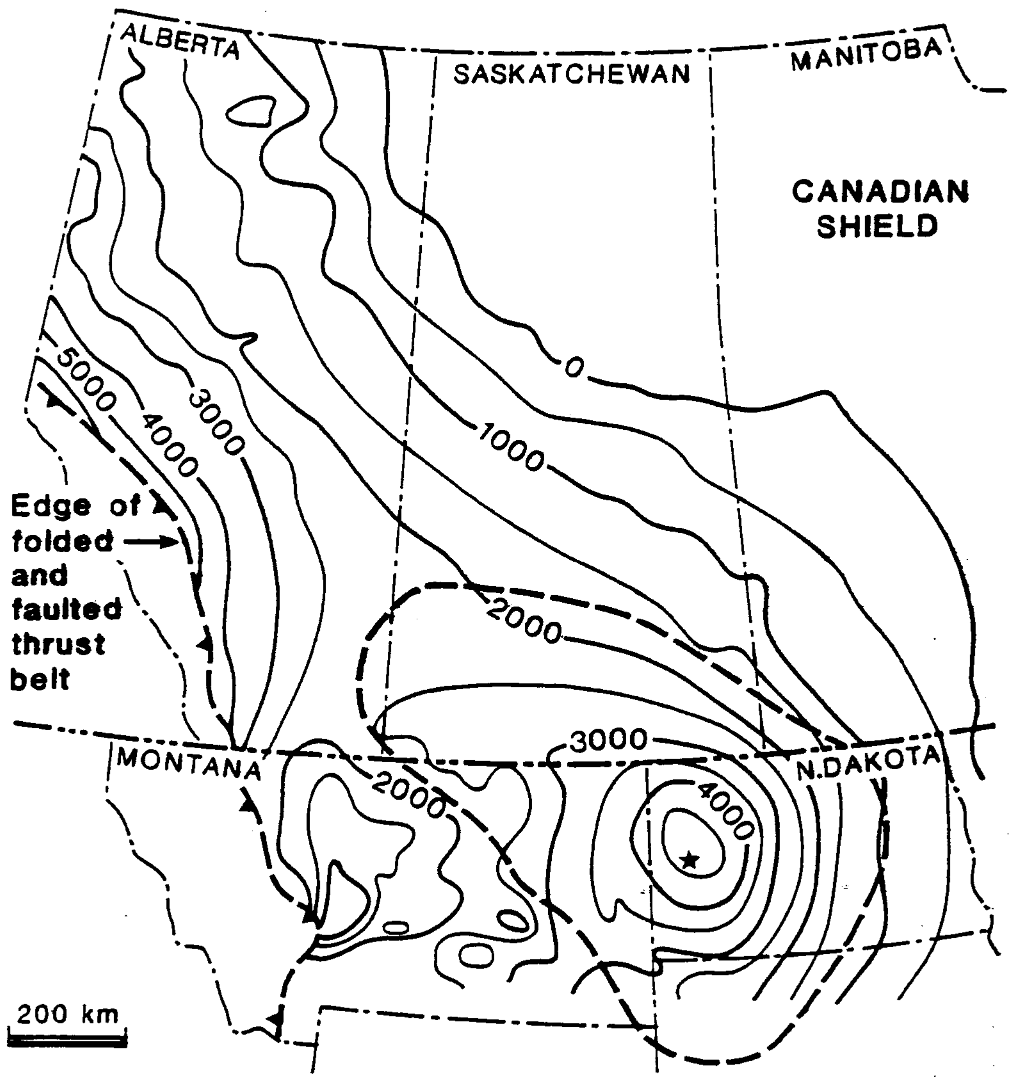

Figure 1: Map of the Williston basin and adjacent areas. Thickness of preserved Phanerozoic rocks in metres with 500-m-interval contour lines. Star indicates the location of the Zabolotony N 1-3-4a well. Dashed lines indicate the limit of the basin. Adapted from Haid (1991).

The subsidence of the Williston basin is quite different from that of other sedimentary basins in North America. Figure 2 compares the subsidence record of four intracratonic basins of North America: the Williston, Michigan, Illinois, and Hudson's Bay basins (data compiled by Haid, 1991; Nunn and Sleep, 1984; and Quinlan, 1987 respectively). In all these basins, the subsidence started between 450 and 525 Ma. The subsidence lasted about 160 Myrs in the Michigan, 250 Myrs in the Illinois, 120 Myrs in the Hudson's Bay and 520 Myrs in the Williston basin. However, the evolution of the Williston basin is linked to the development of the western Canada sedimentary basin and the local basin's subsidence is difficult to discriminate from the regional subsidence. The time of the initiation of the Williston basin is also subject to discussion. In this paper, the subsidence record of the Williston basin and its relation to the development of the western Canada sedimentary basin are discussed in terms of the local basin's history.

The data from Zabolotony N° 1-3-4a well (see fig.1 for location) has been chosen as a reference because the well is located 77 km from the center of the basin and it represents the most complete subsidence record of the Williston basin. The subsidence record, shown on figure 2 is very long and complex with the thickness of sediment reaching 4.65 km. The subsidence in the basin started at ca 520 Ma and ended at present. The subsidence history shows marked increases in subsidence rate during Middle Ordovician (ca 470 Ma), Middle Devonian (ca 387 Ma), Middle Jurassic (ca 180 Ma), and Early Cretaceous (ca 120 Ma). It was suggested by Porter *et al.* (1982), Gerhard *et al.* (1982), Caldwell (1986), Ahern and Mrkvicka (1984) that the local Williston basin's subsidence began only during Middle Ordovician (at 450 Ma) with the deposition of the Tippecanoe sequence (see the stratigraphic column, fig.3 for the name of the different sequences).

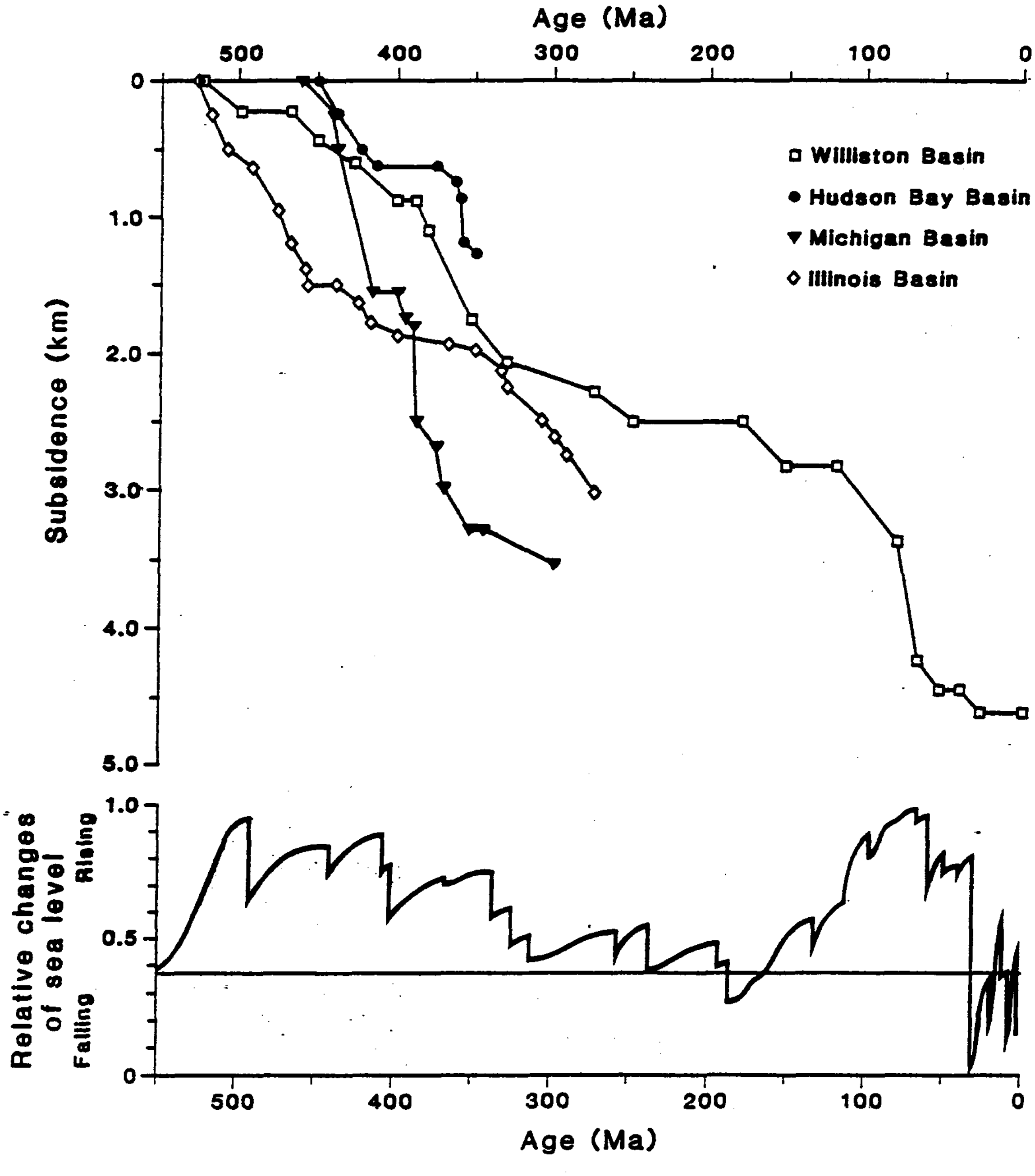

Figure 2: Subsidence of four North American intracratonic sedimentary basins: Michigan, Hudson's Bay, Illinois, and Williston, and sea-level variations (Vail and Mitchum, 1979).

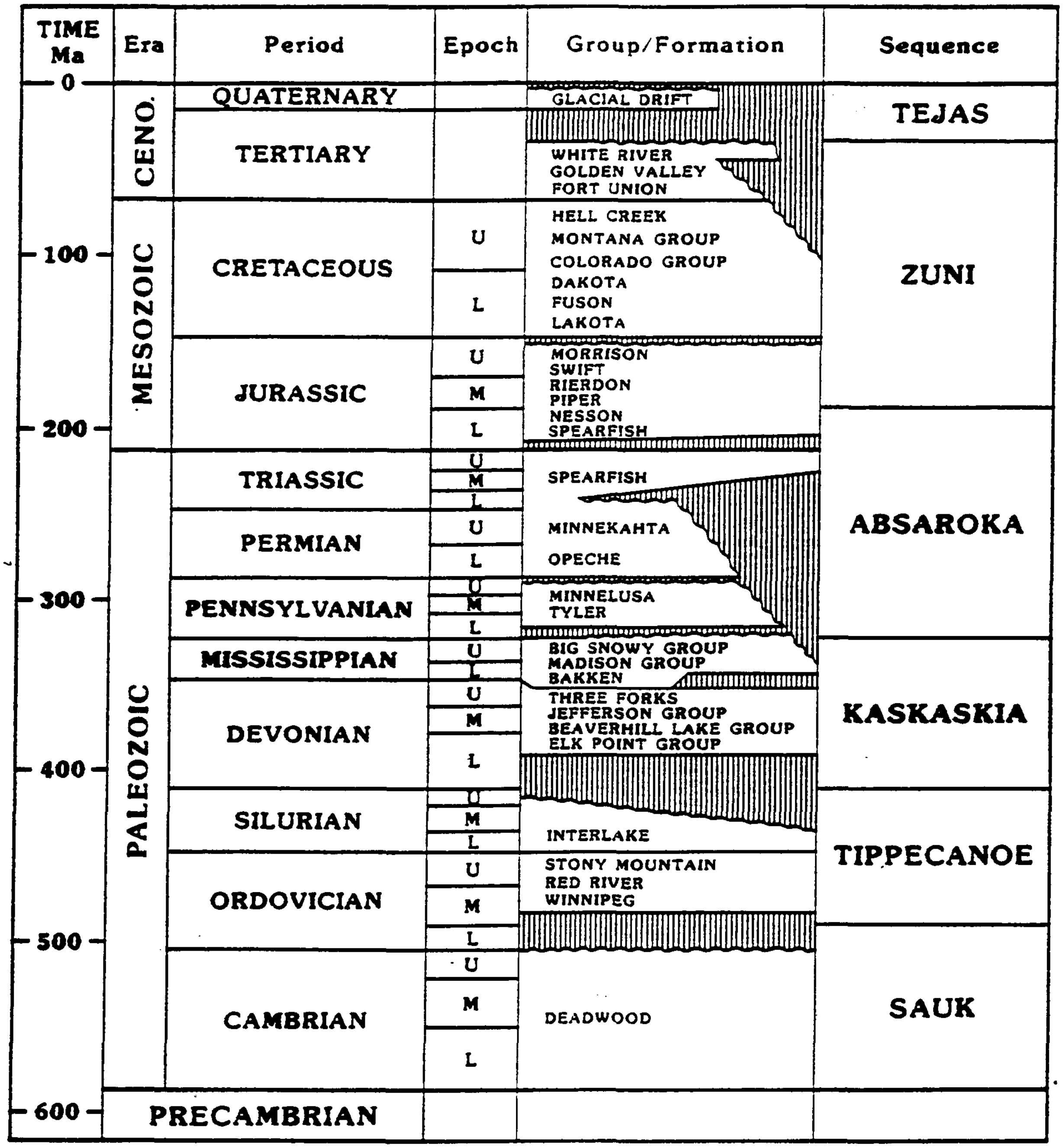

Figure 3: Stratigraphic column of the Williston basin in North Dakota. Adapted from Haid (1991).

This conclusion is based in part on the fact that in the western Canada sedimentary basin, the preserved thickness of the Sauk sequence forms a simple pericratonic wedge (Porter *et al.* 1982). Alternatively, Fowler and Nisbet (1985), Lefever *et al.* (1987), Haid (1991) suggest that the basin's subsidence started with the deposition of the Sauk sequence. The Sauk sequence which is represented in the stratigraphic record of the Williston basin by the Deadwood Formation was mapped in more detail in North Dakota by Lefever *et al.* (1987). These authors analyzed stratigraphic logs and they have shown that the Deadwood Formation thickens (from zero to more than 270 m) towards the center of the basin. They also estimated the tectonic subsidence by using a backstripping procedure. From the isopach maps and tectonic subsidence curves, Lefever *et al.* (1987) concluded that the basin was initiated during deposition of the Member C of the Deadwood formation (at ca 505 Ma). However, the authors note some uncertainty for the ages of the strata of the Deadwood Formation; the ages assigned in the study were obtained by extrapolation with correlative rocks in Montana and North Dakota. The age of the basement rocks beneath the Williston basin estimated by fission-track dating of apatite is about 554 Ma (Crowley *et al.* 1985). The biotite Rb-Sr dating of basement rocks from near the center of the Williston basin gives an age of 520± 30 Ma (Peterman and Hedge, 1964). These ages would also support the idea of a thermal event at the Early-Mid-Cambrian time and consequently that the subsidence in the Williston basin started in Late Cambrian.

The termination of the basin's history has also been in question. It is not clear when the local subsidence of the Williston basin ceased and the basin became part of the foreland basin developing to the east of the Cordilleran foreland thrust belt. The evolution of the foreland Alberta basin was in response to shortening across the Columbian Orogen which translated thrust sheets to the east from the Rocky Montains (Beaumont, 1981). The sediment deposition of the Alberta basin began in the Late Jurassic time (at ca 150 Ma) with the deposition of the Zuni sequence (Beaumont, 1981; Porter *et al.* 1982). An unconformity in the stratigraphic record occurs in the Williston basin from the Late Jurrasic time (at ca 150 Ma) to the Early Cretaceous (at ca 120 Ma). A new increase in subsidence rate began in the Williston basin at ca 120 Ma and continued until the Paleocene time. Furthermore, a change in sedimentary regime in Cretaceous time was remarked by Caldwell (1986), who noted that in Cretaceous time the terrigenous clastic sediments were derived from the Cordillera. It appears that the Williston basin became part of the foreland basin in Cretaceous time (at ca 120). The isopach of 2 km (fig.1), demonstrating a regional trend in the sediment deposition, corresponds to an age of 120 Ma.

Consequently, this study assumes that the local Williston basin's subsidence lasted about 370 Myrs: it started in Cambrian (at ca 520 Ma) and ended to the Late Jurrasic (at ca 150 Ma). During the local subsidence, the total sediment accumulation (2.8 km) remains modest in comparison with the Michigan basin, where about 3.6 km of the sediments were accumulated in only 160 Myrs.

2.2 Crustal structure

In 1977, 1979, and 1981, COCRUST (Consortium for Crustal Reconnaissance Using Seismic Techniques) conducted seismic refraction profiles in the Canadian portion of the Williston basin (southern Saskatchewan and Manitoba). These studies showed that the Williston basin is characterized by a very thick crust and high velocities in the lower crust and the upper mantle.

Hajnal *et al.* (1984) determined a maximum crustal thickness of 54 km for the central portion of the Williston basin. Morel-à-l'Huissier *et al.* (1987) have found that the thickness of crust under the basin varies from 41 km (in southern Manitoba) to 48 km (in southwestern Saskatchewan) with a small (about 3-4 km) thinning of the crust near the center of the basin. Latham *et al.* (1988) reported the results of COCORP deep seismic reflection profiles in northeastern Montana where they estimated the crustal thickness to be 48 km.

The collision of continental fragments during Hudsonian orogeny with the line of collision under the Williston basin, was suggested by Lewry, (1981) and by Green *et al.* (1985). The very thick crust under the Williston basin most probably results from this tectonic event, when the crustal thickness was doubled before being eroded to present level (Fowler and Nisbet, 1985).

Hajnal *et al.* (1984) have determined that P_n velocity was 8.3-8.5 $km.s^{-1}$ below the central part of the basin. Recent interpretation of Morel-à-Huissier *et al.* (1987) suggests that mantle velocities vary from ~ 8 $km.s^{-1}$ to 8.4 $km.s^{-1}$; the authors interpret this variation in terms of mantle anisotropy associated with the trend of the Trans-Hudson orogen. The analysis of Mereu *et al.* (1989) suggests that the Moho under the basin is not a very sharp boundary. Some of the interpretations indicate a high-velocity lower crustal layer (7.0 to 7.6 $km.s^{-1}$) across the Williston basin (Hajnal *et al.* 1984; Morel-à-Huissier *et al.* 1987; Latham *et al.* 1988); these velocities are typical of garnet-granulite (Christensen, 1982).

