1

National Library
of Canada

Bibliothèque nationale
du Canada

CANADIAN THESES
ON MICROFICHE

THÈSES CANADIENNES
SUR MICROFICHE

0-315-04887-5

52781

NAME OF AUTHOR/NOM DE L'AUTEUR Chen-Roon Hwu

TITLE OF THESIS/TITRE DE LA THÈSE Stress, Seismicity and Initiation of Subduction
at Passive Continental Margins

UNIVERSITY/UNIVERSITÉ Carleton

DEGREE FOR WHICH THESIS WAS PRESENTED/
GRADE POUR LEQUEL CETTE THÈSE FUT PRÉSENTÉE M. Sc.

YEAR THIS DEGREE CONFERRED/ANNÉE D'OBTENTION DE CE DEGRÉ 1981

NAME OF SUPERVISOR/NOM DU DIRECTEUR DE THÈSE G. Ranalli

Permission is hereby granted to the NATIONAL LIBRARY OF
CANADA to microfilm this thesis and to lend or sell copies
of the film.

The author reserves other publication rights, and neither the
thesis nor extensive extracts from it may be printed or other-
wise reproduced without the author's written permission.

*L'autorisation est, par la présente, accordée à la BIBLIOTHÈ-
QUE NATIONALE DU CANADA de microfilmer cette thèse et
de prêter ou de vendre des exemplaires du film.*

*L'auteur se réserve les autres droits de publication; ni la
thèse ni de longs extraits de celle-ci ne doivent être imprimés
ou autrement reproduits sans l'autorisation écrite de l'auteur.*

DATED/DATE March 18, 1981 SIGNED/SIGNÉ Chen-Roon Hwu

PERMANENT ADDRESS/RÉSIDENCE FIXE 542 W. 112th Street, #6D
New York, N.Y., 10025
USA

National Library of Canada
Collections Development Branch

Canadian Theses on
Microfiche Service

Bibliothèque nationale du Canada
Direction du développement des collections

Service des thèses canadiennes
sur microfiche

NOTICE

AVIS

The quality of this microfiche is heavily dependent upon the quality of the original thesis submitted for microfilming. Every effort has been made to ensure the highest quality of reproduction possible.

If pages are missing, contact the university which granted the degree.

Some pages may have indistinct print especially if the original pages were typed with a poor typewriter ribbon or if the university sent us a poor photocopy.

Previously copyrighted materials (journal articles, published tests, etc.) are not filmed.

Reproduction in full or in part of this film is governed by the Canadian Copyright Act, R.S.C. 1970, c. C-30. Please read the authorization forms which accompany this thesis.

**THIS DISSERTATION
HAS BEEN MICROFILMED
EXACTLY AS RECEIVED**

La qualité de cette microfiche dépend grandement de la qualité de la thèse soumise au microfilmage. Nous avons tout fait pour assurer une qualité supérieure de reproduction.

S'il manque des pages, veuillez communiquer avec l'université qui a conféré le grade.

La qualité d'impression de certaines pages peut laisser à désirer, surtout si les pages originales ont été dactylographiées à l'aide d'un ruban usé ou si l'université nous a fait parvenir une photocopie de mauvaise qualité.

Les documents qui sont déjà l'objet d'un droit d'auteur (articles de revue, examens publiés, etc.) ne sont pas microfilmés.

La reproduction, même partielle, de ce microfilm est soumise à la Loi canadienne sur le droit d'auteur, SRC 1970, c. C-30. Veuillez prendre connaissance des formules d'autorisation qui accompagnent cette thèse.

**LA THÈSE A ÉTÉ
MICROFILMÉE TELLE QUE
NOUS L'AVONS REÇUE**

STRESS, SEISMICITY, AND INITIATION
OF SUBDUCTION AT PASSIVE CONTINENTAL MARGINS

by

CHEN-ROON HWU, B.Sc.

A thesis submitted to
the Faculty of Graduate Studies and Research
in partial fulfilment of the requirements for
the degree of Master of Science

Department of Geology

Carleton University

Ottawa, Canada

March, 1981

The undersigned hereby recommend to
the Faculty of Graduate Studies and Research
acceptance of this thesis,
submitted by Chen-Roon Hwu, B.Sc.,
in partial fulfilment of the requirements for
the degree of Master of Science.

Chairman, Department of Geology

Supervisor

Carleton University

ABSTRACT

The state of stress induced by differential gravitational loading at a passive continental margin is modelled using a two-dimensional elastic finite-element model. A spreading stress of about 20 MPa under the continental shelf relative to the oceanic crust is calculated. This stress is likely to be superimposed on stresses originated by sediment loads, ice loading and unloading, thermal effects, and plate tectonic forces.

If the accumulated stresses reach a critical value, earthquakes will occur. Earthquakes can occur under relatively low stress if the margin contains pre-existing faults which reduce the strength and concentrate the stresses. The seismicity of the Arctic and eastern Canadian continental margins is analyzed on this basis. Surface loading is the most likely triggering mechanism.

Lithospheric aging and surface loading appear to be critical factors in the initiation of subduction. Subduction may be initiated as a result of the stressing of flexured and fractured lithosphere at a passive continental margin, due to the combined effects of sedimentary loading, phase changes (basalt-eclogite transition) and tectonic forces (ridge push). However, there is no single law relating onset of subduction to age of lithosphere.

ACKNOWLEDGEMENT

I would like to express my sincere appreciation for the assistance and comments of Dr. G. Ranalli, the supervisor of the thesis. I also wish to thank Dr. G.Y. Chao for his continuous encouragement and help. The many excellent suggestions and opinions of numerous members of the computing centre and Professors W.H. Bowes in the Department of Civil Engineering and J.D. Dixon in the Department of Mathematics are gratefully acknowledged. The M. Sc. thesis of R. Stephenson and Ph. D. thesis of D.S. Dean provided guidance as to the design of the computer program. Thanks are also extended to W. Yzerdraat, C. Rees, C. Roots, and C. Kerans who, together with Dr. G. Ranalli, extensively edited earlier drafts of this thesis. Finally, I am very glad to have the opportunity to complete this M. Sc. program at Carleton University, and have enjoyed the instruction and help of the Professors of the Geology Department.

CONTENTS

1. Introduction'	1
2. Stresses	4
2.1 Rheology of the lithosphere	4
2.2 Differential loading stresses	7
2.3 Thermal stresses and thermal effects	15
2.3.1 Temperature distribution within the plate	15
2.3.2 Thermal stresses	18
2.3.3 Subsidence	20
2.4 Flexural stresses induced by sediment or ice	23
2.4.1 Formulae governing vertical displacement	23
2.4.2 Flexural stresses	26
2.5 Driving forces	29
2.6 Synopsis of stresses and forces	30
3. Seismicity at passive continental margins of Canada	32
3.1 The seismicity of the Canadian Arctic continental margin	32
3.1.1 The Yukon-Mackenzie Valley	39
3.1.2 The Arctic continental shelf and slope	39
3.1.3 The eastern Canadian Arctic Islands	41
3.2 The seismicity at the eastern Canadian margin	42
3.3 The causative forces	49
3.3.1 Plate tectonic forces	49

3.3.2 Postglacial uplift	52
3.3.3 Sedimentary loads	55
3.3.4 Pre-existing faults and other features	56
4. Conversion of passive into active continental margin ...	59
4.1 Gravitational instability	59
4.1.1 Penetration of asthenospheric material	61
4.1.2 Basalt-eclogite transition	63
4.2 The initiation of subduction	67
4.3 The maintenance of subduction	69
5. Conclusions	73
5.1 Stress-generating mechanisms at passive continental margins	73
5.1.1 Stresses induced by differential loading	73
5.1.2 Stresses induced by thermal effects	74
5.1.3 Stresses induced by flexure due to sediments or ice	74
5.1.4 Plate tectonic forces	75
5.2 State of stress and earthquakes	75
5.2.1 Arctic continental margin	76
5.2.2 Eastern Canada continental margin	76
5.3 Initiation of subduction	77
6. References	79

FIGURES AND TABLES

Fig. 2.1	Rheological zonation of oceanic lithosphere with a point load	5
Fig. 2.2	Lower crustal oceanward hot creep at a passive continental margin	6
Fig. 2.3	Cross-section of the model of a rifted margin	9
Fig. 2.4	Stress distribution along passive continental margin .....See the Pocket in the back	
Fig. 2.5	Maximum shear stress profiles and crustal strength ..	11
Fig. 2.6	Variations of horizontal deviatoric tensile stress ..	12
Fig. 2.7	Subsidence curve of the Nova Scotia margin	22
Fig. 2.8	Stress distribution with depth within the lithospheric plate due to vertical deflection	27
Fig. 3.1	Geological provinces and geographic locations of Arctic Archipelago	34
Fig. 3.2	Free-air gravity anomalies and seismicity along the Canadian Arctic margin	36
Fig. 3.3	Explanations for the positive free-air gravity anomaly of passive continental margins	38
Fig. 3.4	Earthquake epicentres in northern Canada and adjacent regions	40
Fig. 3.5	Free-air gravity map of offshore eastern Canada	44
Fig. 3.6	Cross-sections and gravity anomalies of the eastern Canadian continental margins	47
Fig. 3.7	Seismicity and focal mechanisms of eastern Canadian margin	48
Fig. 3.8	Maximum horizontal stress direction inferred from the earthquake in the Yukon Territory	50
Fig. 3.9	Deviatoric tension and compression vector for the earthquake in Beaufort Sea	51

Fig. 3.10	Average rates of postglacial uplift for northern North America	53
Fig. 3.11	Geometry of the lithospheric flexural stresses induced by deglaciation	54
Fig. 4.1	Schematic successive relationship to show the evolution of a passive plate	60
Fig. 4.2	Stability fields of rocks for eclogite and basalt	62
Fig. 4.3	Petrological model for the evolution of passive continental margins	64
Fig. 4.4	Sinking slab self-maintaining requirements	71
Table 2.1	The parameters of the model	8

1. INTRODUCTION

There are two different types of continental margins, passive continental margins and active continental margins. These margins separate areas of continental lithosphere from areas of oceanic lithosphere. Passive continental margins are essentially aseismic and are commonly characterized by thick wedges of sediments. Good examples of this are the continental margins on both sides of the Atlantic ocean. The active continental margins, on the other hand, are seismic, often associated with trenches and volcanic arcs, and are loci of metamorphism and orogenic deformation. The plate boundaries bordering the margins of the Pacific ocean are good examples of this. The difference between these types of margins is that an active continental margin is situated at the boundary between two plates (interplate), while a passive continental margin is within one plate (intraplate).