The Bouguer gravity anomaly corrected for the sedimentary fill is positive and shows a regular circular pattern concentric with the basin. Datonji (1981) analyzed this positive anomaly and concluded that it is caused by a load located in the lower crust.

2.3 Models for the Williston basin

The cause for the long subsidence history of the Williston basin has not been elucidated yet. Multiple thermal events, particular initial conditions (in assuming for example, a disk-shaped or a plugs heat source), or a gabbro-eclogite metamorphic transformation are among some tentative explanations for the complex subsidence record. Some among these hypotheses have not been tested or seem inadequate for the Williston basin's evolution.

The idea of multiple thermal events is rejected by the recent studies. The subsidence data from several wells of the Williston basin were analyzed by Haid (1991), who concluded that the basin's subsidence was continuous from the Cambrian to Cretaceous time. Gosnold and Sweeney (1992) suggested a constant paleo-heat flow without thermal input during the past 350 Myrs of the basin's history. This requires that the basin was formed by a single thermal event.

In the mechanical and thermal model of Ahern and Mrkvicka (1984), the subsidence of the Williston basin followed the cooling of a disk-shaped heat source which is 400 km in diameter, 89 km thick, 89 km below the surface, 89 km above the asthenosphere (i.e. the thickness of the lithosphere is 267 km) and initially at a temperature excess of 153°C. The thermal time constant of 225 Myrs was predicted by the model. The early stage (about the first 50-75 Myrs) of the basin's history is not considered; the subsidence curve start in Late Ordovician (at ca 450 Ma) and not in Late Cambrian (at ca 520 Ma) as suggested by the record. Ahern and Mrkvicka (1984) have compared the calculated subsidence with the sediments thickness data from Zabolotony 1-3-4a well. The model does not fit well the subsidence record. This poor fit may be because sea-level was not included in the subsidence calculation or because the mechanism is inadequate.

The formation of the Williston basin by a phase change mechanism was discussed by several authors (Datonji, 1981; Quinlan, 1987; Ahern and Mrkcvicka, 1984; Fowler and Nisbet 1985; Crowley *et al.* 1985; Jerome, 1988; Haid, 1991).

A process for the phase change to trigger subsidence was already proposed by Haxby *et al.* (1976) for the Michigan basin: their model invokes the intrusion of a mantle diapir which heats the lower crust and permits the transformation of metastable garnet-granulite into eclogite under the Michigan basin. Datonji (1981) suggested that the same mechanism caused the formation of the Williston basin.

The cause of the acceleration of the subsidence in the Williston basin is unclear. Several suggestions were made to explain the increase of the subsidence rate in intracratonic basins: eustatic-sea level variations, fluctuations in intraplate stress fields, regional compressive stress during the periods of tectonism, the deep mantle-convection model for the assembly of the supercontinents (Sleep, 1976; Cloetingh, 1988; DeRito *et al.* 1983; Kominz and Bond, 1991). However, it was remarked by Fowler and Nisbet (1985) that the unconformities in the sedimentary record of the Williston basin are correlated with drop in eustatic sea-level. In fact, all periods of subsidence acceleration in the Williston basin except for the Middle Ordovician (at ca 470 Ma), which occurred in early stage of the basin's formation, correlate with a rise of eustatic sea-level. This may indicate that another phenomenon is responsible for the Middle Ordovician acceleration of the subsidence. This phenomenon may be the phase change transformation, because it is delayed relative to the effect of thermal contraction.

The absence of igneous intrusions in the stratigraphic column of the Williston basin (Porter *et al.* 1982) implies that the basin was initiated by a deep heat source located in the mantle lithosphere or at the lithosphere-asthenosphere boundary. The deep thermal event may be related to dynamic upwelling or diapirism of asthenosphere into lithosphere (Crowley *et al.* 1985; Datonji, 1981) or a heating of the lithosphere from below (Crough and Thompson, 1977a,b). The heating of the lithosphere by increased heat flow at the LAB may occur when it moves over a hot spot (Crough and Thompson, 1977a,b).

Heating of the lithosphere can also trigger phase transformations in the lower crust and/or upper mantle. Subsequent cooling may be accompanied by the reverse transformations. This phenomenon may have occurred under the Williston basin; the cooling was accompanied by a phase change in the lower crust and/or in the upper mantle. Although the seismic, if the high-velocity mantle under the Williston basin is causally related or preceded the basin's formation. It will be assumed that high-velocity lower crustal layer, probably corresponding to garnet-granulite facies, is related to the basin's origin.

The positive Bouguer gravity anomaly associated with the mass excess located in the lower crust (Datonji, 1981) supports, the suggestion that phase transition from gabbro into garnet-granulite took place in the lower crust under the Williston basin. Furthermore, the migration of the geometrical center of the basin by 20 km over 450 Myrs (Ahern and Mrkvicka, 1984) and the concentric and concordant isopachs may indicate that the subsidence was caused by an increasing load located near the center of the basin, rather than by other tectonic process (Fowler and Nisbet, 1985). In addition, the gabbro-garnet-granulite transformation causes an acceleration of the tectonic subsidence in the early stage of the basin's evolution and would thus explain the Middle Ordovician pulse of subsidence in the Williston basin.

3. Tectonic subsidence following a sudden heat flow change at the base of the lithosphere.

For the combined thermal contraction-phase transformation model, the tectonic subsidence is obtained as the superposition of the thermal subsidence, caused by contraction of the cooling lithosphere, and the subsidence induced by the phase transformation. Both processes are induced by changing thermal boundary conditions at the lithosphere asthenosphere boundary. The thermal and phase change subsidence were calculated for a sudden heat flow change at the base of the lithosphere.

Thermal subsidence follows the decay of a transient thermal perturbation. The temperature perturbation, Θ_1, is the solution of the 1-D heat conduction equation (Carslaw and Jaeger, 1959):

$$\frac{\partial \Theta_1}{\partial t} = \kappa \frac{\partial^2 \Theta_1}{\partial z^2} \tag{1}$$

with appropriate initial and boundary conditions. κ is the thermal diffusivity, t is time, z is the vertical coordinate (defined positive downward), z = 0 is the surface, z = l is the lithosphere asthenosphere boundary. A list of symbols and the values assumed for some important parameters is given in Table 1.

If thermoelastic effects are neglected (i.e. thermal contraction is in the vertical direction), the thermal subsidence, S_o, is obtained as:

$$S_0(t) = \alpha \int_0^l \{\Theta_1(z, t=0) - \Theta_1(z,t)\} dz \tag{2}$$

where α is the thermal expansion coefficient and l is the thickness of the lithosphere.

Table 1 : List of symbols and values of parameters.

SYMBOL	DEFINITION	VALUE
c	specific heat	700 $J.kg^{-1} K^{-1}$
g	acceleration of gravity	9.8 $m.s^{-2}$
K	thermal conductivity	2 $W.m^{-1} K^{-1}$
l	thickness of the lithosphere	100-250 km
L	latent heat	40 $J.g^{-1}$
P_o	surface pressure excitation	
P_m	pressure of thermodynamic equilibrium	
Q_l	amplitude of heat flow change at LAB	3-10 $mW.m^{-2}$
s	variable of Laplace transform	
S_o	thermal subsidence	300-600 m
S_l	phase change subsidence	300-3000 m
T_o	surface temperature adjusted for crustal heat sources	
T_c	temperature of equilibrium at surface	
T_m	temperature of thermodynamic equilibrium	
z_o	initial depth of the phase boundary	35-60 km
α	coefficient of thermal expansion	$3\cdot10^{-5} K^{-1}$
β	geothermal gradient	5-8 $K.km^{-1}$
γ	inverse slope of the Clapeyron line	0.5-0.7 $K.MPa^{-1}$
Δz_0	amplitude of phase boundary mouvement	7-33 km
Δz_m	displacement interface between two phase	
$\Delta\rho$	density contrast between two phase	0.2 $Mg.m^{-3}$
$\Delta\rho/\rho$	relative density contrast	0.07
Θ_1	thermal perturbation caused by the changing boundary condition	
Θ_2	thermal perturbation caused by latent heat	
κ	thermal diffusivity	$10^{-6} m^2.s^{-1}$
ρ	density of the phase being transformed	3.0 $Mg.m^{-3}$
ρ_m	mantle density	3.3 $Mg.m^{-3}$
ρ_s	sediments density	2.6 $Mg.m^{-3}$
τ	relaxation time for phase boundary motion	1 Myr
τ_o	z_0^2/κ	
τ_l	l^2/κ	200-600 Myrs

After a sudden change in heat flow, Q_l, at the LAB, the thermal subsidence, $S_o(t)$, is obtained as (Mareschal, 1981):

$$S_0(t) = \frac{\alpha Q_l l^2}{2K}\left\{1 - \frac{32}{\pi^3}\sum_{n=0}^{\infty}(-1)^n \frac{\exp\{-(2n+1)^2\pi^2 t/4\tau_l\}}{(2n+1)^3}\right\} \tag{3}$$

where $\tau_l = l^2/\kappa$ is the conduction time for the lithosphere.

The amplitude of the thermal subsidence does not depend on the boundary condition but only on the amplitude of the thermal change at the LAB. A small change in heat flow, 1 $mW.m^{-2}$, causes about 74 m of thermal subsidence (assuming that the lithosphere is 100 km thick, the thermal conductivity $K = 2\ W.m^{-1}K^{-1}$, and the thermal expansion coefficient $\alpha = 3 \cdot 10^{-5}K^{-1}$.

The leading term in the series (3) is $\exp(-\pi^2 t/4\tau_l)$, while the leading term in the series solution for a sudden temperature change at the LAB is $\exp(-\pi^2 t/\tau_l)$ (Hamdani *et al.* 1991). The decay constant of the thermal transient will thus be 4 times longer for a heat flow boundary condition than for a temperature boundary condition (i.e. 200 Myrs vs 50 Myrs for a 100 km thick).

The temperature T_m and the pressure P_m of the two phases in equilibrium are related by the integrated Clausius-Clapeyron equation:

$$T[z_m(t), t] = T_c + \gamma P[z_m(t), t] = T_c + \gamma[g\rho z_m(t) + P_0(t)] \tag{4}$$

where z_m is the depth of the phase boundary, T_c is the temperature of equilibrium at the surface, γ is the inverse slope of the Clapeyron line, g is the acceleration of gravity, P_o is the surface pressure due to sediment loading, and ρ is the density of the crust.

The intersection of the geotherm with the Clausius-Clapeyron curve determines the boundary between the two phases. When the lithosphere cools, and the geotherm is modified, the intersection point moves and determines the movement of the interface between the two phases. When the phase boundary moves up, transformation of the less dense phase into denser phase is accompanied by a subsidence; this phase change induced subsidence, S_1 is obtained as:

$$S_1 = -\frac{\Delta\rho}{\rho}\Delta z_m \tag{5}$$

where $\Delta\rho$ is the density contrast between the two phases, ρ is the density of the phase being transformed, and Δz_m is the displacement of the phase boundary.

The phase transformation causes the release (or absorbtion) of latent heat and a transient thermal perturbation Θ_2, which retards the motion of the phase boundary.

The temperature in the lithosphere is the sum of several components:

$$T(z,t) = T_0 + \beta z + \Theta_1(z,t) + \Theta_2(z,t) \tag{6}$$

where T_o is the surface temperature corrected for shallow radiogenic heat production, β is the geothermal gradient below the heat production layer, Θ_1 is the thermal perturbation caused by changing boundary conditions, and Θ_2 is the thermal perturbation induced by the release of latent heat at the moving boundary.

The phase boundary is determined by the condition of thermodynamic equilibrium:

$$T_m = T_o + \beta z_m(t) + \Theta_1[z_m(t),t] + \Theta_2[z_m(t),t] = T_c + \gamma[g\rho z_m(t) + P_o(t)] \tag{7}$$

The initial depth of the interface is z_o:

$$T(z_o) = T_o + \beta z_o = T_c + \gamma g \beta \rho z_o \tag{8}$$

The phase boundary motion is given by:

$$\Delta z_m(t) = z_m(t) - z_0 = \frac{\Theta_1(z_m(t),t) + \Theta_2(z_m(t),t) - \gamma P_0(t)}{(\gamma g \rho - \beta)} \tag{9}$$

where $\gamma g \rho - \beta$ is the difference between the Clausius-Clapeyron and the geothermal gradients

The equation (9) is a non-linear integral equation because the heat source and the point where temperature is calculated are both moving. A linear approximation was obtained by Mareschal and Gangi (1977a) for small displacement of the phase boundary $z_m(t) \approx z_0$ (i.e. the phase boundary motion is small compared with the depth of the interface).

The phase boundary motion is calculated as a series expansion of which the leading terms are given by (see Appendix A):

$$\frac{\Delta z_m(t)}{\Delta z_o} = 2\sqrt{\frac{t}{\pi\tau_o}}\exp\left(\frac{-\tau_o'}{4t}\right) + \left(\sqrt{\frac{\tau}{\tau_o}} + \sqrt{\frac{\tau_o'}{\tau_o}}\right)\operatorname{erfc}\left(\frac{1}{2}\sqrt{\frac{\tau_o'}{t}} + \sqrt{\frac{\tau}{\tau_o}}\right)\ldots$$

$$+\exp\left(\frac{t}{\tau} + \sqrt{\frac{\tau_o'}{\tau}}\right)\operatorname{erfc}\left(\frac{1}{2}\sqrt{\frac{\tau_o'}{t}} + \sqrt{\frac{t}{\tau}}\right) \tag{10}$$

where the amplitude of the phase boundary movement $\Delta z_o = Q_l z_o / K(\gamma g \rho - \beta)$ and the time constants $\tau_o = z_o^2/\kappa$ and $\tau_o' = (l - z_o)^2/\kappa$.

The amplitude of the subsidence induced by the phase change is directly proportional to the amplitude of the thermal change at LAB and inversely proportional to the difference between the Clausius-Clapeyron and the geothermal gradient. A drop of 1 $mW.m^{-2}$ in heat flow at the LAB could cause the phase boundary to move up by 3840 m and results in 269 m of subsidence (assuming that the lithosphere is 100 km thick, the depth of the phase boundary is 45 km, γ=0.5 $K.MPa^{-1}$, β=8 $K.km^{-1}$, the difference in slope between the geotherm and the Clausius-Clapeyron is 7 $K.km^{-1}$ and the relative density change is 7%).

Equation (10) implies that the phase boundary subsidence is negligible for $t < (l - z_o)^2/\kappa$ (i.e. as long as the thermal perturbation has not reached the phase boundary). The subsidence induced by the phase transformation lags behind the thermal subsidence. The delay depends not only on boundary conditions, but also on the distance between the boundary and the initial position of the phase boundary. For example, for a sudden drop in flow at LAB it is on the order of 10 Myrs for l=100 km, z_o=45 km and 40 Myrs for l=150 km, z_o=45 km.

The time constant τ represents the time necessary for the latent heat to diffuse away from the phase boundary; it determines the maximum velocity of the phase boundary. It is given by:

$$\tau = \left\{ \frac{L}{2\sqrt{\kappa}\,c(\gamma g \rho - \beta)} \right\}^2 \tag{11}$$

where L is the latent heat, and c is the specific heat.

For the gabbro-garnet-granulite transition γ is on the order 0.5-0.7 $K.MPa^{-1}$ (Ito and Kennedy, 1971; Ahrens and Schubert, 1975). The latent heat released from this transformation is on the order 40 $J.g^{-1}$ (assuming that the temperature of the transition is 1000 K and the relative density change is 7%).

Mareschal and Gangi (1977a) have estimated τ to be on the order of 1 Myr for the gabbro-eclogite transition or less which is much smaller than τ_o and τ_o'. For the gabbro-garnet granulite transition τ is on the order 0.6 Myrs for $(\gamma g\rho - \beta)$= 6 K.km^{-1} and 0.08 Myrs for $(\gamma g\rho - \beta)$=10 K.km^{-1}. This implies that it is the rate of temperature change and not the rate of latent heat release that controls the phase boundary motion and that for all practical purposes the effect of the latent heat can be neglected. In that case, the phase boundary motion is given by:

$$\frac{\Delta z_m(t)}{\Delta z_0} = 1 - \frac{8l}{\pi^2 z_0} \sum_{n=0}^{\infty} (-1)^n \frac{\sin((2n+1)\pi z_0/2l)}{(2n+1)^2} \exp(-(2n+1)^2\pi^2 t/4\tau_l) \qquad (12)$$

Figure 4 shows the total tectonic subsidence after a change in heat flow at the LAB. Time is relative to τ_l; the total subsidence is compared with the thermal subsidence ($\alpha Q_l l^2/2K$). On Figure 4a, the depth of the phase boundary (i.e. τ_o/τ_l) is constant and the curves are calculated for different ratios of phase change to thermal subsidence. On Figure 4b, the subsidence induced by the phase change is assumed equal to the thermal subsidence and the curves are calculated for different depths to the phase change (i.e τ_o/τ_l).

The tectonic subsidence is amplified by two processes: (1) isostatic adjustments, and (2) additional migration of the phase boundary due to the weight of the sediment load. For a one-dimensional model, flexural effects are neglected and isostatic adjustments are directly included in the calculations through an isostatic amplification coefficient. For typical values of sediment and mantle density, the isostatic amplification is about 4.7.