According to the generally accepted plate-tectonic hypothesis, a passive continental margin may evolve into an active continental margin. Passive continental margins were once plate boundaries, dividing continents that were rifted and then drifted apart resulting in oceanic expansion. The passive continental margin thus marks the original line of splitting, or is bounded by transform faults with the adjacent lithosphere. The initial stages of the evolution of a passive continental margin involve uplift and stretching, a process accompanied by faulting. Subsequently the margin slowly cools

and subsides, allowing large amounts of sediment to be deposited. The erosion of crustal layers results in crustal thinning and supplies sediments for deposition at a passive continental margin. Thus the passive continental margin is stressed by thermal effects, sedimentary loading, plate-tectonic forces and differential gravitational loads. As the plate ages, these stresses accumulate at the passive continental margin. If the combined stresses reach the critical point where they can induce earthquakes, seismicity occurs. Sufficiently large accumulated stresses may even initiate subduction, changing the margin into an active one. Thus the state of stress at a passive continental margin plays a very important role in the plate tectonics hypothesis.

In Chapter Two of this thesis, a model of the effects of differential gravitational loading^o is analyzed by the finite element method. For the purpose of comparison, a discussion about the state of stress induced by thermal stress, sedimentary loading (including glacial loading and unloading), and plate tectonic forces is also included. In Chapter Three, the knowledge of the state of stress at a passive continental margin is applied to the seismicity around Canada's passive continental margins. When the critical stress for seismicity has been reached, an earthquake is induced by sudden release of the accumulated stresses which are caused by different mechanisms such as sedimentary loading, and post-glacial rebound. Pre-existing faults play a very important role in seismicity and in the initiation of subduction because they not only reduce the crustal

strength but they also concentrate the stress. Finally, in Chapter Four, the state of stress is related to the initiation of subduction. The relatively denser eclogite which occurs at the passive continental margin causes a gravitational instability. If a fractured plate is loaded by thick sediments and denser material at one of its passive continental margins, it develops further gravitational instability, and subduction might begin. Once the passive continental margin has started to subduct, the process will be self-sustaining if there is sufficient horizontal push.

2. STRESSES

The mechanisms causing the accumulation of stress at passive continental margins are: (i) lateral density differences and variable crustal thickness (Bott and Dean, 1972), termed differential loading in this thesis, (ii) temperature changes (e.g. Sleep, 1971; Sleep and Snell, 1976; Turcotte and Oxburgh, 1976; Turcotte et al., 1977a), (iii) sedimentation, erosion, and glacial loading and unloading (e.g. Walcott, 1972; Neugebauer and Spohn, 1978; Hasegawa, et al., 1979), and (iv) plate-driving forces such as mantle convection currents and plate boundary forces.

2.1. *Rheology of the lithosphere*

The rheology of lithosphere (whose thickness ranges from 75 km under oceans to 110 km or more under continents, cf. Walcott, 1970b) is directly related to the state of stress in the lithosphere. It will not only depend on rock type, pressure and temperature but also on strain and strain rate. Based on the calculation of flexural stresses, a rheological zonation of oceanic lithosphere with a point load has been proposed by Beaumont (1979). Flexure-induced stresses are compared with an olivine deformation map (Fig. 2.1). The results show that cataclastic flow is to be expected in the upper 15 km of the plate and ductile plastic flow is predicted for the

Prediction of the Rheology of the Flexed Lithosphere
from Olivine Deformation Maps

Rheological Zonation of the Lithosphere

Fig. 2.1. Rheological zonation of oceanic lithosphere with a point load. The olivine deformation map (from Ashby and Verrall, 1977) is for a 0.1 mm grain size. Dashed lines are strain rates. The base of the lithosphere is defined by a temperature of $1300^{\circ}C$. Short dashed lines are lithospheric stress profiles (from Beaumont, 1979).

lowermost 20 km of lithosphere. The model is similar to that proposed by Bott (1973) based on seismic data. In Bott's model, crustal thinning at a passive continental margin is caused by oceanward hot creep of the lower crust (Fig. 2.2).

Fig. 2.2. Diagram showing lower crustal creep at a passive continental margin. The oceanward creep of the lower continental crustal material is accompanied by wedge subsidence of the brittle layer above (from Bott, 1973).

Beaumont (1979) has given a maximum estimate of approximately 50 km for the elastic layer of the lithosphere (see Fig. 2.1). The elastic thickness, however, is variable, depending on age and thermal properties of the lithosphere and on load factors (volume, rate, duration: cf. also Watts, 1978). The elastic thickness could be as low as 20-30 km (e.g. Kuszir and Bott, 1977; Neugebauer and Spohn, 1978). Keen et al. (in press) presented a more detailed rheological

lithospheric model for passive continental margins. Beneath the thick sediment layer there is a thermal lithosphere which is separated from the asthenosphere by the temperature $T_b = 1350^\circ\text{C}$. The thermal lithosphere consists of an elastic layer and a plastic layer separated by another temperature boundary, T_r ($450^\circ\text{C} \leq T_r \leq 750^\circ\text{C}$). The elastic layer is subdivided into a brittle-elastic layer (cataclastic flow) on top, and a layer underneath which has an elastic-ductile rheology.

The yield stress (i.e. strength) for the elastic layer depends on material properties and environmental conditions. Estimates range from as low as 100 MPa (Turcotte and Oxburgh, 1976) to as high as 1,000 MPa (Turcotte *et al.*, 1977a). At greater depths, material deforms by plastic flow, and the stress depends on the strain rate.

2.2. Differential loading stresses

Bott and Dean (1972) were the first to use the finite element method to analyze stresses due to lateral density variations and variable crustal thicknesses at passive continental margins. The stress distribution was considered to depend only on lateral density differences and variations in the crustal thickness. The shear stress profiles obtained from their calculations show that the greatest shear stress amounts to more than 30 MPa and occurs at a depth of about 10 km below the margin. For comparison, I have analyzed a two-dimensional lithospheric model 40 km thick by 397.5 km

long. The model is subjected to the topography-dependent gravitational force and is characterized by a purely elastic rheology. It is analyzed by the finite element method (for a detailed exposition of the method, cf. Zienkiewicz, 1971). It is subdivided into different layers with different thicknesses and material properties. Fig. 2.3 and Table 2.1 show the model and the parameters used in the computation of stress distribution.

Table 2.1

1. Thicknesses and layers of the model

average depth of the ocean	4.5 km
sediment thickness in ocean	1.5 km
" " in slope-rise	10.0 km
" " in shelf	5.5 km
crustal thickness in ocean	5.5 km
" " in continent	31.5 km

2. Densities (kg/m^3) and Young's Moduli (10^5 MPa)

zone 1	2400	0.4
" 2	2850	1.0
" 3	2850	1.3
" 4	2850	1.1
" 5	3320	1.0

3. Poisson's ratios (estimated from the available seismic velocities of longitudinal and transverse waves)

zone 1	0.27
" 2	0.28
" 3	0.29
" 4	0.26
" 5	0.25

Fig. 2.3. Cross-section of the model of a rifted passive continental margin.

The stresses in the model are produced by variations in body force (gravity) due to topography and to density variations. The base of the structure is allowed to move only in the horizontal direction, the top is taken to be free, and the sides are allowed to move only vertically. The model has two constrained basal corner nodes which are fixed horizontally and vertically. The computation ignores initial and thermal stresses. Fig. 2.4 (in back pocket) shows the resulting stress distribution. Since the Poisson's ratios are about 0.25 throughout the model, the horizontal stresses are approximately one third of the vertical stresses. The maximum and minimum principal stresses are approximately vertical and horizontal respectively.

Fig. 2.5 shows the profiles of maximum shear stresses in Sections A, B, C and D of Fig. 2.3. They are calculated according to the formula $\tau_{\max} = \frac{1}{2} (\sigma_1 - \sigma_3)$ where σ_1 and σ_3 are principal maximum and minimum stress respectively. The maximum shear stresses are about 45° from the vertical. Section A is in the area of the continental shelf; Section B is in the area of the continental slope but near the shelf edge; Section C is also in the area of the continental slope but closer to the slope-rise break; Section D represents the oceanic crust. The shaded area in Fig. 2.5 roughly shows the elastic crustal strength ($\sigma_{\text{yield}} \geq 100 \text{ MPa}$). Actually the crustal strength increases with pressure, but decreases with rising temperature, higher incidence of fractures, and increasing pore water pressure. The reason why these values are used is that

Fig. 2.5. Profiles of maximum shear stress in different sections of the margin. The shaded area shows the probable range for the crustal strength. The stresses are calculated according to the parameters of Table 2.1.

Fig. 2.6. Variations of horizontal deviatoric tensile stress with depth in different sections. The shaded area represents the oceanward differential spreading force.

the stress level required to maintain the topography of mountains is of the order of 100 MPa (Turcotte and Oxburgh, 1976) and the maximum laboratory strength of rock can reach 1,000 MPa (Griggs et al., 1960). The ductile layer beneath the elastic layer with a temperature boundary $T_r = 450^\circ\text{C}$, plays an important role in the process of stress relaxation. In the ductile layer, the crustal strength rapidly decreases to a negligible amount. For instance, Walcott (1970b) described the strength of the asthenosphere as being of the order of one MPa or less.

From three-dimensional considerations, the horizontal deviatoric stresses can be expressed by the tensorial formula $\sigma'_{ij} = \sigma_{ij} - \frac{1}{3} \sigma_{kk} \delta_{ij}$ where δ_{ij} is the Kronecker tensor. It can be simplified to: $\sigma'_{xx} = \frac{1}{2} (\sigma_{xx} - \sigma_{zz})$ in the horizontal direction for a two-dimensional problem. If $\sigma_{xx} > \sigma_{zz}$, then there is horizontal deviatoric compression, and in the opposite case there is horizontal deviatoric tension. Thus, when the horizontal stress is less than the vertical stress, horizontal deviatoric tension results. Since the horizontal stress is approximately 1/3 of vertical stress in the model, there can only be horizontal deviatoric tension throughout the model. Fig. 2.6 shows the horizontal deviatoric tension as calculated from the model. Because maximum and minimum principal stresses are approximately coincident with vertical and horizontal stresses respectively, values in Fig. 2.6 are similar to those in Fig. 2.5. There is a stronger tendency towards deviatoric horizontal tension in the continental shelf section than in the oceanic section because of the topographic relief and the density differences between these sections. The

horizontal deviatoric stress difference is approximately 20 MPa over a depth range of about 20 km. This results in a spreading stress at the continental shelf caused by differential loading relative to the oceanic crust. This result corresponds to the preferential normal faulting on the continental side described by Bott and Dean (1972). The total spreading force can be calculated by the integration of the deviatoric stress difference with depth: $\int_0^{Z_{\text{shatic}}} \Delta \sigma'_{xx} dz$. An alternative way to simplify the integration is to measure the shaded area of Fig.2.6 with the help of grid paper. The oceanward shelf spreading force has a magnitude of about 4.0×10^{11} N/m. It should be noted that ridge pushing forces of about 1.07×10^{12} N/m (Harper, 1975) and up to 2.5×10^{12} N/m (Jacoby, 1970) have been calculated. Therefore the oceanic lithosphere will be compressively stressed by ridge push from the oceanic side and by the gravitational spreading force from the continental shelf. The conclusion that the oceanic crust is under compressive stress is supported by fault plane solution studies (Sykes and Sbar, 1973; Sykes, 1978).