In the linear approximation, different causes of the phase boundary motion can be superposed. Consequently, the effects of pressure and temperature on the phase boundary

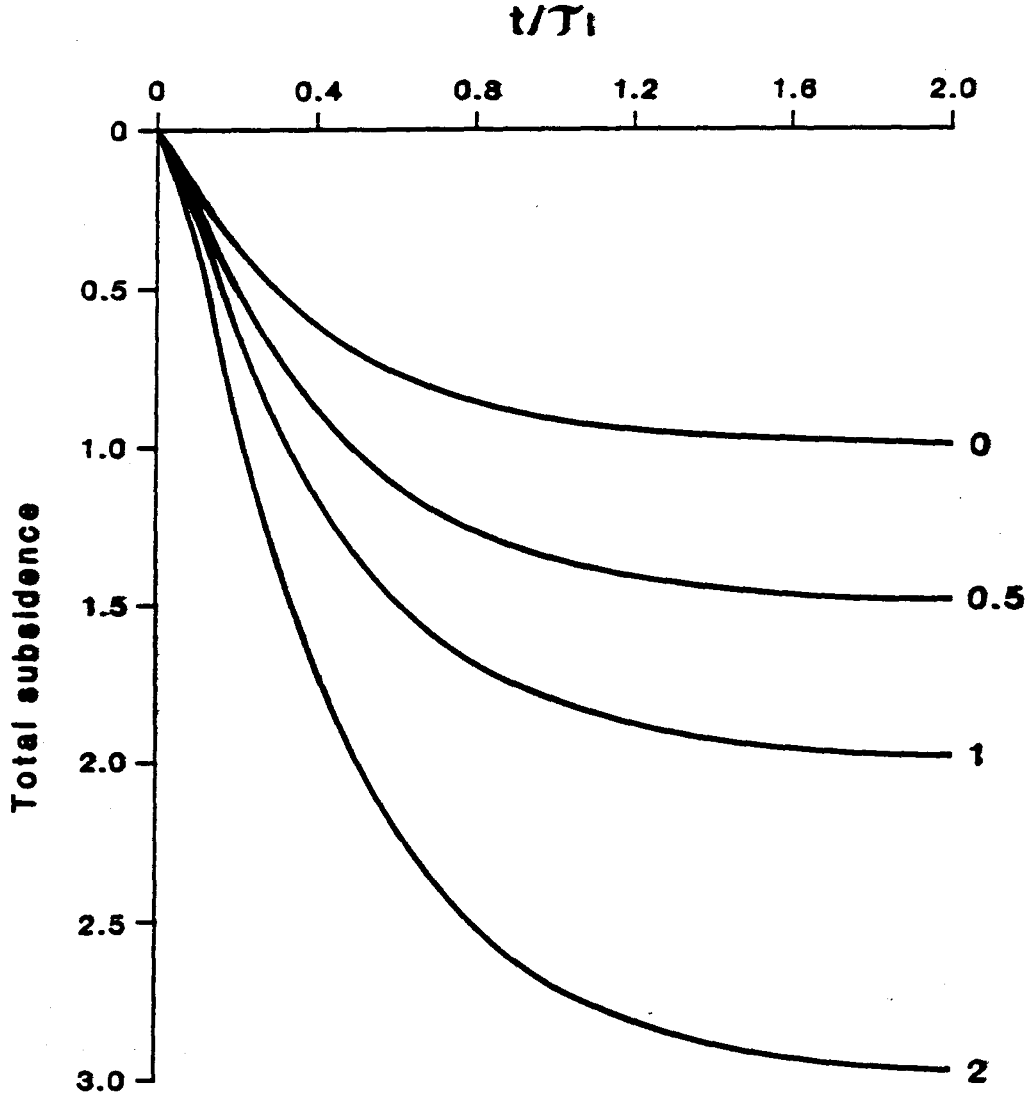

Figure 4a: Total subsidence after a change in heat flow at the LAB. The total subsidence is normalized to the amplitude of thermal subsidence. Time is relative to l^2/κ. The different curves correspond to different ratios of phase change to thermal subsidence. $\tau_o/\tau_l = 0.1$

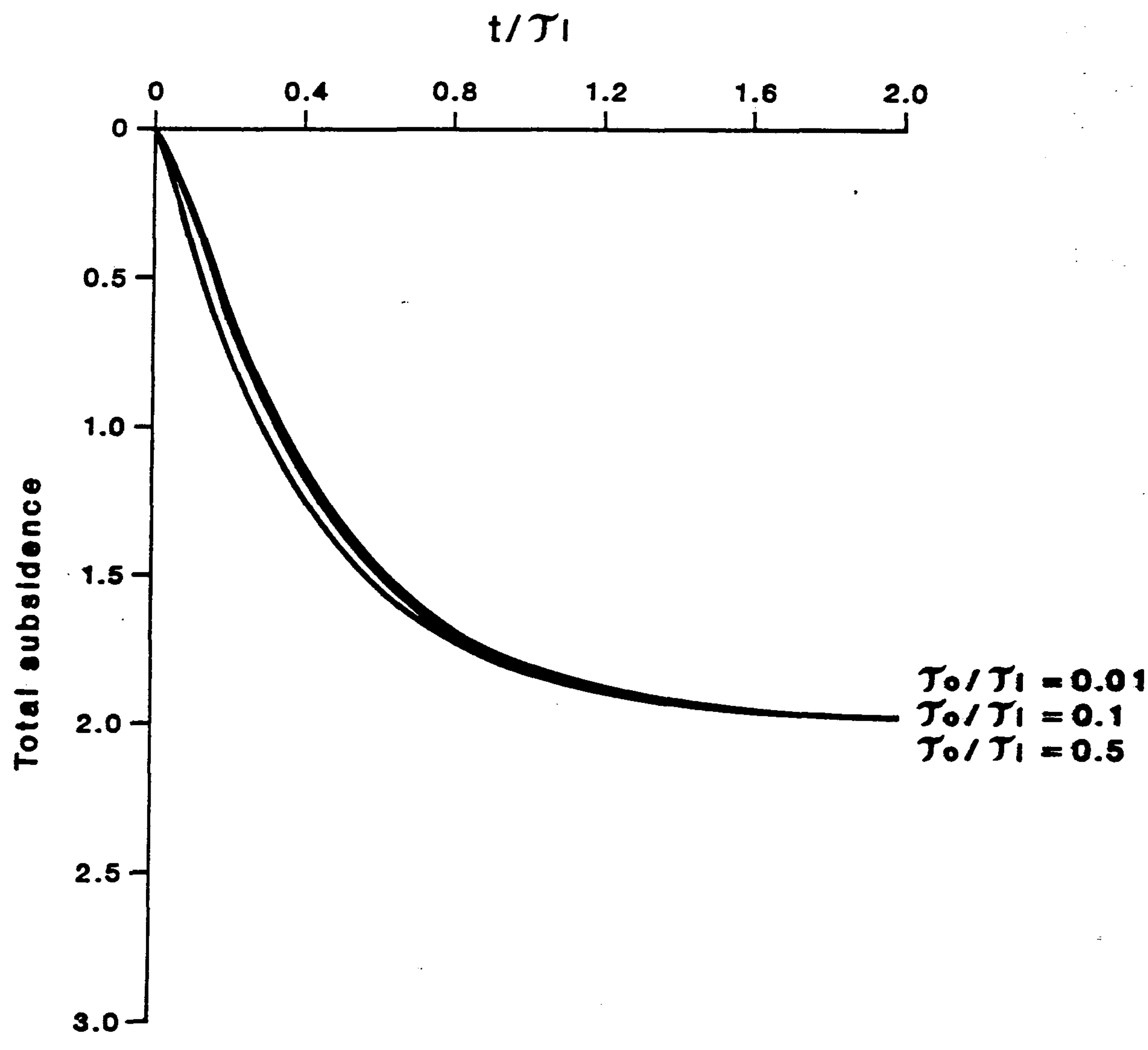

Figure 4b: Total subsidence after a change in heat flow at the LAB. The total subsidence is normalized to the amplitude of thermal subsidence. Time is relative to l^2/κ. The different curves correspond to different values of initial depth to phase boundary (i.e. τ_o/τ_l).

were superposed. The motion of the phase boundary following a change in pressure at constant rate (i.e. constant rate of sediment deposition) is obtained by Hamdani *et al.* (1991) It gives:

$$\Delta z_m(t) = \frac{-\gamma \overline{P}'_o \tau}{(\gamma g \rho - \beta)} \left\{ 1 - 2\sqrt{\frac{t}{\pi\tau}} + \frac{t}{\tau} - \exp(t/\tau)\,\mathrm{erfc}(\sqrt{t/\tau}) \right\} \tag{13}$$

If the inverse slope of the Clausius-Clapeyron $\gamma = 0.5\,\mathrm{K.MPa^{-1}}$, and the difference in slopes between the Clausius-Clapeyron and the geotherm is 7 $\mathrm{K.km^{-1}}$, this effect is not very large: the phase boundary migrates by 2 m for 1 m of sediment deposited and causes 14% additional subsidence.

4. Analysis of the models proposed for the Williston basin

The cooling of the lithosphere and the subsidence induced by phase boundary movement following a drop in temperature at the LAB adequately explained the evolution of the Michigan basin, where 3.6 km of sediments accumulated in relatively short time of about 160 Myrs.

The longer subsidence history of the Williston basin requires that either: (1) cooling of the lithosphere followed a drop in heat flow at LAB, or (2) cooling of the lithosphere resulted from reduced temperature at the LAB but the lithosphere was much thicker that 150 km. Calculations were performed to test these two hypotheses.

For a temperature at the base of the lithosphere equal to 1500°C, the average geothermal gradient in the crust was assumed to be 10 $\mathrm{K.km^{-1}}$ for l=250 km and 14 $\mathrm{K.km^{-1}}$ for l=150 km. The gradients in the mantle are 5 $\mathrm{K.km^{-1}}$ for l= 250 km, and 8 $\mathrm{K.km^{-1}}$ for l=150 km, corresponding to mantle heat flow equal to 10 $\mathrm{mW.m^{-2}}$ and to 16 $\mathrm{mW.m^{-2}}$ (with

$K= 2W.m^{-1}K^{-1}$) at the time of the basin's initiation. Figure 5 shows the subsidence history calculated for a temperature drop at the LAB. The model assumes that tectonic subsidence is caused by the thermal contraction and the transformation of gabbro into garnet-granulite in the lower crust. A temperature drop of 193 K is assumed to have occurred at 512.5 Ma. The initial thickness of the lithosphere was assumed 250 km, the geothermal gradient 5 $K.km^{-1}$, the difference between Clausius-Clapeyron and geothermal gradient 11.5 $K.km^{-1}$ (with γ=0.55 $K.MPa^{-1}$) and the initial depth of the phase boundary 48 km. Airy isostasy with isostatic amplification coefficient 4.7 was assumed, corresponding to sediment density $\rho_s = 2.6$ $Mg.m^{-3}$, and mantle density $\rho_m = 3.3$ $Mg.m^{-3}$. Eustatic effect were included and the sea-level variations determined by Vail and Mitchum (1979) were used to calculate the sediment accumulation; sediment deposition occurs only when the basin is below sea-level; during periods where sea-level drops below the basin's surface there is no accumulation, but erosion is not included in the calculation. The calculations were made for different amplitudes in sea-level changes: the "standard" amplitudes, 1.5 times the standard amplitude, and 0.5 times the standard amplitude. The calculations show that, the return to the thermal equilibrium is completed in 350 Myrs, the phase change subsidence was retarded by about 75 Myrs and produced an acceleration of subsidence at 440 Ma (fig.5). About one third of the tectonic subsidence is caused by gabbro-garnet-granulite transformation. The subsidence data of Zabolotony 1-3-4a well, located near the center of the basin can be compared with the calculations. The results do not fit well the initial subsidence data, because they show that about 2.2 km of sediment accumulated rapidly in the 100 Myrs following basin initiation. The model predicts correctly the total sediment thickness but not their ages.

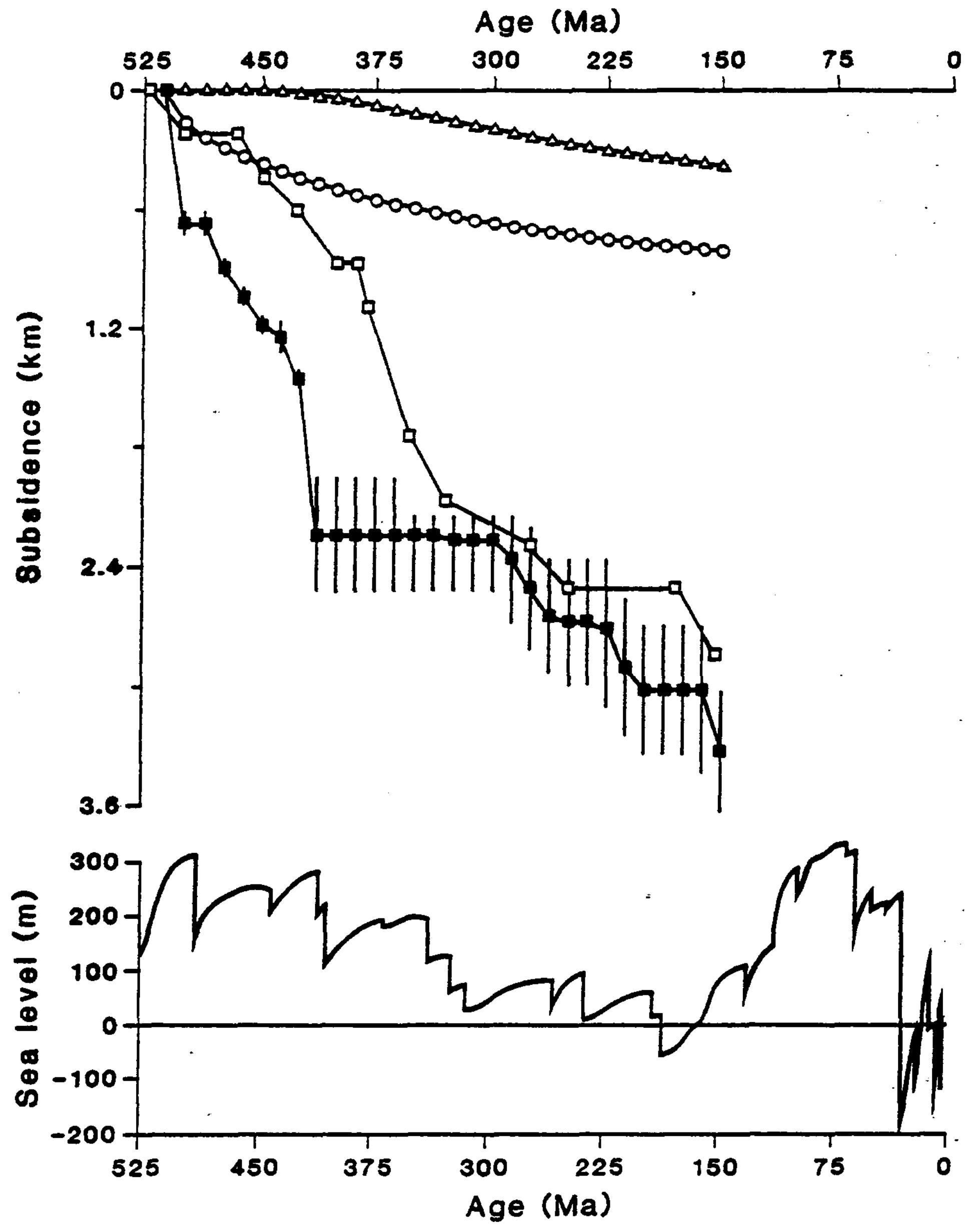

Figure 5: Calculated subsidence for the thermal contraction and phase change following a sudden temperature drop (193 K) at the LAB (250 km). The calculated subsidence is compared with the subsidence record. Sea-level variations are shown below. Open triangles represent the phase change induced subsidence (without isostatic amplification), open circles the thermal subsidence, plain squares the calculated total sediment accumulation including isostasy and sea-level changes, and open squares the recorded subsidence. The error bars on the calculated subsidence shows the effect of increasing or decreasing by 50 per cent the amplitude of the sea-level variations

The temperature boundary condition does not account for the Williston's basin evolution because the subsidence starts rapidly regardless of the initial lithospheric thickness; only the duration of the subsidence will be affected by the initial lithospheric thickness (Mareschal, 1981). In addition, the initial thicknesss of the lithosphere 250 km is in contradiction with the assumption that initially hot lithosphere cools down. Also, such a model requires the final mantle heat flow to be low (about 8.1 $mW.m^{-2}$), lower than suggested for the Canadian Shield by the heat flow data (about 12 $mW.m^{-2}$ from Pinet *et al.* 1991).

A change in heat flow at the LAB provides a better approximation to the mechanism of lithospheric cooling and the subsidence in the Williston basin. Figures 6a,b compare the subsidence record with the results of two alternative models for the formation of the Williston basin. The tectonic subsidence is caused:

1) by thermal subsidence after a 7.2 $mW.m^{-2}$ heat flow drop at the LAB (fig.6a);

2) by a combination of thermal subsidence and phase transformation of gabbro into garnet-granulite, after a 3.2 $mW.m^{-2}$ drop in heat flow (fig.6b). The initial depth of the phase boundary is 48 km, the relaxation time for the phase boundary, τ, is 1 Myr, the difference between the Clausius-Clapeyron and the geothermal gradient, $\gamma g \rho - \beta$, is 8.5 $K.km^{-1}$ (with γ=0.55 $K.MPa^{-1}$ and β=8 $K.km^{-1}$) and relative density change between the two phases 0.07.

The isostatic amplification and the effect of the sea-level variations on the tectonic subsidence are introduced as above. For both models, the initial lithospheric thickness is assumed 150 km, and the heat flow at the LAB drops at 512.5 Ma. The predicted subsidence could not fit the observations if the time of basin initiation is 523 Ma. This may be because the ages of the strata of the Deadwood formation or the amplitude of the changes of the sea-level in Late Cambrian time are probably overestimated. The same problem was pointed by Haid (1991) during the estimation of the tectonic subsidence.

Figures 6a,b compare the predicted subsidence with the data from well Zabolotony 1-3-4a. The calculations show that, for 150 km thick lithosphere and heat flow boundary condition, the return to thermal equilibrium is completed in approximatively 350 Myrs. The thermal contraction model (fig.6a) does not fit as well the initial stage of the subsidence records. Some acceleration of subsidence occurs at 487.5 Ma as a result of sea-level rise. The combined thermal contraction-phase boundary motion model (fig.6b) explains well the first 175 Myrs of subsidence. The phase boundary motion is retarded by approximatively 40 Myrs and produces the acceleration of subsidence at 470 Ma. After 150 Myrs, the phase change induced subsidence dominates and produces most of the subsidence. Drops in sea-level between 500-475 Ma and 425-400 Ma coincide with unconformities. The acceleration of the subsidence at 387 Ma is caused by sea-level rise. The predicted subsidence between 260-180 Ma is greater than observed; this is probably because the effect of erosion was not included.

5. Discussion and conclusion.

The duration of the subsidence is determined by the initial lithospheric thickness and by the boundary condition. For a drop in heat flow at the LAB (at 150 km), the lithosphere returns to equilibrium in approximatively 350 Myr.

The amplitude and duration of Williston basin subsidence is explained relatively well by a single episode of cooling following decrease in heat flow at the LAB.