Hasegawa et al. (1979) have studied the effects of varying the geometry of uncompensated sediment wedges on the stress distribution. A maximum distributed normal load of about 20 MPa was used. The deviatoric stress is variable with depth, with the shape of the sedimentary load, and with the distance from the locus of loading. The horizontal deviatoric stress peak, ranging from 10 - 15 MPa at the central line of loading, is always located at a depth of about 40 km. This depth corresponds to the seismicity in the Beaufort Sea.

Hasegawa's deviatoric horizontal stresses, induced by loading from the top, can be superimposed on the stresses resulting from differential loading as calculated in this thesis to obtain a higher level of deviatoric stress.

2.3. Thermal stresses and thermal effects

Thermal effects are among the most important stress contributors. A plate expands when it is initially heated during or before the rifting stage; however, as it moves away from the oceanic ridge, it cools and contracts. These phenomena play a significant role in generating the thermal stress distribution at passive continental margins. Sleep (1971), Sleep and Snell (1976), and Turcotte *et al.* (1977a) have shown that the lithosphere is initially uplifted by thermal expansion and subsequently thinned by subaerial erosion of the resultant uplift and by subcrustal thinning (e.g. necking due to tension, and subcrustal current erosion). Following this initial phase, oceanic lithosphere subsides due to cooling, and due to underplating thickening (Dewey and Kidd, 1977), as it moves away from the oceanic ridge.

2.3.1. Temperature distribution within the plate

In order to understand the thermal effects, it is necessary to analyze the temperature distribution within the plate. The hot

material intrudes along the axis of the ridge with a temperature T_m (cf. Sleep, 1969, for a detailed analysis), and cools by conduction to the surface. As the oceanic lithosphere becomes older and moves farther away from the ridge, its temperature decreases. Thus, applying the theory of thermal diffusion to a two-dimensional model, the temperature distribution within the plate can be seen as a function of depth and distance from ridge; this distance being directly related to the age t (or to use a symbolic expression: $T(X(t), Z)$, where X symbolizes the horizontal direction, and Z the vertical direction). Instead of using the involved classical method to solve the temperature distribution problem, Turcotte and Ahern (1977) used a simpler method (similarity solution method) to obtain the temperature distribution for the cooling plate. On the basis of analogy to three-dimensional heat transfer in a rectangular parallelepiped (Carslaw and Jaeger, 1959), Turcotte and Ahern (1977) considered the subsidence of the basement with a sediment load, and concluded that the temperature distribution should satisfy the following equations:

$$\frac{\partial T_1}{\partial t} + v \frac{\partial T_1}{\partial z} = \kappa_s \frac{\partial^2 T_1}{\partial z^2} \quad 2.1$$

$$\frac{\partial T_2}{\partial t} + v \frac{\partial T_2}{\partial z} = \kappa_m \frac{\partial^2 T_2}{\partial z^2} \quad 2.2$$

where the subscript 1 refers to the sediments and the subscript 2 to the basement, t = time, T = temperature, κ_s and κ_m = thermal

diffusivities in sediment and basement respectively, v = subsidence rate, and z = depth (positive downward).

The temperature distribution solution is given by a error function (Abramowitz and Stegun, 1965) based on the initial, boundary, and matching conditions between sediment and basement:

$$\frac{T_2 - T_1}{T_m - T_1} = \text{erf} \left\{ \frac{z}{2[\kappa_m(t-t_0)]^{1/2}} \right\} \quad 2.3$$

Since thermal stresses develop at relatively low temperatures ($T_p \leq 450^\circ\text{C}$), equation 2.3 can be simplified as a linear expansion, that is

$$\begin{aligned} T_2 &= T_1 + \frac{(T_m - T_1)z}{[\pi\kappa(t-t_0)]^{1/2}} \\ &= T_1 + \beta_{(t)} z \end{aligned} \quad 2.4$$

where

$$\beta_{(t)} = \frac{(T_m - T_1)}{[\pi\kappa(t-t_0)]^{1/2}}$$

Other calculations, such as those by Sleep (1971), reach essentially similar conclusion. The cooling of the lithosphere generates thermal stresses when the temperature becomes sufficiently low to prevent relaxation by flow.

2.3.2. Thermal stresses

Thermal stresses have been calculated by many authors such as Sleep (1971), Turcotte (1974), Turcotte and Oxburgh (1976), Sleep and Snell (1976), Turcotte et al. (1977a). As the lithosphere cools through a yield temperature T_p , the thermal stresses are frozen in and retained. Turcotte et al. (1977a) proposed a way to calculate thermal stress based on plane stress elasticity where the mean horizontal stress parallel to and perpendicular to the ridge is zero, assuming horizontal strain and no flexure. Due to the changes of temperature T and horizontal strain ϵ , the thermal stress is governed by

$$d\sigma = \frac{E}{(1 - \nu)} (d\epsilon + \alpha_L dT_2) \quad 2.5$$

where E = Young's Modulus, ν = Poisson's ratio, α_L = linear coefficient of thermal expansion of the lithosphere, and σ = stress. On the assumption that the total mean stress is zero for the region $T \leq T_p$ (or $z \leq z_p$), the general solution is:

$$\sigma = \frac{E \alpha_L}{1 - \nu} \left[z\beta - \frac{1}{2} (T_p - T_s) \ln\beta \right] + C \quad 2.6$$

where T_s = surface temperature, and C = integral constant.