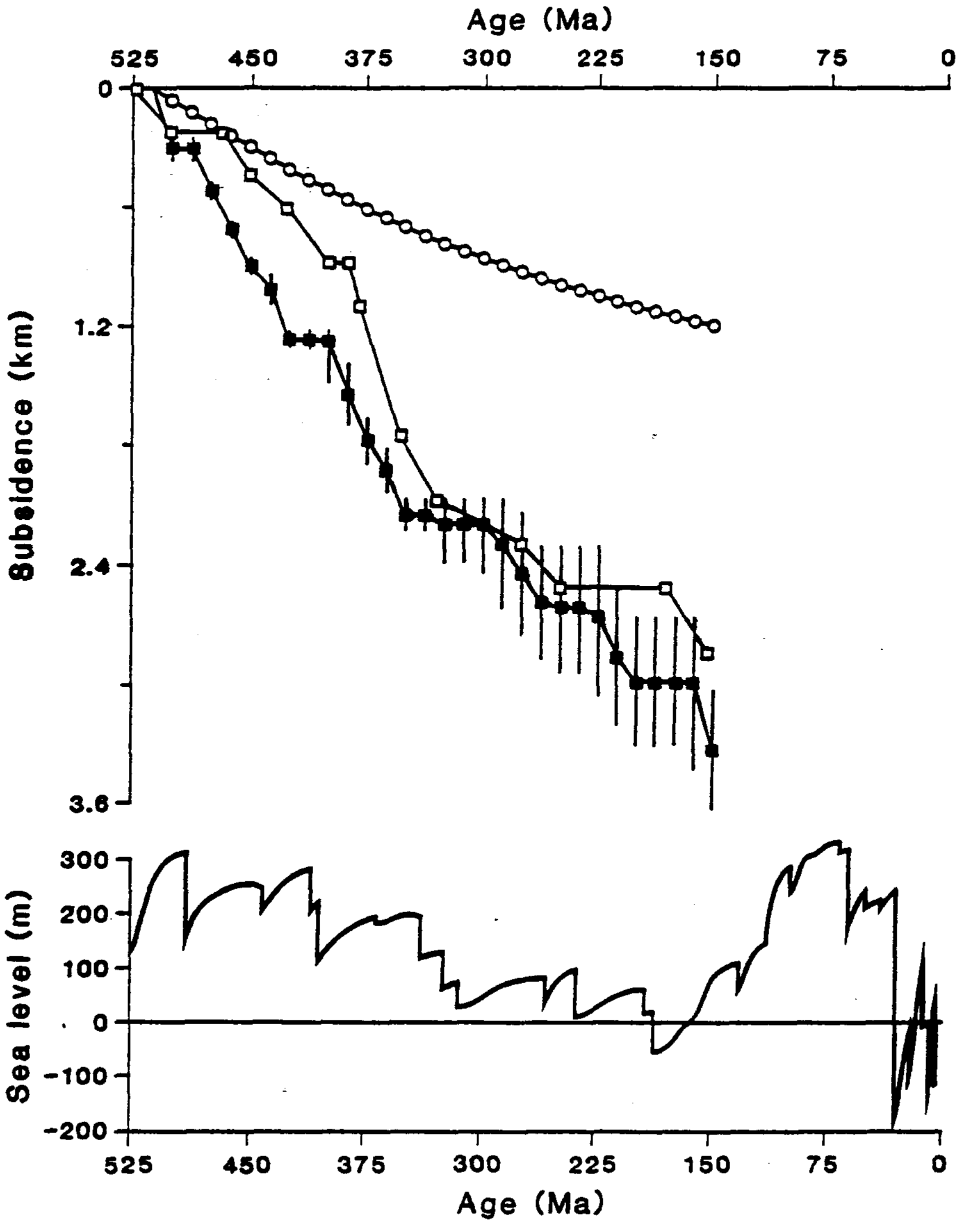

Figure 6a: Calculated subsidence for thermal contraction following a 7.2 mW.m^{-2} heat flow drop at the LAB (150 km). The calculated subsidence is compared with the subsidence record. Sea-level variations are shown below. Open circles represent the thermal subsidence, plain squares the calculated total sediment accumulation including isostasy and sea-level changes, and open square the recorded subsidence. The error bars on the calculated subsidence shows the effect of increasing or decreasing by 50 per cent the amplitude of the sea-level variations.

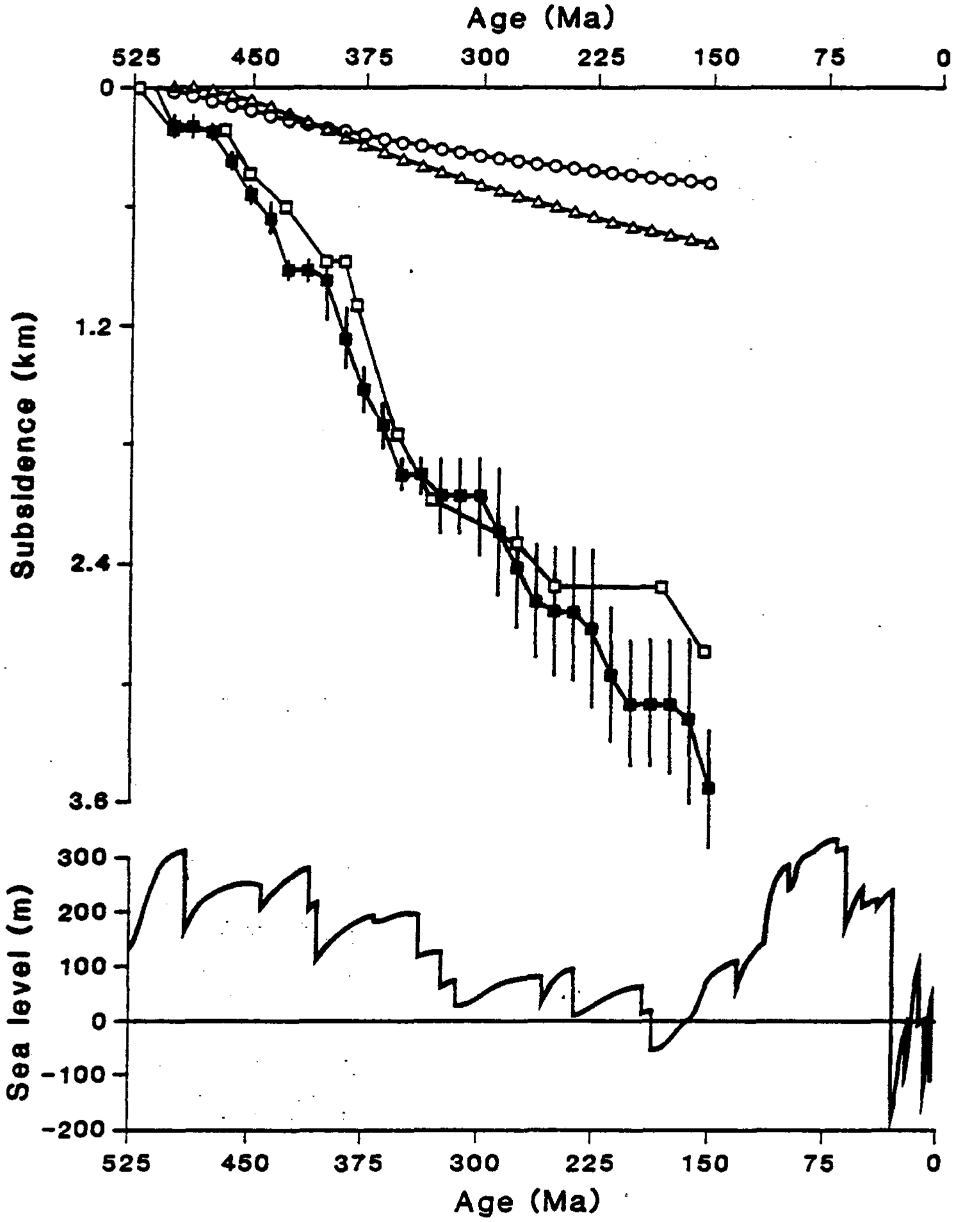

Figure 6b: Calculated subsidence for thermal contraction and phase change following a 3.2 $mW.m^{-2}$ heat flow drop at the LAB (150 km). The calculated subsidence is compared with the subsidence record. Sea-level are shown below. Open triangles represent the phase change induced subsidence (without isostatic amplification), open circles the thermal subsidence, plain squares the calculated total sediment accumulation including isostasy and sea-level changes, and open squares the recorded subsidence. The error bars on the calculated subsidence shows the effect of increasing or decreasing by 50 per cent the amplitude of the sea-level variations.

A simple thermal contraction following a substantial heat flow change at the LAB (about 7.2 $mW.m^{-2}$) could also explain the gross features of the subsidence record of the Williston basin. However, the heat flow change must be large and implies a 367 K drop in temperature at the LAB; larger than the 100-250 K estimated difference between hot plumes and normal mantle (White and McKenzie, 1989; Sleep, 1990; Griffiths and Campbell, 1991). Furthermore, such a drop in heat flow implies that the final mantle heat flow in the Williston basin is about 8.8 $mW.m^{-2}$, lower than most mantle heat flow estimates.

A mechanism involving a lower crustal phase transformation does not require excessive temperature change in the lithosphere. The calculations demonstrate that:

1) For the parameters chosen in the model, the delay of the phase change subsidence is about 40 Myrs.

2) 150 Myrs after the basin's initiation, the phase change becomes the major cause of tectonic subsidence.

3) Different boundary conditions (heat flow vs temperature drop at the LAB) explain the longer subsidence history in the Williston than in the Michigan.

The detailed subsidence record (in particular the first 175 Myrs) is best fitted by a superposition of thermal contraction and phase motion. The model supposes that the local basin's subsidence started about 512.5 Ma and ended to 150 Ma. The unconformity between 500-470 Ma is the result of the moderate subsidence and a drop in sea-level. After that time (at 470 Ma), the phase transformation began and caused an acceleration in subsidence. However, unless the sea-level history and the effect of eustatic changes on sediment deposition history are perfectly understood, it is difficult to draw definite conclusions from the apparent acceleration of subsidence in the early stage of the evolution of the Williston basin. According to the model, the Williston basin's subsidence terminated at 150 Ma. A new episode of the subsidence began at 120 Ma when the basin became the part of the foreland basin cannot considered as the local basin's subsidence.

A 3.2 $mW.m^{-2}$ in heat flow drop at the LAB is consistent with the estimated mean excess heat flow at the hotspot (Davies, 1988a; Sleep, 1990). The model implies that the recent mantle heat flow is 12.8 $mW.m^{-2}$. This is compatible with the thermal structure of the lithosphere suggested for the Canadian Shield by several authors (Drury and Taylor, 1987; Mareschal *et al.* 1989; Pinet *et al.* 1991).

This model is compatible with the seismic structure of the crust. If the phase boundary was initially at the Moho depth (today at 48 km), it has moved up by about 13 km and is located at 35 km. This agrees well with refraction and reflection seismic data that show a high-velocity (7.1-7.6 $km.s^{-1}$) lower crustal layer, between a velocity discontinuity at about 32-40 km depth and the Moho at about 41-48 km depth. The model is also compatible with the gravity data. The density of the garnet-granulite layer explains the positive Bouguer anomaly observed over the Williston basin.

The investigated boundary conditions (heat flow and temperature drop at the LAB) reflect different mechanisms of transferring heat in the lower lithosphere. Liu and Chase (1989) suggested that the mechanisms of heat transfer at the LAB depend on the strength of the mantle plumes; a strong plume is associated with secondary convection and for a weak mantle plume, thermal conduction is dominant. Changing heat flow is associated to conduction. The temperature boundary condition is more appropriate when the lithosphere is being heated by secondary convection in the asthenosphere. The lithospheric material in contact with the plume is engulfed by the plume current as soon as its temperature reaches the temperature of the plume (Yuen and Fleitout, 1985; Liu and Chase, 1989).

We speculate that the lithosphere under the Williston basin was heated by a weak plume and that heat flow at the LAB increased. A stronger plume may have affected the Michigan basin studied previously and heating of the lithosphere by secondary convection caused temperature to rise at the LAB. Subsequent cooling of the lithosphere, when the plate moves away from the plume, produced the drop in temperature or heat flow at the LAB. This waning of the thermal anomaly caused the subsidence of the basins by thermal contraction and the phase changes in the lithosphere. This study demonstrates that the combined effect of these phenomena with an appropriate boundary condition is a plausible mechanism of intracratonic basins's formation.

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Appendix A. Sudden change in heat flow at the LAB.

The solutions for the thermal subsidence and the phase motion following a sudden heat flow change at LAB will be derived. The temperature perturbation Θ_1 following a stepwise change in heat flow Q_l at the LAB is obtained solving the heat equation, with its initial and boundary conditions.

The Laplace transform of the heat equation is:

$$\frac{s}{\kappa}\Theta_1(z,s)=\frac{\partial^2\Theta_1}{\partial z^2}+\frac{\Theta_0(z)}{\kappa} \qquad (A.1)$$

where Θ_0 is the initial condition, which vanishes.

This equation is solved with the following boundary conditions:

$$\Theta_1(z=0,s)=0 \qquad (A.2a)$$

$$K\frac{\partial\Theta_1}{\partial z}(z=l,s)=\frac{Q_l}{s} \qquad (A.2b)$$

The solution is obtained as:

$$\Theta_1(z,s)=\frac{Q_l}{Ks}\sqrt{\frac{\kappa}{s}}\,\frac{\sinh\left(\sqrt{\frac{s}{\kappa}}\;z\right)}{\cosh\left(\sqrt{\frac{s}{\kappa}}\;l\right)} \qquad (A.3)$$

If the thermal contraction is in vertical direction, the Laplace transforme of the thermal subsidence, S_o is given by:

$$S_0(s)=\alpha\int_0^l\left\{\frac{\Theta_1(z,t=0)}{s}-\Theta_1(z,s)\right\}dz \qquad (A.4)$$

Consequently, the thermal subsidence is obtained by integrating (A.3):

$$S_o(s)=\frac{\alpha Q_l l^2}{K\tau_l s^2}\left\{1-\mathrm{sech}\left(\sqrt{\frac{s}{\kappa}\,l}\right)\right\} \qquad (A.5)$$

where $\tau_l = l^2/\kappa$

Inverting the Laplace transform yields (Mareschal, 1981):

$$S_o(t) = \frac{\alpha Q_l l^2}{2K}\left\{1 - \frac{32}{\pi^3}\sum_{n=0}^{\infty}(-1)^n \frac{\exp\{-(2n+1)^2\pi^2 t/4\tau_l\}}{(2n+1)^3}\right\} \tag{A.6}$$

The motion of the phase boundary is determined by the non-linear equation (9). Mareschal and Gangi (1977a) used a linear approximation and reduced this equation to a linear integral equation. The Laplace transform of its solution:

$$z_m(s) = \frac{z_o}{s} + \frac{\Theta_1(z_o,s) + \Theta_2(z_o,s) - \gamma P_o(s)}{(\gamma g \rho - \beta)} \tag{A.7}$$

where Θ_1 is the temperature perturbation caused by the changing boundary condition (A.3) and Θ_2 is the temperature perturbation caused by the release of latent heat by the moving boundary. In the lineare approximation Θ_2 is given by:

$$\Theta_2(z_m,t) \approx \frac{L}{c}\int_0^t \frac{\dot{z}_m(t')dt'}{2\sqrt{\pi\kappa(t-t')}} \tag{A.8}$$

Introducing the solution for $\Theta_1(z_o,s)$ in (A.7) yields the Laplace transform of the phase boundary motion as:

$$\Delta z_m(s) = \frac{Q_l z_o}{K(\gamma g \rho - \beta)s}\frac{1}{\sqrt{\tau_o s}}\frac{1}{\sqrt{\tau s}+1}\frac{\sinh(\sqrt{\tau_o s})}{\cosh(\sqrt{\tau_l s})} \tag{A.9}$$

It can be expanded as:

$$\frac{\Delta z_m(s)}{\Delta z_o} = \frac{1}{\sqrt{\tau_0}}\ \frac{1}{s^{3/2}(\sqrt{\tau s}+1)}\ \sum_{n=0}^{\infty}(-1)^n\{\exp(-((2n+1)l - z_0)\sqrt{s/\kappa})\ldots$$

$$-\exp(-((2n+1)l + z_0)\sqrt{s/\kappa})\} \tag{A.10}$$

where the amplitude of the phase boundary motion $\Delta z_0 = Q_l z_0 / K(\gamma g \rho - \beta)$.

Inversion of the Laplace transform yields:

$$\frac{\Delta z_m(t)}{\Delta z_0} = \sum_{n=0}^{\infty} \exp\left(-\frac{\tau_n'}{4t}\right)\left\{2\sqrt{\frac{t}{\pi\tau_o}} - \frac{\sqrt{\tau}+\sqrt{\tau_n'}}{\sqrt{\tau_o}} w\left(\frac{1}{2}\sqrt{\frac{\tau_n'}{t}}\right) + \sqrt{\frac{\tau}{\tau_o}}\, w\left(\frac{1}{2}\sqrt{\frac{\tau_n'}{t}} + \sqrt{\frac{t}{\tau}}\right)\right\} \ldots$$

$$-\exp\left(-\frac{\tau_n''}{4t}\right)\left\{2\sqrt{\frac{t}{\pi\tau_o}} - \frac{\sqrt{\tau}+\sqrt{\tau_n''}}{\sqrt{\tau_o}} w\left(\frac{1}{2}\sqrt{\frac{\tau_n''}{t}}\right) + \sqrt{\frac{\tau}{\tau_o}}\, w\left(\frac{1}{2}\sqrt{\frac{\tau_n''}{t}} + \sqrt{\frac{t}{\tau}}\right)\right\} \quad (A.11)$$

where $\tau'_n = [(2n+1)l - z_o]^2/\kappa$ and $\tau''_n = [(2n+1)l + z_o]^2/\kappa$ and $w(x) = \exp(x^2) erfc(x)$

If the characteristic time, τ, for diffusion of the latent heat is neglected (i.e. it is much smaller than time for conduction across the lithosphere, then the transform (A.9) has only simple poles in $s=0$ and $s = -(2n+1)^2\pi^2/4\tau_l$ and the inverse transform is obtained as the series:

$$\frac{\Delta z_m(t)}{\Delta z_0} = 1 - \frac{8l}{\pi^2 z_0} \sum_{n=0}^{\infty} (-1)^n \frac{\sin((2n+1)\pi z_0/2l)}{(2n+1)^2} \exp(-(2n+1)^2\pi^2 t/4\tau_l)$$

$$(A.12)$$

CHAPITRE III

PALEO HEAT-FLOW OF EASTERN CANADA'S PASSIVE MARGINS

Yvette Hamdani and Jean-Claude Mareschal

Abstract

Tectonic subsidence rates, derived from several wells in the Labrador Sea and offshore Nova-Scotia and Newfoundland, were used to estimate the paleo heat-flow along the margins of eastern Canada. If tectonic subsidence is caused by the cooling of the lithosphere only, the transient component of the heat-flow (i.e. the heat-flow in excess of background heat-flow) can be estimated directly from subsidence data.