Assuming in a plane stress case that $T_m = 1100^\circ\text{C}$, $T_s = 0^\circ\text{C}$, $E = 6.0 \times 10^4$ MPa, $\nu = 0.25$, $\alpha_L = 10^{-5}^\circ\text{C}^{-1}$, $\beta = 12^\circ\text{C/km}$, and

$T_p = 300^\circ\text{C}$, the surface thermal stress is 190 MPa. The stress, however, can be as high as 320 MPa if T_p is taken as 700°C . The thermal effect could be an important contributor to stress in a plate. In addition to differential loading stress, sedimentary loading stress, plate driving induced stress, and flexural stress.

Stress induced by the addition or removal of overburden, which is also associated with thermal effects, was first proposed by Haxby and Turcotte (1976). Turcotte and Oxburgh (1976) introduced a formula to calculate the induced stress and then concluded that stress due to overburden tends to cancel thermal stress; the overburden stress is compressive and the thermal stress is tensional. This provides the way to relieve the high thermal stress as well as cracking.

Thermal effects also induce stress by increasing gravitational loads. Sleep and Snell (1976) proposed a method to calculate the increase in weight per unit area of a column of lithosphere due to isotropic thermal contraction. The relationship is

$$P = -\frac{2}{3} \alpha T_e \rho g z \quad 2.7$$

where P = weight increase per area of a column of lithosphere due to thermal contraction

α = volume coefficient of thermal expansion

T_e = average excess temperature

z = depth

ρ = density

The density is assumed to be constant when temperature decreases.

The load resulting from increasing weight generates stress. Assuming a loading at a given time, the governing formula for horizontal deviatoric stress is (Jaeger and Cool, 1969)

$$\sigma_{xx} - \sigma_{zz} = (2Py\pi) \sin(\theta_1 - \theta_2) \cos(\theta_1 + \theta_2) \quad 2.8$$

where θ_1 and θ_2 (for the right and left respectively) are the clockwise angles formed at the two edges of the strip by the horizontal and the direction towards the point for which knowledge of the stress is required.

In fact, the calculation of horizontal deviatoric stress is more complicated because the temperature decays with time.

2.3.3. Subsidence

Keen et al. (in press) classify tectonic subsidence of passive continental margins into two types. Initial subsidence, as discussed for instance by McKenzie (1978), results from isostatic adjustment in the vicinity of anomalously dense material emplaced when continental lithosphere is extensionally necked and partly replaced by denser material from the asthenosphere. According to McKenzie's model, the crust is first thinned before the onset of sea-floor spreading. After initial subsidence, thermal subsidence occurs as the newly formed plate cools, contracts and thickens (Sleep, 1971). McKenzie's (1978) model also shows that the total subsidence due to thermal

effects is dependent on the degree of initial stretching. These explanations shed some light on the observations from present day passive continental margins which are characterized by a zone of thinned continental crust extending for some hundred kilometers across the ocean-continent boundary (Keen and Hyndman, 1979), and by the occurrence of thick sedimentary wedges at the shelf-slope-rise system.

The subsidence history of a rifted continental margin is naturally recorded in the marine sediments deposited upon it after continental separation has occurred. Well data (cf. Keen and Hyndman, 1979) show that subsidence is most rapid soon after initial rifting and slows as time progresses. Assuming that there is no sediment loading and that cooling and subsidence commence at time = 0, the subsidence is governed by the equation (Turcotte and Ahern, 1977):

$$Z(\rho_w - \rho_m) + \int_0^\infty \rho_m \alpha_m (T_m - T_2) dz = 0 \quad 2.9$$

where Z is the subsidence of the oceanic lithosphere.

Substitution of 2.4 into 2.9 and integration give

$$Z = \frac{2 \rho_m \alpha_m (T_m - T_s)}{(\rho_m - \rho_w)} \left(\frac{\kappa_m t}{\pi} \right)^{1/2} \quad 2.10$$

where the subscript m refers to the mantle, subscript w to the ocean water, t is time, and T_s is surface temperature.

Keen (1979) proposed a similar simple equation

$$Z = (300 \pm 80) t^{1/2} \quad 2.11$$

where Z is in metres, t in Ma,

to estimate the amount of subsidence of the Nova Scotia margin.

Fig. 2.7 shows the subsidence of the Nova Scotia margin with $(\text{time})^{1/2}$.

Fig. 2.7. Subsidence curve of the Nova Scotia margin. Open circles represent the subsidence of oceanic crust in the Atlantic Ocean calculated by Parsons and Sclater (1977); black dots are well data from Sable Island (from Keen and Hyndman, 1979).

Equation 2.11 shows that the thermal subsidence of a plate will increase with the square root of geological time. There is, however, a break in the curve at 50-80 Ma. (cf. also Sleep, 1971; King et al., 1975). Other mechanisms, such as normal faulting (Bott and Dean, 1972), phase changes in the lower crust (Neugebauer and Spohn, 1978; Spohn and Neugebauer, 1978; Beaumont, 1979) and sedimentary loading also contribute to passive continental margin subsidence. Subsidence, therefore, can be considered a function of two main processes. Firstly, cooling causes lithosphere contraction and densification resulting in subsidence. Secondly, accumulating sediments act as a load, depressing the lithosphere and allowing more sediments to be deposited. The total thickness of sediments at mature passive continental margin can be well above 10 km (King et al., 1975; Keen and Hyndman, 1979).