For the Nova-Scotia and the Labrador Sea margins, the analysis shows that the transient component of the heat-flow decreased markedly from 28-56 $mW.m^{-2}$, immediately after rifting, to a present value between 7 and 14 $mW.m^{-2}$ depending on boundary conditions. The analysis shows a distinctive difference between the evolution of the Labrador Sea and that of the northeastern Newfoundland margin where it is estimated that the transient component of the heat-flow exceeded 100 $mW.m^{-2}$ after rifting and dropped rapidly to very low values. The study suggests that the subsidence of the Newfoundland margin was not caused by cooling but mostly by continuing extension after continental breakup and/or ductile deformation in the lower crust and upper mantle. The estimated transient heat-flow was used to calculate the average present surface heat-flow in eastern Canada's margins, which is the sum of heat-flow from the mantle, crustal heat production, and the transient cooling of the lithosphere. The calculated heat-flow compares well with heat-flow density data obtained from bottom hole temperature measurements.

1. Introduction

Passive continental margins are found on both sides of the Atlantic, in the Indian Ocean along the African coast and Australia, and surrounding Antarctica (e.g. Dingle, 1980; Veevers, 1980). The crustal structure of several parts of these continental margins has been studied by seismic and gravity surveys. These studies have shown that passive continental margins are characterized by thick sediments deposited over a crystalline crust thinned by extension. Along the oldest margins of eastern North America (ca 180 Ma), such as the east coast of the U.S. and Nova Scotia, 6 to 12 km of sediments were deposited; in younger margins (ca 90-100 Ma), such as the Labrador Sea and northeastern Newfoundland, 3 to 10 km of sediments accumulated (Keen *et al.*, 1975, 1987; Keen and de Voogd, 1988; van der Linden, 1975; Balkwill, 1987; Grant, 1987; Keen *et al.* 1990).

Passive continental margins form after continental breakup (Vogt and Ostenso, 1967). Their evolution follows two stages. During the rifting stage, asthenospheric upwelling and extension produce crustal and lithospheric thinning, and uplift accompanied by subaerial erosion. The subsidence of the margin sometimes begins at the end of the rifting stage. During the drifting stage, the thermal perturbation caused by asthenospheric upwelling decays. The lithosphere cools and thickens because of excessive flow of heat at the surface and/or reduced flow of heat at the base. Thermal subsidence is caused by the decay of the initial thermal perturbation and the thickening of the lithosphere. The subsidence is accompanied by sediment deposition.

"Tectonic subsidence" refers to that part of subsidence caused by the tectonic processes acting during formation and evolution of the margins exclusive of isostatic adjustments and sea-level changes. During rifting and the early stages of margin evolution, subsidence is affected by lithospheric extension and by the thermal transient (Royden and Keen, 1980). During the later stages of margin evolution, the decay of the thermal transient is the principal cause of tectonic subsidence (Sleep, 1971; Sleep and Snell, 1976; Keen, 1979). In addition, flexure of the lithosphere caused by the sedimentary load amplifies the subsidence. Tectonic subsidence can be determined from sediment thickness by "backstripping" (Steckler and Watts, 1978). Backstripping requires an assumption on the mechanism of isostatic adjustment (i.e. the flexural rigidity of the lithosphere), the prior knowledge of sea level changes, and information on sediment compaction history.

Following Turcotte and Oxburgh (1967), several authors have investigated the thermal subsidence of the cooling oceanic lithosphere (Parker and Oldenburg, 1973; Parsons and Sclater, 1977). Calculations predict that the subsidence of the cooling lithosphere is proportional to the square root of the age of the sea floor. The observations support these calculations, at least for a lithosphere younger than 80-100 Ma; the flattening of the bathymetry at greater age is well explained by the plate model (e.g. Sclater and Francheteau, 1970; Parsons and Sclater, 1977). A similar analysis was also applied to the subsidence of passive margins by Sleep (1971) and Sleep and Snell (1976); Steckler and Watts (1978), and Keen (1979) analyzed subsidence from the margins of eastern North America and showed that the tectonic subsidence of these margins is proportional to $t^{1/2}$, and is thus well accounted for by the cooling and thermal contraction mechanism.

McKenzie (1978) proposed that the subsidence of sedimentary basins follows an episode of mechanical stretching of the lithosphere. The subsidence is caused by thermal contraction, cooling and thickening of the lithosphere initially thinned by stretching. The stretching model was applied to many intracontinental basins, such as the North Sea Basin (Sclater et Christie, 1980) and the Pannonian Basin (Royden *et al.*, 1983a, b). Le Pichon and Sibuet (1981) and Beaumont *et al.* (1982) applied the same analysis to continental margins and suggested that the stretching factor varies with depth. Non uniform stretching of the lithosphere was proposed by Royden and Keen (1980) and Keen *et al.* (1987) to explain the evolution of eastern Canada's passive continental margins.

If the stretching factors are known, it is possible to calculate the thermal evolution and the heat-flow of the subsiding margin. Alternatively, it is possible to determine the heat-flow directly from the subsidence history, independently of any assumption on mechanism. Keen (1979) used a half-space cooling model to determine a global heat-flow history from a best fit to the subsidence data of eastern Canada's margins. Mareschal (1987, 1991) suggested that, since the thermal subsidence is caused by the cooling of the lithosphere, the past heat-flow in excess of the steady state could be calculated directly from the tectonic subsidence rate and he estimated the paleo heat-flow in several sedimentary basins. For the North Sea Basin, there were two episodes of rapid subsidence ca 120 Ma and ca 60 Ma when the heat-flow increased by about 45mW.m^{-2}. The high subsidence rate of the Pannonian Basin requires very high heat-flow in excess of steady state, varying from about 115 mW.m^{-2} at 8 Ma to 35 mW.m^{-2} today. The estimated heat-flow was highly variable during the evolution of the Michigan Basin; this variability may be caused by errors in the backstripping procedure that result from uncertainties in the past sea level variations. Finally, for the North American margin offshore New-York, the excess heat-flow has decreased from 80 mW.m^{-2} at 150 Ma to 15 mW.m^{-2} today.

The purpose of the present study is the determination from the tectonic subsidence rate of the paleo heat-flow in excess of steady state along eastern Canada's passive margins. The study demonstrates very distinctive evolutions for the different parts of eastern Canada's margins. Excess of the paleo heat-flow off Labrador and Nova Scotia varied from about 60 $mW.m^{-2}$ to less than 10 $mW.m^{-2}$ today. The very rapid initial subsidence in northeastern Newfoundland requires the transient part of the heat-flow to have been at least 120 $mW.m^{-2}$; this excessively high value suggests that the mechanism of subsidence was not exclusively thermal.

2. Method of analysis.

The tectonic subsidence of passive margins is directly related to their thermal history (Sleep, 1971; Steckler and Watts, 1978). During rifting, the lithosphere is heated and undergoes thermal expansion. Subsequently, it cools, contracts, and subsides during the drifting stage. Subsidence by thermal contraction is directly proportional to the total quantity of heat lost during cooling, which is the difference between the heat-flow at the base and at the surface of the lithosphere. The relationship between the surface heat-flow in excess of steady state and subsidence depends on the assumption on heat-flow at the lithosphere asthenosphere boundary (LAB), i.e. on boundary and initial conditions (Mareschal, 1991).

2.1 Cooling with constant heat-flow at the LAB.

If heat-flow is constant at the LAB, the excess surface heat flux is directly proportional to the rate of tectonic subsidence. The relationship between the excess heat-flow, Δq, and the subsidence rate, $\dot{h}$, is given by (Mareschal, 1987):

$$\Delta q(t) = \frac{K}{\alpha\kappa}\dot{h} = \frac{\rho c}{\alpha}\dot{h}$$

where K is the thermal conductivity, κ is the thermal diffusivity, ρ is the density of the lithosphere, c is the thermal capacity, and α is the thermal expansion coefficient.

Typical values for the physical parameters would be $\rho = 3{,}200\ kg.m^{-3}$, $c = 700\ J.kg^{-1}.K^{-1}$, and $\alpha = 3.10^{-5}\ K^{-1}$ (Clark, 1966). For these values, the ratio of excess heat-flow (in $mW.m^{-2}$) to the subsidence rate (in $m.Myr^{-1}$) is about 2.3 $mW.Myr.m^{-3}$ (Mareschal, 1987). Alternatively, $1 mW.m^{-2}$ of excess heat flow is accompanied by 0.42 m of subsidence.

2.2 Cooling with constant temperature at the LAB.

Defining the LAB as an isotherm implies that temperature, and not heat-flow, is constant at the base of the lithosphere. Then, the lithosphere cools partly because of increased heat-flow at the surface, and partly because of reduced heat-flow at the base. Consequently, the heat-flow in excess is less than predicted with the assumption of constant flux at the LAB:

$$\Delta q(t) \leq \frac{\rho c}{\alpha} \dot{h}$$

In general, the excess heat flux depends on initial as well as boundary conditions. However, because the memory of the initial thermal conditions is damped out, the lithosphere will cool as much by increased surface heat-flow as by reduced heat from below the LAB. Therefore, the surface heat-flow will tend to (Mareschal, 1987):

$$\Delta q(t) = \frac{\rho c}{2\alpha} \dot{h}$$

This will be valid after a time on the order of 15 Myrs. For the same values of the thermal parameters as used above, the ratio of excess heat-flow to subsidence rate will thus tend to 1.15 $mW.Myr.m^{-3}$.

3. Development and subsidence of eastern Canada's margins.

The passive margins of eastern North America were formed by the breakup of the continents around the Atlantic: the development of the margin of Nova Scotia followed the separation of the African and North American plates in Early Jurassic (at ca 175 Ma), the formation of the Labrador margin resulted from the separation of Greenland and North America, and the evolution of the margin of Newfoundland followed the separation of Europe from North America in Late Cretaceous time (at ca 92 Ma and 100 Ma respectively) (Klitgord and Schouten, 1986; Srivastava, 1978; Roest and Srivastava, 1989). Tectonic reconstruction of the evolution of the Labrador Sea suggests that the southern part of the margin separated first. There is some question about the formation of the northeastern margin of Newfoundland. According to Enachescu (1987), rifting started at ca 200 Ma and continued episodically with several reactivation periods until 80 Ma; most of the tectonic subsidence took place during drifting.

The tectonic subsidence is determined by "backstripping" the sediment accumulation data (e.g. Steckler and Watts, 1978). Figure 1 shows the location of several wells from the Nova-Scotia, Labrador, and Newfoundland continental margins that were used to determine tectonic subsidence in this study. The name and the location of the wells are listed listed in Table 1. The subsidence data were backstripped by Keen (1979) who did not include sediment compaction and sea-level changes for the Nova-Scotia margin because the variations in sea-level are poorly known for the Jurassic. Figures 2a, 2b, and 2c show the subsidence determined in these wells as a function of $t^{1/2}$ where t is the time after the initiation of drifting.

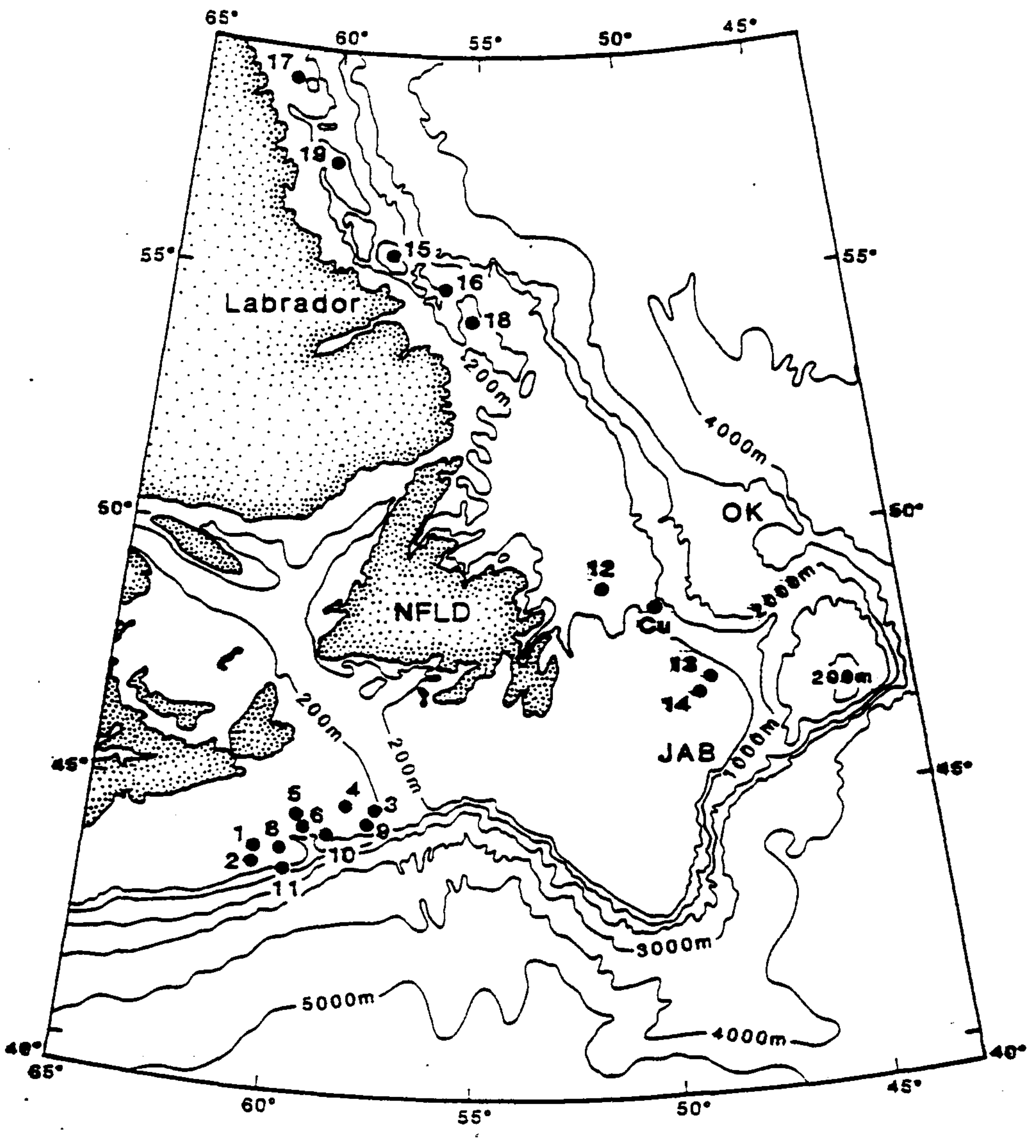

Figure 1: Map of eastern Canada's margin showing the location of the wells used to determine subsidence rates (adapted from Keen, 1979). A list of the wells and their location is given in Table 1. OK: Orphan Knoll, JAB: Jeanne d'Arc Basin.

3.1 Nova-Scotia. Northern Labrador Sea.

Tectonic subsidence of Nova Scotia and the Labrador Sea is proportional to $t^{1/2}$, where t is time after the initiation of subsidence. The best estimate of the slope of the tectonic subsidence vs squareroot of age line is 300 (± 80) $m.Myr^{-1/2}$ (Keen, 1979).

Royden and Keen (1980) have compared uniform and non-uniform stretching models with the subsidence of the margins of Nova-Scotia and of the Labrador Sea. The difference in rheology causes differences in extension rate between the brittle upper crust and the ductile lower crust and mantle and is the cause of non-uniform stretching. This analysis of the subsidence determined in 5 wells in Nova-Scotia and 4 wells in the Labrador Sea concluded that: (1) stretching was more or less uniform for the margins of Nova-Scotia with a stretching factor on the order of 1.75 to 2.0; (2) in the Labrador Sea, the stretching factor was on the order of 1.3 to 1.6 in the crust, and varied between 2.5 and 10 in the lower crust and mantle.

In the northern part of the Labrador Sea, the increased sedimentation on the margin during the Pliocene and the Pleistocene may be not related to thermal events; Royden and Keen (1980) have suggested that it is the result of recent glaciations while Cloetingh et al. (1990) have assumed that it is the effect of changing intraplate stress regime. They relate the acceleration of the sedimentation in the Pliocene (at ca 5 Ma) to a late Neogene plate reorganization and associated stress changes. This part of the subsidence record has not been considered because it seems to be not thermal in origin.

Table 1. List and location of the wells used for the paleo-heat-flow estimation.