2.4. Flexural stresses induced by sediment or ice

2.4.1. Formulae governing vertical displacement

Nadai (1963) determined the vertical deflection of a viscoelastic slab subjected to a system of loading, including a distributed pressure load P and substratum-reaction pressure $q = -kw$, where k is an empirical constant, and w is the vertical deflection. The substratum is assumed to have a slightly greater density and to behave as a fluid. The equation for the vertical deflection is, in the case of a one-dimensional load,

$$D \frac{\partial^4 \dot{W}}{\partial X^4} = -k \left(\dot{W} + \frac{W}{t_e} \right) + \dot{P} + \frac{P}{t_e} \quad 2.12$$

where X = horizontal direction

W = vertical deflection

D = flexural rigidity = $\frac{Eh^3}{12(1-\nu^2)}$

h = thickness of the slab

ν = Poisson's ratio

t_e = viscoelastic decay time = $\frac{3\eta}{E}$

E = Young's Modulus

η = viscosity of the substratum

P = load

and dots denote time derivatives.

The deflection equation for an elastic slab resting on a fluid substratum has been simplified by Walcott (1970b)

$$D \frac{d^4 W}{dx^4} + (\rho_m - \rho_c) Wg = P \quad 2.13$$

where ρ_m and ρ_c are densities of the substratum and of the slab, respectively, and g is the acceleration of gravity.

The general solution for Equation 2.13 is

$$W = W_0 \exp \left(-\frac{X}{\alpha} \right) \left(\cos \frac{X}{\alpha} + \sin \frac{X}{\alpha} \right) \quad 2.14$$

where $\alpha^4 = \frac{4D}{r_s} = \frac{Eh^3}{3(1-\nu^2)r_s}$ = flexural parameter

$r_s = \Delta \rho g$

$\Delta \rho = \rho_m - \rho_c$ = density difference

$W_0 = \frac{P}{4\alpha^2 \rho_c g}$ = maximum deflection of the slab

The exact solution for the deflection depends on the boundary and initial conditions, and on loading, such as constant loads or periodic function loads.

The top of the lithosphere can be subjected to a combination of loads produced by water, sediment or ice. A variety of solutions for different loading conditions have been presented by Walcott (1970b), Turcotte et al. (1977a), and Stein et al. (1979), among others.

Flexure is a very important feature of the vertical deformation of the lithosphere. It is characterized by flexural rigidity of the plate and its thickness. The flexural rigidity of the lithosphere is deduced from observations of the wavelength and amplitude of bending in the vicinity of supercrustal loads. Walcott (1970a) described a way to calculate the flexural rigidity from these observations. Given the flexure of a floating elastic sheet due to a load P , the flexural parameter α defines the wavelength $\lambda = \pi\alpha$ and, in part, the amplitude $W_0 = \frac{P}{4\alpha^2\rho_c g}$. Once λ is known from the observations, the flexural parameter α can be found.

According to the definition of the flexural parameter $\alpha = \left[\frac{Eh^3}{3(1 - \nu^2)\Delta\rho g} \right]^{1/4}$, the thickness of the plate h can be defined. The deformation amplitude W_0 can also be found by considering the density of the lithosphere and the flexural rigidity. Comparing the deformation as inferred from the observations with the deformation computed from an assumed load and different flexural parameters, the best-fit flexural rigidity and thickness can therefore be found.

2.4.2. Flexural stresses

The strains due to vertical displacement in the case of plane strain and one-dimensional load are (Beaumont, 1979)

$$\begin{aligned}\epsilon_{xx} &= -Z \frac{\partial^2 W}{\partial X^2} \\ \epsilon_{zz} &= \left(\frac{\nu}{1 - \nu} \right) \epsilon_{xx}\end{aligned}\quad 2.15$$

where twice differentiating 2.14 gives

$$\frac{\partial^2 W}{\partial X^2} = 2 W_0 \frac{1}{a^2} \exp\left(-\frac{X}{a}\right) \left[\sin\left(\frac{X}{a}\right) - \cos\left(\frac{X}{a}\right)\right] \quad 2.16$$

The significant stresses are in the horizontal direction, and can be calculated by the stress-strain relationship:

$$\begin{aligned}\sigma_{xx} &= \frac{E}{1 - \nu^2} \epsilon_{xx} = - \frac{EZ}{1 - \nu^2} \frac{\partial^2 W}{\partial X^2} \\ \sigma_{zz} &= \nu \sigma_{xx} = - \frac{\nu EZ}{1 - \nu^2} \frac{\partial^2 W}{\partial X^2}\end{aligned}\quad 2.17$$

The corresponding deviatoric stress tensor,

$$\sigma'_{ij} = \sigma_{ij} - \frac{1}{3} \sigma_{kk} \delta_{ij}$$

for the two-dimensional case is:

$$\sigma'_{xx} = \frac{1}{2} (\sigma_{xx} - \sigma_{zz}) \quad 2.18$$

If the origin of the coordinate system (X, Z) is taken at the centre of the plate, $x, z = X, Z$ and all deviatoric stresses reach a relative maximum at the surface of the plate $Z = \pm \frac{d}{2}$, and are zero on the neutral surface $Z = 0$. Fig. 2.8 shows the stress distribution with depth within the lithospheric plate due to the vertical deflection W .

Fig. 2.8. Stress distribution with depth within the lithospheric plate due to the vertical deflection W . Stresses decrease downward from the upper surface of the plate, are zero on the neutral surface, and increase with opposite sign to the same maximum at the base of the plate (from Beaumont, 1979). Note that the state of stress depends on position relative to the axis of the load.