Location	Number and name of well	Latitude	Longitude
Nova Scotia	1. Cohasset D-42	43°51'	60°37'
	2. Cree E-35	43°44'	60°35'
	3. Dauntless D-35	44°44'	57°20'
	4. Esperanto K-78	44°47'	58°11'
	5. Micmac J-77	44°36'	59°26'
	6. Missisauga H-54	44°33'	59°22'
	8. Sable C-67	43°56'	59°55'
	9. Sachem D-76	44°35	57°40'
	10. Sauk A-57	44°16'	58°37'
	11. Triumph P-50	43°39'	59°51'
South Labrador Sea			
	15. Bjarni H-81	55°30'	57°42'
	16. Gudrid H-55	54°54'	55°52'
	18. Leif M-48	54°17'	57°07'
North Labrador Sea			
	17. Karlsefni H-13	58°32'	61°46'
	19. Snorri J-90	57°19'	59°57'
Newfoundland			
	12. Bonavista C-99	49°08'	51°14'
	13. Dominion O-23	47°22'	48°18'
	14. Flying Foam I-13	47°02'	48°46'

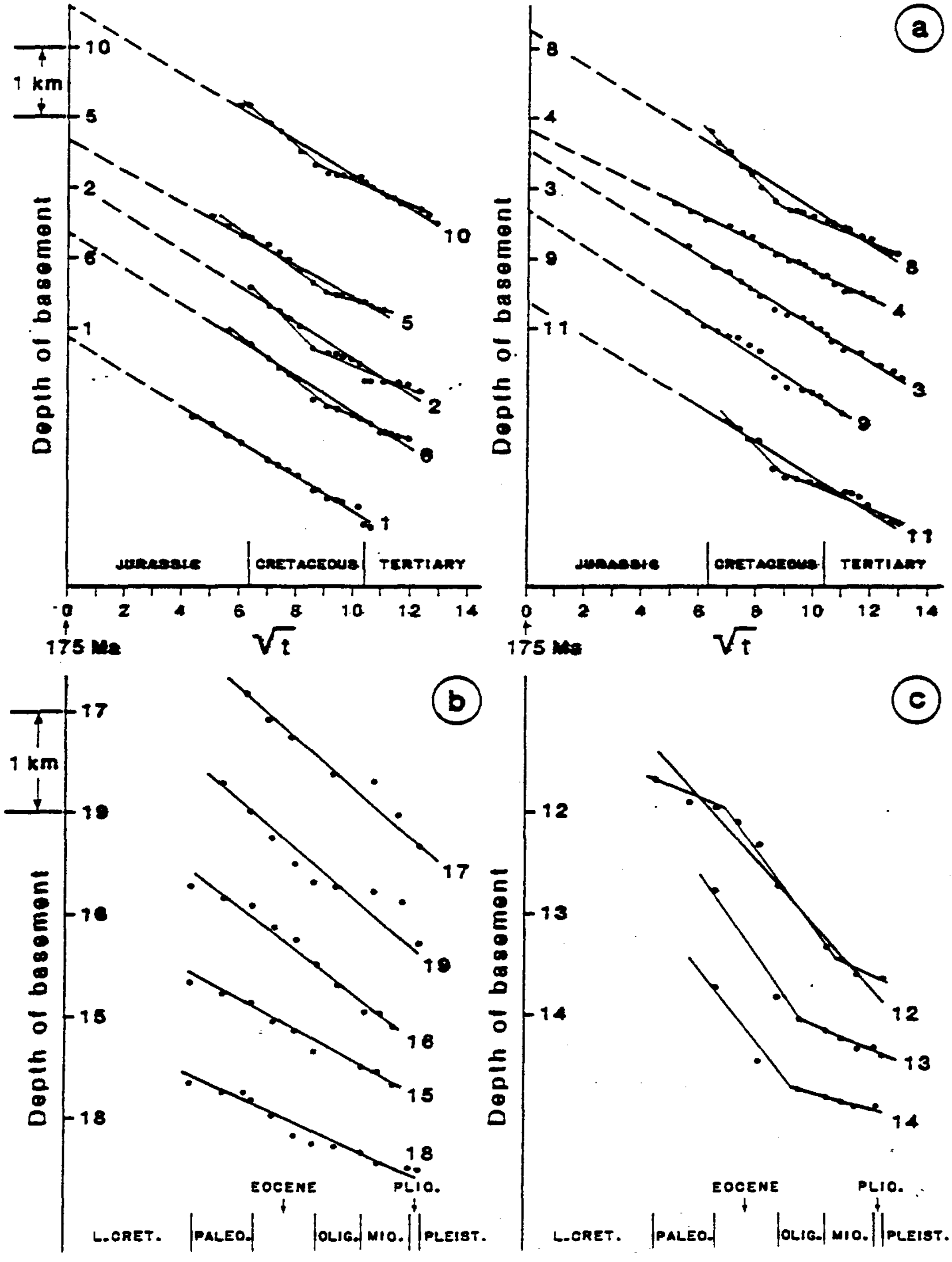

Figure 2: Subsidence as a function of $\sqrt{t}$ determined for the wells on eastern Canada's margin. For the wells of Nova-Scotia (a), time is measured from 175 Ma, for the Labrador Sea (b) and for Newfoundland (c), time is measured from 75 Ma. The best fitting straight line to the subsidence vs $\sqrt{t}$ data were computed by Keen (1979). For Newfoundland, two different lines are necessary to fit the break in the slope of the subsidence vs $\sqrt{t}$ data at 40 Ma.

3.2 Southern Labrador Sea. Newfoundland.

The evolution of the southern part of the Labrador Sea and the northeastern part of the Newfoundland margin started at about the same time as drifting (i.e. at ca 92 Ma in Labrador Sea and at ca 100 Ma in Newfoundland). The data clearly show that the tectonic subsidence of northeastern Newfoundland (Fig. 2c), which was extremely rapid during the first 20 Myrs, is distinctively different from that of the Labrador Sea margin (Fig. 2b). An initial period of extremely rapid subsidence (with an average slope of the tectonic subsidence vs squareroot of age curve on the order of 450 m $Myr^{-1/2}$) was followed by a period of attenuated subsidence rate (Keen, 1979). Keen *et al.* (1987) identified three episodes of stretching in northeastern Newfoundland with stretching factors much higher in the lower than in the upper part of the lithosphere. They suggested that the extension of the lithosphere continued in the margin of Newfoundland and the Orphan Basin after continental breakup. On the other hand, Grant (1987) suggested that the continental margins east of Newfoundland formed by vertical displacement possibly related to a dense body under the basin.

4. Thermal history of the margins.

The relationships between the excess surface heat flux and the tectonic subsidence rate were used to constrain the paleo heat-flow in the margins of Nova-Scotia, Newfoundland, and Labrador. The tectonic subsidence rate is not calculated for individual wells, but by stacking together the subsidence records of several nearby wells that have experienced similar tectonic evolution. Figures 3, 4, and 5 compare the paleo heat-flow determined directly from subsidence data with the heat-flow calculated for the stretching models.

4.1 Nova-Scotia.

Figures 3a, 3b, 3c, and 3d show the average tectonic subsidence rate and the excess heat-flow estimated directly from the subsidence rate for the Nova-Scotian margin. It is compared with the heat-flow determined for the stretching model of McKenzie (1978) with the extension factors estimated by Royden and Keen (1980). Details on the calculation of the heat-flow for the non-uniform stretching model are provided in the Appendix A. The tectonic subsidence rate was obtained by averaging between wells that are close together and have similar subsidence history, i.e. the following wells: (a) 1, and 2, (b) 8, and 11, (c) 5, 6 and 10, and (d) 3, 4 and 9. These figures indicate that the excess heat-flow was very high (28-56 $mW.m^{-2}$) at the beginning of the drifting stage. Subsequently, as the lithosphere returned to thermal equilibrium, the excess heat-flow decreased to the present low value of less than 5 $mW.m^{-2}$. The small fluctuations of the tectonic subsidence rate and heat-flow are probably caused by errors because sea level variations were not included in the backstripping procedure. The heat-flow calculated for the Nova-Scotian margin is close to that estimated for the eastern U.S. (Mareschal, 1987).

4.2 Labrador Sea.

The tectonic subsidence rate for the margin of the Labrador Sea was calculated by averaging data from wells 15, 16, and 18 located in the southern part of the margin and from wells 17 and 19 in the northern part. On Figure 4a and 4b, the excess heat-flow determined directly from the subsidence rate is compared with the heat-flow calculated for the non-uniform stretching model proposed by Royden and Keen (1980).

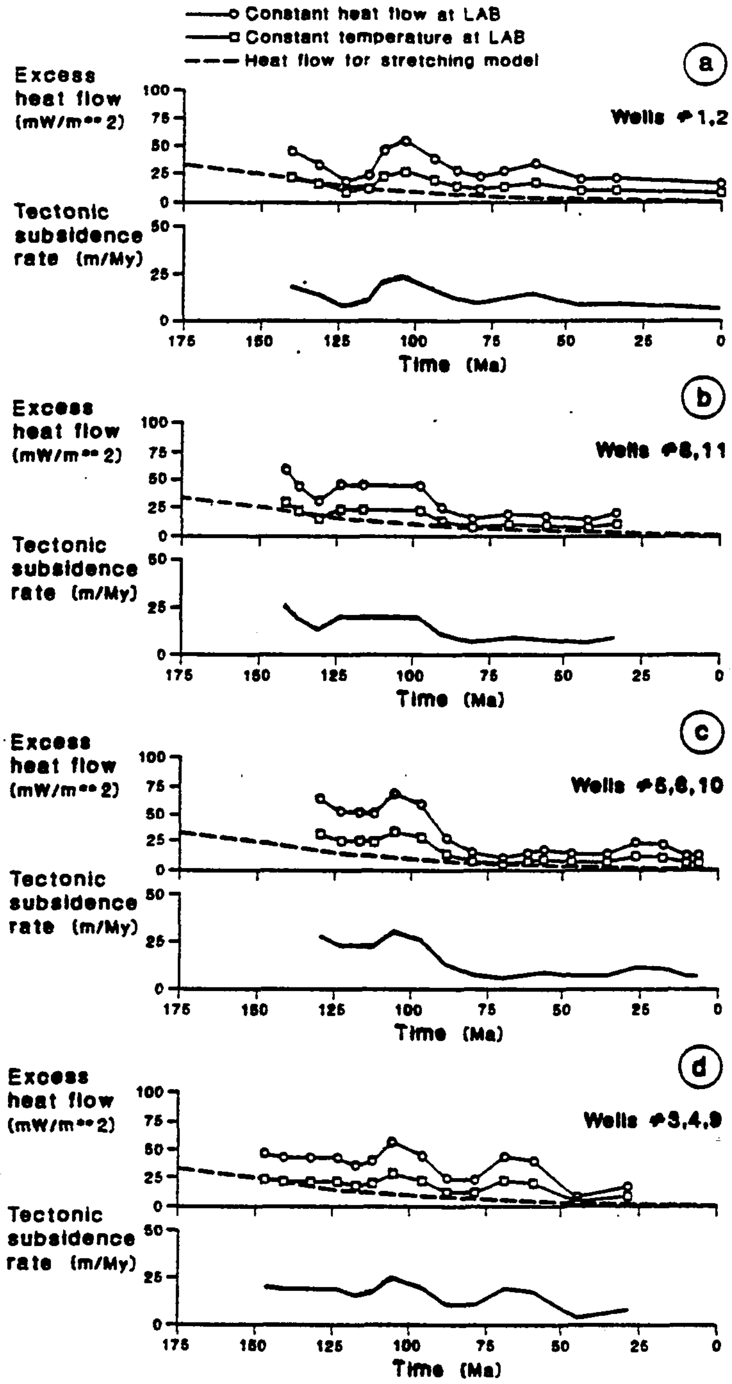

Figure 3: Subsidence rate and past heat-flow (in excess of background) determined directly from subsidence data and calculated for a uniform stretching model for the margin of Nova-Scotia. The subsidence rate is determined by averaging between wells that are close together. Past heat-flow dropped from 28 to 56 mW.m^{-2} (depending on boundary condition) at 150 Ma to close to zero today.

4.3 Newfoundland.

Figure 5 shows the average tectonic subsidence rate and the excess heat-flow for the margin of Newfoundland. The excess heat-flow determined from the tectonic subsidence rate is compared with the theoretical heat-flow excess calculated for non-uniform stretching model with stretching factors suggested by from Keen and Barrett (1981). The excess heat-flow of the Newfoundland margin was extremely high (>100 $mW.m^{-2}$) during the early stages of drifting, but it decreased rapidly to its present value which is close to zero.

5. Discussion.

The results of the present analysis can be compared with present heat-flow data, constraints from thermal maturation analysis, and other studies of the evolution of the margins.

5.1 Comparison with present heat-flow.

The calculated excess heat-flow and assumed crustal heat production and mantle heat-flow were also used to estimate the present heat-flow along the different margins, which can be compared with heat-flow measurements.

The present heat flux was calculated as the sum of three components:

$$q = q_H + q_r + \Delta q$$

where q_r is the reduced heat-flow which includes the heat-flow from the mantle and the contribution of deep crustal heat sources, q_H is the contribution of heat sources in the shallow crust, and Δq is the heat flux in excess (i.e. transient effect following the heating of the lithosphere during break-up). These different components were estimated to obtain the present heat-flow of the margins studied.

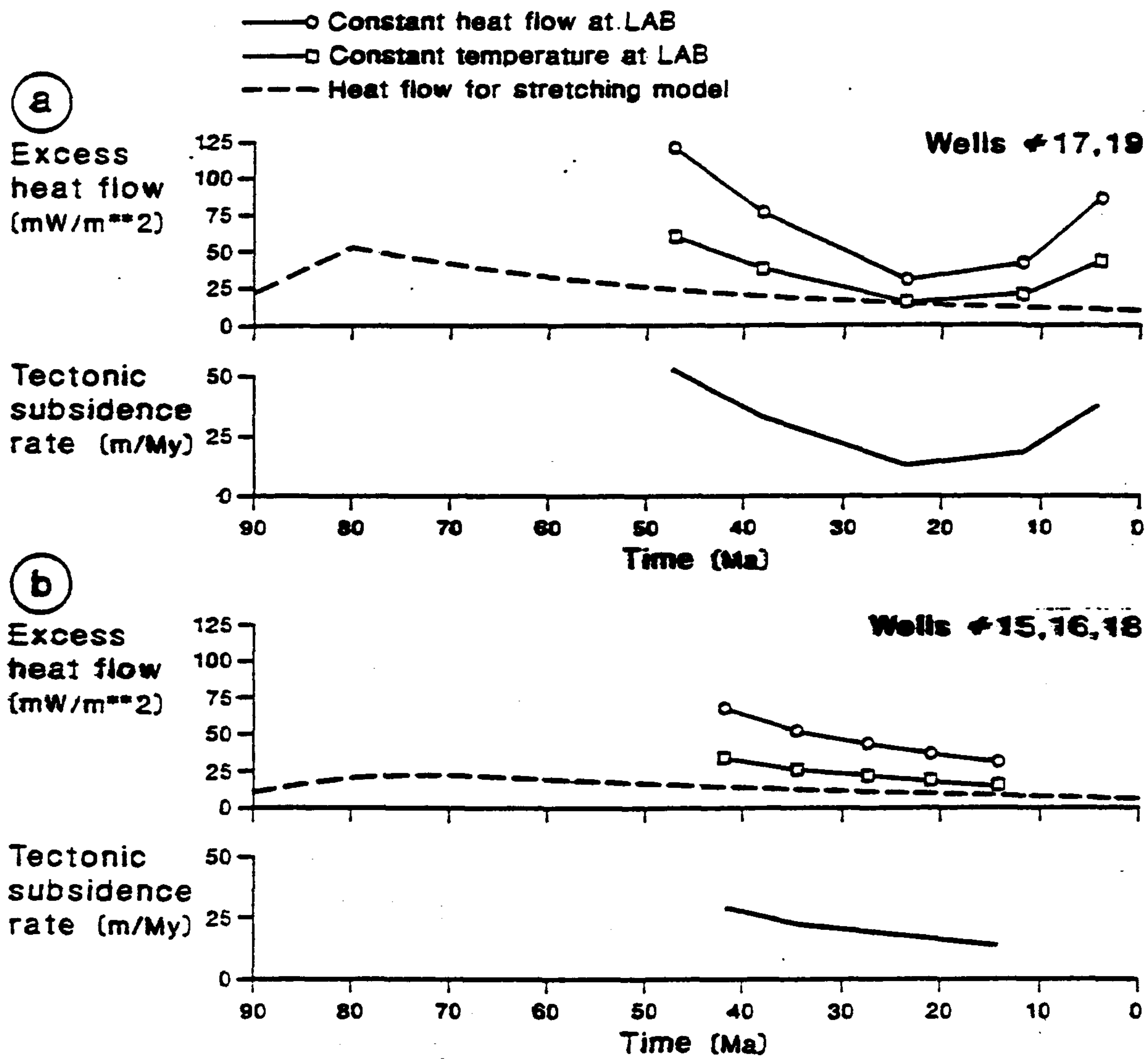

Figure 4: Subsidence rate and past heat-flow for different wells for the Labrador Sea. It is compared with the heat-flow predicted by the non-uniform stretching for an opening time of 90 Ma. With a stretching factor larger in the lower crust and mantle than in the upper crust, the heat-flow increases during the first 15 Myr.

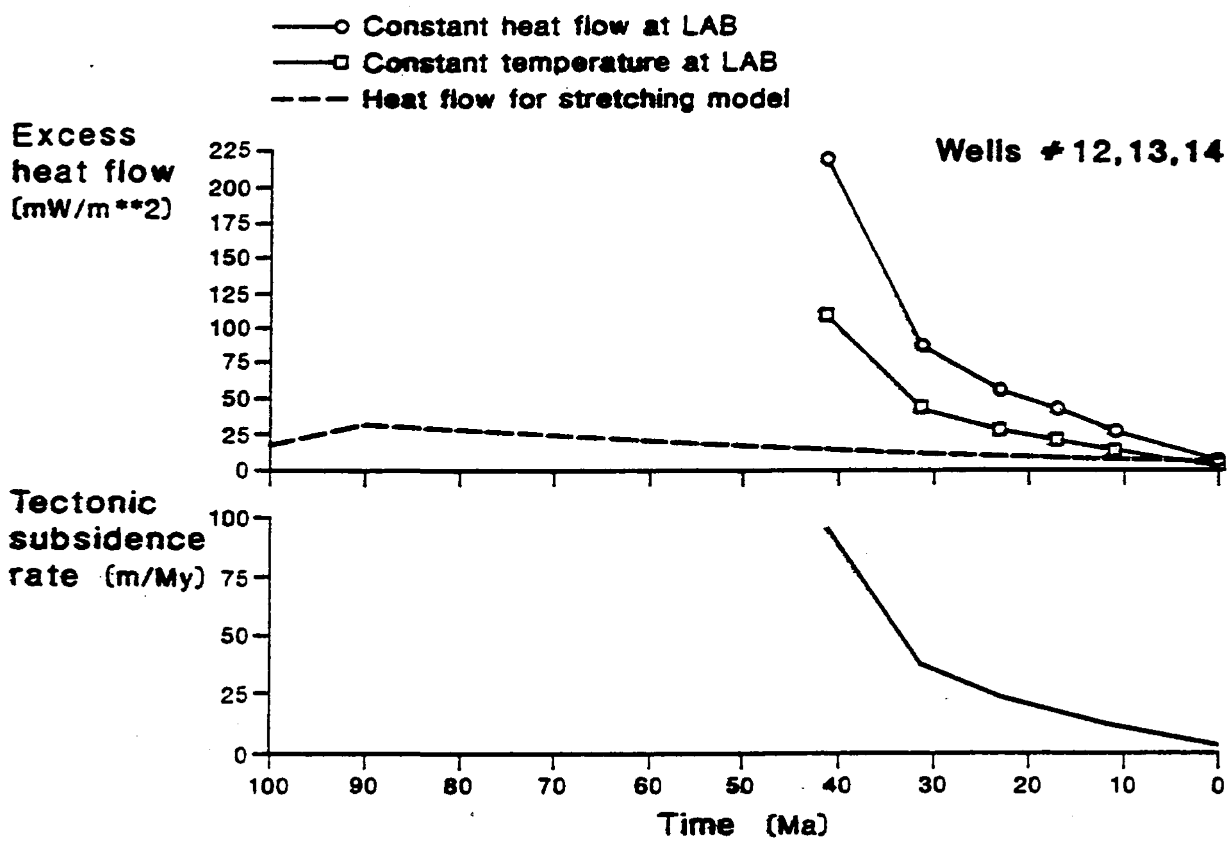

Figure 5: Subsidence rate and past heat-flow in Newfoundland. It is compared with the heat-flow predicted by a non-uniform stretching model and opening time of 100 Ma. The very high subsidence rate ca 40 Ma requires an extremely high heat-flow in excess of background or it implies that tectonic subsidence was not thermal.

A value of 27 $mW.m^{-2}$ was assumed for the reduced heat-flow q_r throughout these calculations. Hyndman *et al.* (1979) suggest that the reduced heat-flow in Nova-Scotia is 31.9 $mW.m^{-2}$. This value is close to those proposed for the eastern U.S. (Roy *et al.*, 1968), stable continental shields (Rao and Jessop, 1975), and the Grenville Province (Pinet *et al.*, 1991). Several authors have estimated that the reduced heat-flow in stable continental regions is on the order of 27 $mW.m^{-2}$ (e.g. Morgan and Sass, 1984; Pujol *et al.*, 1985).

Table 2 summarizes the estimated contribution of crustal heat production to the present day heat-flow. Hyndman *et al.* (1979) estimated the heat production of the Paleozoic sediments and granites of the upper crust of Nova-Scotia to be on the order of 2.1 $\mu W.m^{-3}$. Keen and Lewis (1982) have estimated the heat generation in the sediments to be on the order of 1 $\mu W.m^{-3}$. The value of 0.8 $\mu W.m^{-3}$ retained by Royden and Keen (1980) for heat production of the Precambrian crust in Labrador is close to 1 $\mu W.m^{-3}$ measured on samples from four Labrador Shelf wells (Issler and Beaumont, 1987). The thickness of the upper radioactive crustal layer before stretching (h) was divided by the stretching factor, except for the wells of Newfoundland's where it was deduced from seismic refraction data (Keen and Barrett, 1981). The average contribution of the upper crystalline layer to the surface heat-flow was assumed to be 8 $mW.m^{-2}$ for the Nova Scotian margin, 6-8 $mW.m^{-2}$ for the Labrador Sea and 12 $mW.m^{-2}$ along the Newfoundland margin. The heat production of the sediments was assumed to be 0.9 $\mu W.m^{-3}$; the sediments' contribution to surface heat-flow was 7-9 $mW.m^{-2}$ for the Nova Scotian margin and 2-3 $mW.m^{-2}$ for the Labrador Sea and Newfoundland margins. Table 3 compares various estimates of the present heat-flow along these margins. For the Labrador Sea and Newfoundland, the present study yields values on the order of 45-50 $mW.m^{-2}$ much lower than the values between 60 and 90 $mW.m^{-2}$ calculated by Reiter and Jessop (1985).

Table 2. Contribution of the crustal heat production to the present surface heat-flow

Location	Sediments			β	δ	Upper crystalline layer			Total contribution to heat flow
	Mean thickness (km)	Mean heat production ($\mu W.m^{-3}$)	Contribution to surface heat flow ($mW.m^{-2}$)			Thickness (km)	Mean heat production ($\mu W/m^{-3}$)	Contribution to heat flow ($mW.m^{-2}$)	($mW.m^{-2}$)
Nova Scotia									
(wells 3,4,9)	10.0	0.9	9.0	2.0		3.7[1]	2.1	7.8	16.8
(wells 5,6,10)	9.0	0.9	8.1	2.0		3.7[1]	2.1	7.8	15.9
(wells 8,11)	9.5	0.9	8.6	2.0		3.7[1]	2.1	7.8	16.4
(wells 1,2)	7.5	0.9	6.8	2.0		3.7[1]	2.1	7.8	14.6
South Labrador sea									
(wells 15,16,18)	2.2	0.9	2.0	2.5	1.3	7.7[1]	1.0	7.7	9.7
North Labrador sea									
(wells,17,19)	3.2	0.9	2.9	7.5	1.65	6.1[1]	1.0	6.1	9.0
Newfoundland									
(wells 12,13,14)	3.1	0.9	2.8	3.0	1.5	12.0[2]	1.0	2.0	14.8

(1) calculated as thickness of the upper radioactive crustal layer before stretching (h) / stretching factor.
h=7.5 km for Nova Scotia (Hyndman *et al.* 1979)
h=10 km for Labrador sea and Newfoundland (Pinet *et al.* 1991)
(2) from seismic refraction Keen and Barrett (1981)

These values are based on estimates of the thermal conductivity of the sediments that might be too high. MacKenzie *et al.* (1985) calculated lower heat-flow values based on thermal conductivity measurements on samples from offshore eastern Canada. These latter values are in agreement with those of Issler and Beaumont (1986). Alternatively, the reported high heat-flow could be caused by perturbations due to water circulation in the sediments. Correia *et al.* (1990) show that fluid flow affects the heat-flow in the Jeanne d'Arc Basin where they estimate that the average heat-flow is on the order of 66 $mW.m^{-2}$ and they obtain 60 $mW.m^{-2}$ and 64 $mW.m^{-2}$ for the wells 13 and 14 investigated in this study.

The heat-flow estimate for the Nova-Scotian margin, on the order of 45-50 $mW.m^{-2}$, is in good agreement with heat-flow data and previous estimates. The average of 14 heat-flow measurements along the margin of Nova-Scotia is 48 $mW.m^{-2}$ (Lewis and Hyndman, 1976). The models of Issler and Beaumont (1986, 1987) and of MacKenzie *et al.* (1985) yield 40 to 50 $mW.m^{-2}$ for the margins of Nova-Scotia and the Labrador Sea. According to these results, the average heat flux would be 42.5 $mW.m^{-2}$ for the region of wells 15, 16, and 18 and 44.6 $mW.m^{-2}$ for the region of wells 5 and 6 (see Table 3).

The heat-flow estimated directly from subsidence rate compares better with the heat-flow data for a constant temperature boundary condition at the LAB. This boundary condition implies that some of the cooling takes place by reduced heat-flow at the base of the lithosphere and, therefore, it yields lower estimates of the past heat-flow.

Table 3. Measured and estimated present heat flow

Location	q_m	q_{bl}	Present average heat flow from subsidence data		β	δ	Present average heat flow for stretching model	Heat flow from BHT
			q const at LAB	T const at LAB				
Nova Scotia	27							
(wells 3,4,9)		16.8	60.6	52.2	2.0		45.1[1]	65(av.)[4]
(wells 5,6,10)		15.9	50.7	46.6	2.0		44.2[1],44.6[2]*	68.7(av.)[4]
(wells 8,11)		16.4	56.8	50.1	2.0		44.7[1],41.1[2]**	78.5(av.)[4]
(wells 1,2)		14.7	52.0	46.8	2.0		42.9[1]	58 (from well#2)[4]
South Labrador sea	27							
(wells 15,16,18)		9.7	57.3	47.0	2.5	1.3	42.8[1],42.5[3]	90 (from well#15)[4]
North Labrador Sea								
(wells 17,19)		9.0	-	-	7.5	1.65	45.1[1],45.1[3]	73 (from well#17)[4]
Newfoundland	27							
(wells 12,13,14)		14.8	49.6	45.7	3.0	1.5	49.1[1]	60(av.)[5]

Heat flow in $mW.m^{-2}$

(1) calculated with the stretching factors proposed by Royden and Keen (1980), Keen and Barrett (1981).

(2) estimated by MacKenzie *et al*. (1985)

(3) estimated by Issler and Beaumont (1986, 1987)

(4) measured by Reiter and Jessop (1985)

(5) average value estimated from wells 12,13 (Correia *et al*. 1990) and well Cumberland B-55 (Reiter and Jessop, 1985)

av.=average

* from wells 5, and 6 only

** from well 8 only

5.2 Comparison with stretching models.

Table 4 summarizes the results for the three margins studied and compares them with the heat-flow calculated for the uniform or non-uniform stretching models. Because the stretching models assume a constant temperature at the base of the lithosphere, they predict heat-flow values that are in good agreement with the excess heat-flow calculated directly from tectonic subsidence rate with constant temperature at the LAB. This does not imply that this boundary condition is more appropriate. Extrapolation of the curves yielded an estimate of present excess heat-flow except for wells 17,19 in the northern part of the Labrador Sea where the acceleration of the tectonic subsidence from ca 20 Ma to present is not thermal in origin.

The difference between stretching models and direct estimates is evident for the Labrador margins; this might be caused by uncertainties in backstripping and/or the estimation of the stretching factors. The difference is particularly striking for the Newfoundland margin where the direct estimates indicate very high excess heat-flow (>100 $mW.m^{-2}$ ca 40 Ma). This corroborates previous analyses concluding that the margins of the Labrador Sea and those of northeastern Newfoundland have experienced distinctive thermal and tectonic evolutions. If the subsidence accelerated because of continuing extension during the drifting stage, all the estimates based on the assumption that the subsidence is thermal will be in error. Alternatively, if very rapid cooling took place because of high stretching factor and extreme lithospheric thinning, the thermal estimates would be valid.

Table 4. Summary of past and present heat-flow perturbation (mW.m^{-2}) from subsidence data

Location	Initial Δq from subsidence data		Present Δq from subsidence data		β	δ	Initial Δq for stretching model	Present Δq for stretching model
	q constant at LAB	T constant at LAB	q const at LAB	T const at LAB				
Nova Scotia								
(wells 3,4,9)	45.4	22.7	16.8	8.4	2.0		20.6	1.3
(wells 5,6,10)	58.4	29.2	7.8	3.9	2.0		23.2	1.3
(wells 8,11)	64.0	32.0	13.4	6.7	2.0		18.0	1.3
(wells 1,2)	46.0	23.0	10.4	5.2	2.0		22.9	1.3
South Labrador Sea								
(wells 15,16,18)	66.8	33.4	20.6	10.3	2.5	1.3	14.6	6.1
North Labrador Sea								
(wells,17,19)	120.4	60.2	?	?	7.5	1.65	26.2	9.1
Newfoundland								
(wells 12,13,14)	226.0	113.0	7.8	3.9	3.0	1.5	17.5	7.3

Excess heat flow for stretching models is calculated with the stretching factors proposed by Royden et Keen (1980), Keen and Barrett (1981).

5.3 Comparison with thermal maturation models.

Issler (1984) calculated the Time-Temperature Index (TTI) for Nova-Scotia margin by assuming that the temperature gradient in the sediment did not change with time. A value of 26.6 $°C.km^{-1}$ was assumed for these calculations corresponding to the present average temperature gradient which varies between 19-35$°C.km^{-1}$ in wells of the Nova-Scotian margin (Reiter and Jessop, 1985). A model following MacKenzie *et al.* (1985) assumes a constant temperature gradient of 20$°C.km^{-1}$ for the margins of Nova-Scotia. The thermal maturity studies are compatible with similar gradients for the Grand-Banks region south east of Newfoundland, and slightly higher gradients on the order of 30$°C.km^{-1}$ for the Labrador Shelf (Issler,1984). The recent measurement of matrix conductivity on samples from deep exploratory wells in Nova-Scotia yield a mean value of 1.88 $W.m^{-1}.K^{-1}$ (Keen and Beaumont,1990) and thus an average heat-flow of 40-50 $mW.m^{-2}$, compatible with the estimates of present heat-flow obtained by this study. The results of the present study show that the heat-flow from the basement changed much during the subsidence but they do not invalidate the calculations of TTI. Indeed, the thermal regime in the sediments is affected by several opposing factors: the increased heat-flow during subsidence is absorbed to warm up the deposited sediments. Calculations of the temperature gradient in the sedimentary sections have shown that it is approximately constant because these two factors compensate for each other (Balling, personnal communication). In addition, thermal maturation data, which are mostly affected by the maximum temperature conditions, are a poor indicator of the entire thermal history of the basin. Usually, the temperature will be highest when the sediments reach the greatest depth of burial, at the end of the basin's history. The thermal maturation data are less sensitive to the early subsidence history when the depth of burial is small and consequently the temperature remains low despite the high heat-flow.

6. Conclusions.

This study has provided an estimate of the thermal regime and history of eastern Canada's passive margins. These estimates compare well with the present heat-flow data from the margin. The heat-flow estimated directly from subsidence data (even with a temperature boundary condition at the LAB) is in general slightly higher than the heat-flow calculated for various stretching models. The direct determination of past heat-flow offers the advantage that it is not affected by errors on estimates of stretching factors and opening time. However, it is affected by noise in the data and errors in the backstripping procedure. The effect of these errors is to add some high frequency noise to the estimated heat-flow, but the consequence for the long term trends will be small.

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Appendix A. Excess heat-flow for uniform and non uniform stretching.

The thermal perturbation, ΔT, is solution of the one dimensional heat equation with boundary and initial conditions (Carslaw and Jaeger, 1959):

$$\frac{\partial \Delta T}{\partial t}(z,t) = \kappa \nabla^2 \Delta T(z,t) \qquad (A.1)$$

where κ is the thermal diffusivity.

For a uniform stretching factor β, the appropriate boundary conditions are:

$$\Delta T(z=0,t) = \Delta T(z=L,t) = 0$$

and the initial condition is:

$$\Delta T(z,t=0) = \frac{(\beta-1)T_L z}{L} \quad z < L/\beta$$

$$\Delta T(z,t=0) = T_L(1-z/L) \quad L/\beta < z < L$$

where T_L is the temperature at the LAB.

The surface heat-flow is (McKenzie, 1978):

$$\Delta q(t) = 2q_e \sum_{n=1}^{\infty} \frac{\beta}{n\pi} \sin\left(\frac{n\pi}{\beta}\right) \exp(-n^2\pi^2\kappa t/L^2) \qquad (A.2)$$

where q_e is the equilibrium heat-flow ($q_e = K\, T_L/L$ where K is thermal conductivity).

The heat-flow for non uniform stretching with a factor δ for the upper crust and β for the lower crust and upper mantle can be derived from Roydèn and Keen (1980) (see fig. A-1 for the initial conditions). It gives for the heat-flow:

$$\Delta q(t) = 2q_e \sum_{n=1}^{\infty} b'_n \exp(-n^2\pi^2\kappa t/L^2)$$

where:

$$b'_n = \frac{(\delta - \beta)}{n\pi}\sin\left(\frac{n\pi y}{L\delta}\right) + \frac{\beta}{n\pi}\sin\left(\frac{n\pi y}{L\delta} + \frac{n\pi(L-y)}{L\beta}\right)$$

where y is the initial crustal thickness.

In the calculations, the following values were assumed for the parameters: thermal diffusivity, $\kappa = 1\mathrm{x}10^{-6}$ m^2. s^{-1}, equilibrium heat-flow, $q_e = KT_L/L = 33$ mW.m^{-2}, temperature at the LAB, $T_L = 1333$°C, crustal thickness before stretching, y = 35 km, lithospheric thickness, L = 125 km.

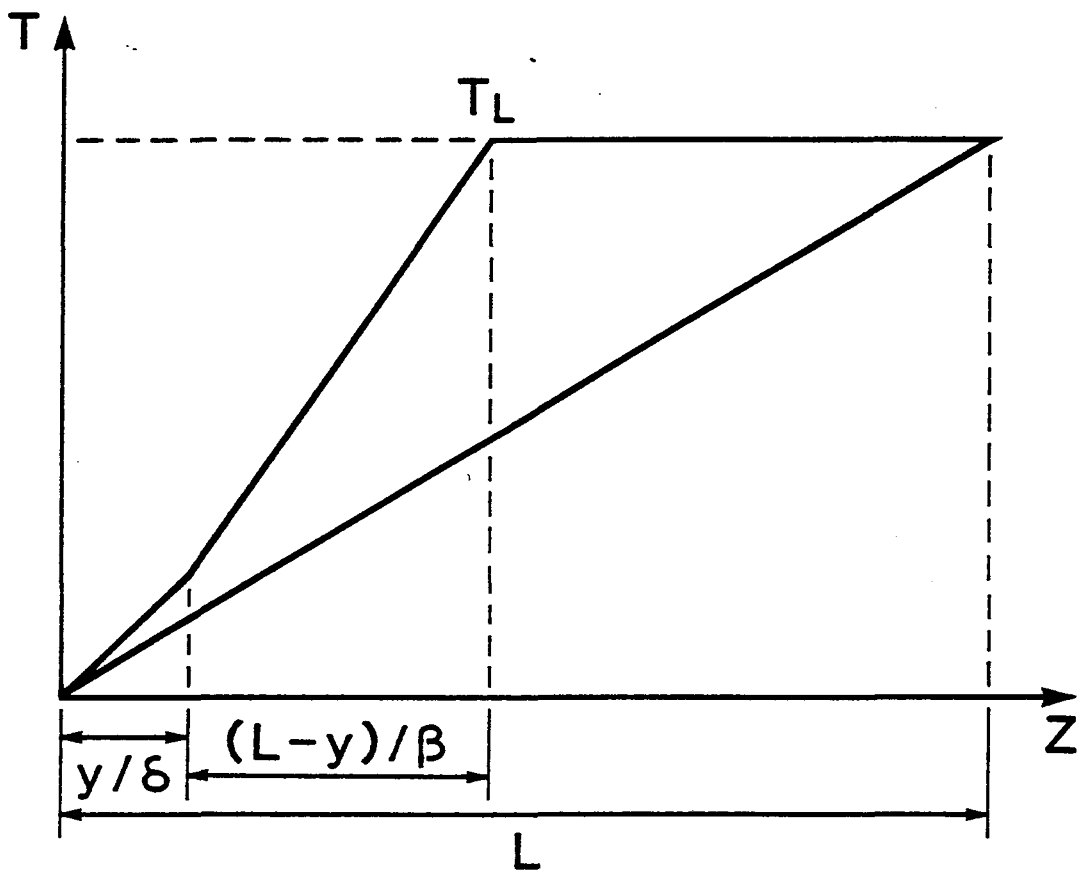

Figure A-1: Initial conditions for the non-uniform stretching model.

CONCLUSIONS

Les deux premiers chapitres de cette thèse sont consacrés au développement et à l'application de modèles de subsidence tectonique pour les bassins sédimentaires du Michigan et du Williston. Les modèles supposent que la subsidence est le résultat de la contraction thermique et de changements de phase suivant une perturbation thermique (une diminution instantanée de la température ou du flux de chaleur) à la base de la lithosphère. La subsidence causée par les changements de phase est proportionnelle au déplacement de l'interface qui sépare deux phases. Les solutions analytiques pour la subsidence thermique et les approximations analytiques pour le déplacement de l'interface séparant deux phases sont calculées à l'aide de la technique de la transformée de Laplace. Mathématiquement, le déplacement de l'interface est déterminé par la solution d'un problème de Stefan. Ce problème est difficile à résoudre car il est non-linéaire. Le déplacements a été calculé à l'aide d'une approximation linéaire proposée par Mareschal et Gangi (1977a). Cette approximation est valide pour les petits déplacement de l'interface (c.a.d. le déplacement de l'interface reste petit par rapport à la profondeur de l'interface). L'effet du poids des sédiments sur le changement de phase a été calculé pour un taux de sédimentation constant. Dans le cas de l'approximation linéaire, différentes composantes du déplacement de l'interface peuvent être superposées; de cette manière, l'effet de la sédimentation sur le déplacement de l'interface a été ajouté à celui du refroidissement.

Les conclusions générales qui découlent des modèles mathématiques étudiés dans les deux premiers chapitres sont les suivantes:

1) La subsidence thermique dépend des conditions aux limites choisies; le temps de retour à

l'équilibre thermique est 4 fois plus long pour un changement brusque du flux de chaleur que pour un changement brusque de température à la base de la lithosphère. Si la lithosphère a une épaisseur de 100 km, l'équilibre thermique s'établit en 50 Ma après une diminution soudaine de la température et en 200 Ma après une diminution soudaine du flux de chaleur à la base de la lithosphère. Le temps de retour à l'équilibre dépend de l'épaisseur de la lithosphère au début de la formation du bassin (voir aussi Mareschal, 1981).

2) L'amplitude de la subsidence thermique est proportionnelle à l'amplitude du changement thermique à la base de la lithosphère.

3) L'amplitude de la subsidence causée par les changements de phase est directement proportionnelle à l'amplitude du changement thermique à la base de la lithosphère et inversement proportionnelle à la différence entre le gradient géothermique et le gradient de la courbe de Clausius-Clapeyron.

4) La subsidence due aux changements de phase est retardée par rapport à la subsidence thermique et ce délai dépend des conditions aux limites choisies et de la distance entre la base de la lithosphère et la position initiale de l'interface qui sépare deux phases. Il varie entre 1-100 Ma.

Pour le bassin du Michigan, le modèle proposé combine la contraction thermique et la transformation métamorphique d'une granulite à grenat en éclogite dans le manteau supérieur à la suite d'un changement brusque de la température à la base de la lithosphère. L'évolution du bassin a été calculée en tenant compte des variations du niveau de la mer et de l'amplification isostatique (selon le mécanisme d'Airy). La subsidence dans le bassin du Michigan commence à 460 Ma. Les calculs reproduisent les traits généraux de la subsidence de ce bassin. En particulier, l'accélération de la subsidence autour de 440 Ma est une conséquence du retard (de 20 Ma) du phénomème de changement de phase par rapport à la contraction thermique et la discordance dans la stratigraphie (entre 420 et 400 Ma) semble être causée par une baisse du niveau de la mer. Le modèle requiert la présence d'éclogite sous la discontinuité de Mohorovicic (Moho). La densité de cette couche d'éclogite peut expliquer l'anomalie de Bouguer positive, observée sous le bassin du Michigan.

La longue subsidence du bassin du Williston (370 Ma) nécessite une condition aux limites différente à la base de la lithosphère ou une lithosphère initialement très épaisse. Trois mécanismes pour l'évolution de ce bassin ont été comparés: 1) la subsidence tectonique est le résultat du changement de phase (du gabbro en granulite à grenat) et de la contraction thermique suite à une diminution brusque de la température à la base d'une lithosphère très épaisse (250 km); 2) la subsidence est le résultat seulement de la contraction thermique causée par une diminution substantielle (7.2 $mW.m^{-2}$) du flux de chaleur à la base de la lithosphère; 3) la subsidence tectonique est produite par la combinaison du changement de phase de gabbro en granulite à grenat dans la croûte inférieure et de la contraction thermique suite à une diminution du flux de chaleur (de 3.2 $mW.m^{-2}$) à la base de la lithosphère. Dans les deux derniers cas, l'épaisseur de la lithosphère est supposée de 150 km. Les calculs de la subsidence tiennent compte de l'effet de l'amplification isostatique et des variations du niveau de la mer.

Les résultats obtenus démontrent qu'une diminution de la température à la base d'une lithosphère épaisse n'explique pas la subsidence observée. Ce modèle prédit correctement l'accumulation totale de sédiments, mais pas leur âge. La réponse thermique de la lithosphère au changement de température est rapide et environ 2.4 km de sédiments s'accumulent en 100 Ma, contrairement à ce qui est observé dans le bassin du Williston. De plus l'épaisseur initiale implique une lithosphère froide et incompatible avec le refroidissement ultérieur.

La modélisation démontre qu'un changement de 7.2 $mW.m^{-2}$ du flux de chaleur à la base de la lithosphère explique mieux la subsidence du bassin du Williston. Toutefois, l'amplitude du changement du flux est très élevée et correspond à une variation de température de l'ordre de 367 K à la base de la lithosphère.

Les résultats qui satisfont le mieux les données ont été obtenus en supposant que la subsidence du bassin du Williston est le résultat de la contraction thermique et de la transformation métamorphique du gabbro en granulite à grenat dans la croûte inférieure à la suite d'une diminution du flux de chaleur (3.2 $mW.m^{-2}$) à la base de la lithosphère. La subsidence du bassin du Williston commence à 512.5 Ma. Le modèle explique la discordance dans l'accumulation de sédiments (entre 500 et 470 Ma) par une faible subsidence et une baisse de niveau de la mer. Une accélération de la subsidence se produit à 470 Ma, parce que le changement de phase commence à influencer la subsidence tectonique environ 40 Ma après de l'initiation du bassin. Selon le modèle, la subsidence du bassin du Williston proprement dit est terminée à 150 Ma et un nouvel épisode de subsidence commence à 120 Ma avec l'ensemble du bassin d'avant-pays formé à l'est des Montagnes Rocheuses.

Ce modèle est en accord avec les données sismiques montrant une couche dans la croûte inférieure du bassin ayant une vitesse correspondant à la granulite à grenat et les données gravimétriques donnant une anomalie de Bouguer positive. De plus, ce modèle requiert un changement de température à la base de la lithosphère compatible avec ceux suggérés par les modèles de plume dans le manteau (White et McKenzie, 1989; Sleep, 1990; Griffiths and Campbell, 1991) ainsi qu'avec le flux de chaleur mantélique de l'ordre 12 $mW.m^{-2}$ estimé pour la province du Supérieur (Pinet *et al.* 1991).

Par conséquent, les conclusions concernant des modèles proposés pour expliquer la formation des bassins du Michigan et du Williston se résument de la manière suivante:
1) Les changements de phase ont affecté considérablement l'évolution de ces bassins. Après un certain temps (50 Ma pour le Michigan et 150 Ma pour le Williston), la subsidence due aux changements de phase représente la majeure partie de la subsidence tectonique.
2) Le retard du changement de phase fourni une explication pour l'accélération de subsidence 20-40 Ma après le début de la formation de ces bassins.
3) La différence de la durée de la subsidence dans les bassins du Michigan et du Williston peut s'expliquer par différents mécanismes d'interaction thermique entre l'asthénosphère et la lithosphère qui sont reflétés par la condition aux limites; une diminution brusque de la température à la base de la lithosphère pour le bassin du Michigan et une diminution du flux de chaleur pour le bassin du Williston.

Toutefois, à cause du grand degré d'incertitude dans les variations de niveau de la mer (dans le temps et dans l'amplitude), il est impossible d'exclure définitivement les variations de niveau de la mer comme cause de l'accélération de la subsidence au stade inital de la formation de ces bassins. Seule une meilleure connaissance des variations de niveau de la mer permettra de résoudre cette question.

Dans le troisième chapitre, le paléo-régime thermique des marges de la Nouvelle-Écosse, de la mer de Labrador et du nord-est de Terre-Neuve a été déterminé à partir du taux de la subsidence tectonique, en supposant que la subsidence est due uniquement à la contraction thermique. Les résultats obtenus sont les suivants:

1) Pour les marges de Nouvelle-Écosse et la partie sud des marges de la mer de Labrador, l'excès moyen du paléo-flux de chaleur au début du stade de dérive était de l'ordre 28-56 $mW.m^{-2}$ et sa valeur actuelle est environ de 7-14 $mW.m^{-2}$.

2) Pour la partie nord des marges de la mer de Labrador, la valeur initiale de l'excès du flux de chaleur est de 60-120 $mW.m^{-2}$, mais sa valeur actuelle n'a pu être estimée parce que l'augmentation récente du taux de subsidence n'est pas d'origine thermique.

3) L'excès du flux de chaleur des marges de Terre-Neuve a été extrêmement élevé (100-200 $mW.m^{-2}$) durant les 20 premiers millions d'années de la phase de dérive et sa valeur actuelle tend vers zéro.

Le flux de chaleur en excès estimé a été comparé avec le flux calculé pour le modèle de striction (uniforme ou non-uniforme selon la région étudiée). Les résultats correspondent à la condition de température constante à la base de la lithosphère. Toutefois, il existe une différence évidente pour les marges de Labrador et de Terre-Neuve, qui pour le Labrador peut être due à une inexactitude dans le "backstripping" ou/et dans l'estimation des facteurs de striction.

L'analyse des résultats met en évidence la particularité des marges du nord-est de Terre-Neuve et suggère une évolution tectonique différente de celle des marges de la mer de Labrador. Les causes de la subsidence anomalique des marges du nord-est de Terre-Neuve restent à préciser. Elles pourraient être liées à l'extension de la lithosphère durant la phase de dérive et à la déformation ductile de la croûte inférieure et/ou du manteau supérieur. Si ces phénomèmes ont effectivement eu lieu, une partie de la subsidence tectonique n'est pas d'origine thermique et le paléo-régime thermique ne peut être déterminé à partir des données de subsidence.

Le flux de chaleur des marges a été estimé à partir des valeurs de l'excès actuel du flux de chaleur et en tenant compte du flux de chaleur réduit, de la production de chaleur des sédiments et de la croûte supérieure. Pour les marges de Nouvelle-Écosse, de la mer de Labrador et celles de nord-est de la Terre-Neuve, le flux estimé est de l'ordre de 45-50 $mW.m^{-2}$. Ces valeurs sont comparables avec le flux de chaleur mesuré (Lewis and Hyndman, 1976) et les modèles (Issler et Beaumont, 1986, 1987; MacKenzie *et al.* 1985). Par ailleurs, la comparaison suggère que les résultats sont compatibles avec une température constante à la base de la lithosphère. Cette condition implique, que la lithosphère se refroidit par conduction de la chaleur vers la surface et par une réduction du flux à la base de la lithosphère.

Le gradient géothermique suggéré par cette étude pour les marges de Nouvelle-Écosse est de l'ordre 24-27°$C.km^{-1}$. Il reste comparable avec le gradient mesuré (Reiter et Jessop, 1985; Issler, 1984) et démontre de cette façon la validité de la technique utilisée. Cette technique permet une estimation directe du flux de chaleur à partir des données de subsidence qui n'est pas affectée par l'estimation du facteur de striction et l'approximation de l'épaisseur de la lithosphère.

Cette recherche complète le répertoire des mécanismes de subsidence des bassins sédimentaires et approfondit la connaissance de l'évolution thermique des marges continentales passives de l'est du Canada. Les modèles d'évolution des bassins intracratoniques du Michigan et du Williston sont plausibles et compatibles avec leur histoire sédimentaire, avec les données gravimétriques et sismiques. Ces modèles montrent le rôle qu'ont pu jouer des changements de phase durant la formation de ces bassins. L'étude de l'évolution thermique a mis en évidence l'évolution tectonique particulière des marges du nord-est de Terre-Neuve.

Toutefois certains problèmes restent à résoudre. Les modèles conçus pour les bassins du Michigan et du Willistion sont uni-dimensionels. Ils ne tiennent pas compte de l'effet de la flexure, mais introduisent l'amplification isostatique dans les calculs de la subsidence. Ils restent valides seulement pour la modélisation de la subsidence au centre des bassins. Les études futures devront être portées sur le développement des modèles bi-dimensionels de subsidence à l'aide de méthodes numériques (c.a.d. d'éléments finis). D'une part, les modèles bi-dimensionels permettront de tenir compte de l'effet de la flexure sur la subsidence tectonique de ces bassins circulaires (en supposant une charge sédimentaire axisymétrique à la surface). D'autre part, une résolution numérique permet d'introduire dans les calculs l'augmentation de l'épaisseur de la couche élastique et de la rigidité en flexion avec le refroidissement de la lithosphère; ceci permettra de tenir compte de la variation de l'épaisseur de la lithosphère durant l'évolution des bassins. D'autres bassins intracratoniques (les bassins d'Illinois et de la Baie d'Hudson) devront également être étudiés. Différents mécanismes de subsidence (les changements de phase, la striction, la convection dans le manteau supérieur) seront testés comme hypothèse de la formation de ces bassins.

Les données de subsidence des marges du nord-est de Terre-Neuve illustrent la complexité de la formation de cette région et démontrent que la contraction thermique suite à une striction n'est pas la seule cause de la subsidence tectonique. Il apparaît difficile de donner une explication de la subsidence anomalique de cette région. Néanmoins, certaines caractéristiques géophysiques suggèrent que les changements de phase dans la croûte inférieure en sont peut être la cause. Les données sismiques suggèrent une vitesse de croûte inférieure de l'ordre 7.4 $km.s^{-1}$ (Keen et Barrett, 1981) qui est celle de la granulite à grenat. L'anomalie d'air libre est atypique des marges continentales passives (Keen *et al.* 1990) par son amplitude (supérieure à 100 mGal) et par l'absence d'anomalies négatives sur ses flancs. Grant (1987) suggère qu'elle résulte de la présence d'un corps dense dans la croûte inférieure, correspondant peut être à la présence d'une couche de granulite à grenat.

Pour ces marges, une étude plus approfondie est nécessaire. L'étude comprendra:
1) La collecte, l'analyse et le "backstripping" des données de la subsidence provenant d'autres puits de forage disponibles.
2) Le développement de modèles numériques bi-dimensionels de subsidence tectonique combinant l'effet de la contraction thermique et des changements de phase et en supposant comme condition initiale une striction instantanée de la lithosphère. Les modèles bi-dimensionels vont permettre de tenir compte de l'effet de la conduction latérale de chaleur et de la rhéologie de la lithosphère sur la subsidence.
3) L'application de ce modèle à des marges du nord-est de Terre-Neuve.
Le développement de modèles de la subsidence tectonique incluant l'effet de changement de phase permettra de tester le rôle des changements de phase dans l'évolution des marges de Terre-Neuve.

Il serait également utile d'appliquer les résultats à l'estimation des températures dans les sédiments pour déterminer la maturation de la matière organique. Une méthodologie d'estimation de paléo-température dans les sédiments devrait donc être développée dans les prochains travaux de recherche. Elle combinera l'estimation de paléo flux de chaleur à partir des données de la subsidence avec les méthodes de continuation vers le bas. L'effet de réchauffement des sédiments sera pris en considération.

Les valeurs de l'ITT (l'indice de temps-température de la maturité) pour différentes formations sédimentaires seront calculées à partir des paléo-températures estimées. L'ITT qui caractérise le degré de maturité sera déterminé dans le but de définir les horizons sédimentaires suffisamment matures pour contenir des quantités importantes d'hydrocarbures. L'application pour diverses marges passives potentiellement riches en hydrocarbures sera présentée.

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APPENDICE

NOTE EXPLICATIVE CONCERNANT LES DONNÉES DE LA SUBSIDENCE ET LES CALCULS DE L'AMPLIFICATION ISOSTATIQUE

Il faut noter que les données de la subsidence du bassin du Willison montrées sur la figure 1 du chapitre I et celles montrées sur figures 2, 5, 6a,b du chapitre II proviennent de deux puits de forage différents; ceci explique la différence de la forme et de la durée des courbes de la subsidence. Les données de la subsidence (selon Fowler and Nisbet, 1985) utilisées dans le chapitre I proviennent du puits de forage situé en Saskatchewan, mais dont les auteurs ne citent pas ni les coordonnées géographiques ni la distance au centre du bassin. Ensuite, (Haid, 1991) a publié dans sa thèse d'autres données de subsidence du bassin du Williston. Par conséquent, les données provenant du puits de forage le plus proche du centre du bassin (Zabolotony N 1-3-4a) ont été choisies pour comparaison avec les modèles de subsidence.

Le coefficient de l'amplification isostatique utilisé pour modéliser la subsidence du bassin du Michigan (chapitre I) a été calculé selon la relation suivante:

$$f_l = \frac{\rho_m - \rho_w}{\rho_m - \rho_s} = 3.3 \qquad (1)$$

où

ρ_m est la densité du manteau=3.3 Mg.m^{-3}

ρ_w est la densité de l'eau de mer =1.Mg.m^{-3}

ρ_s est la densité des sédiments= 2.6 Mg.m^{-3}

La relation (1) suppose qu'une colonne de l'eau de mer est remplacée par une colonne de sédiments. Toutefois, les modèles de subsidence des bassins du Michigan et du Williston calculent le dépôt de sédiments à partir du niveau de la mer; ceci implique l'absence de l'eau de mer au moment de l'initiation des bassin. Dans ce cas, le coefficent de l'amplification isostatique devrait être calculé selon la relation suivante:

$$f_l = \frac{\rho_m}{\rho_m - \rho_s} = 4.7 \qquad (2)$$

La relation (2) a été appliquée pour modéliser la subsidence du bassin du Williston.

Concernant le bassin du Michigan, la correction du coefficient de l'amplification isostatique peut être compensée par une réduction de l'amplitude du changement de la température au moment de l'initiation du bassin (c.a.d. la diminution de la température sera de 170 K au lieu de 240 K).