Stresses decrease downward from the upper surface of the plate, are zero on the neutral surface, and increase with opposite sign to the same maximum at the base of the plate. The absolute stress maximum $\sigma_{xx(\max)}$ occurs at $(0, \pm \frac{d}{2})$ and is (Beaumont, 1979):

$$\begin{aligned}\sigma_{xx(\max)} &= \pm \frac{d}{2} \left(\frac{E}{1 - \nu^2} \right) \left[2 W_0 \left(\frac{1}{a^2} \right) \right] \\ &= \pm d \left(\frac{E W_0}{1 - \nu^2} \right) \left(\frac{r_s}{4D} \right)^{1/2} \\ &= \frac{3}{2} \sigma'_{xx(\max)}\end{aligned}\tag{2.19}$$

Beside the previous discussion about the stress distribution with depth, Walcott (1972) presented a horizontal flexural stress distribution for a model loaded by an oceanward prograding sedimentary wedge; the maximum tensile flexural stress occurs near the edge of the continental shelf, and the maximum width of the sediment load necessary to cause elastic failure is about 200 km to 300 km. In order to produce the highest flexural stress at the base of the lithosphere, the maximum width of sediment load can not exceed 300 km in Walcott's (1972) estimates because for greater widths flexural stress decreases due to the dispersion of the load. These sediments give rise to a positive free-air gravity anomaly. The thickest part of the sediment wedge coincides with the highest positive free-air gravity anomaly and the highest compressive downward flexural stresses at the top of the lithosphere. The bending stress at the base of the lithosphere beneath the point of maximum flexure would be about

100 MPa, using a lithospheric thickness of 100 km, and Young's modulus 10^5 MPa. This kind of stress level may be sufficient to produce elastic failure. Turcotte *et al.* (1977a) calculated the flexural stresses at a passive continental margin based on slightly different deflection equations. The stress would be 50 MPa at the edge of the subsiding passive continental margin, assuming:
 $E = 6.0 \times 10^4$ MPa, $\nu = 0.25$, $\rho_s = 2,200 \text{ kg/m}^3$, $\rho_c = 2,800 \text{ kg/m}^3$,
 $\rho_m = 3,300 \text{ kg/m}^3$, thickness of lithosphere = 30 km, and W' (initial depth) = 2.25 km.

2.5. *Driving forces*

The forces acting on a plate include a push from the ridge, essentially due to an increase of the thickness of the lithosphere with time (Jacoby, 1970), a drag force from flow in the upper mantle, and a pull force from the subducting slab if one is attached to the plate in question. In order to understand the compression induced by the horizontal force at a passive continental margin, Neugebauer and Spohn (1978) described creep deformation in a lithosphere stressed by horizontal loading of 0.4×10^{10} N/m from the ridge (assuming that the continental margin is fixed). Apart from vertical loading (which was discussed previously in Sections 2.2, 2.3, and 2.4), Neugebauer and Spohn (1978) showed that increasing horizontal loading (such as ridge push) tends to reduce the width of the subsiding rise, and to cause uplift of the landward continental crust and of the

seaward oceanic lithosphere due to the bending effects of the lithosphere (Watts and Talwani, 1974). These phenomena suggest that stress will be concentrated at the continental rise.

Sykes and Sbar (1973) and Sykes (1978) have argued that the intra-plate state of stress is primarily due to the forces driving the plate. These forces might re-activate pre-existing fault systems. It seems that the driving forces might contribute significantly to the generation of stresses in the lithosphere, although a quantitative estimate of their effect is difficult.

2.6. *Synopsis of stresses and forces*

The spreading stress induced by differential gravitational loading on the continental side of the passive continental margin is about 20 MPa greater than that on the oceanic side. The horizontal spreading force can therefore be estimated at about 4×10^{11} N/m. Without structural flexure, sedimentary loading on top of a passive continental margin produces stress profiles with a maximum horizontal tensile stress of about 10 - 15 MPa between depths of 40 - 50 km. Actually, the elastic lithosphere would become flexured in time, due to the great accumulation of sediments. Thus the state of stress would be dominantly controlled by the flexural stress. Beneath the maximum deflection, the stress

profiles show that the compressive stresses at the top surface of the lithosphere range from about 100 MPa (Walcott, 1972), (300 MPa in Cloetingh et al., in press), up to a maximum of 1,000 MPa (Beaumont, 1979). These values decrease with depth to zero at the neutral surface, where tensile stresses begin to increase with depth, finally to reach the same magnitude as the compressional stress at the top of the lithosphere. The horizontal variations of flexural stress show that the maximum tensile flexural stress occurs near the edge of the continental shelf (Walcott, 1972; Turcotte et al., 1977a). The stress magnitude at the edges of the continental shelf was estimated by Turcotte et al. (1977a) to be 50 MPa. Stein et al. (1979) estimated that flexural tensile stress due to post-glacial rebound is approximately 10 - 15 MPa.

Thermal stresses are more complex than the stresses from other sources. Turcotte and Oxburgh (1976) and Turcotte et al. (1977a) argued on the basis of plane strain and plane stress elasticity that thermal stresses might have a significant effect on the evolving plate. However, thermal stress could be released by fracturing or faulting due to lithospheric cooling and erosion of the uplifted crust in the early stages of sea-floor spreading.

Plate-tectonic forces might contribute a significant stress at passive continental margins. The flexured passive continental margin, characterized by thick sediments at the continental rise, would be compressed by a horizontal ridge push. A force as low as 0.4×10^{10} N/m (Neugebauer and Spohn, 1978) would be sufficient.

3. SEISMICITY AT PASSIVE CONTINENTAL MARGINS OF CANADA

Earthquakes are induced by the sudden release of stress. Barazangi and Dorman (1968) presented a worldwide seismicity distribution for the period 1961 through 1967. The global seismic belts are located along plate boundaries. There is, however, seismic activity occurring within the plates, which is termed intraplate seismicity. Intraplate seismicity along passive continental margins is interesting because it may shed some light on the state of stress there. Two seismic areas along passive continental margins around eastern and northern Canada are discussed here: the Canadian Arctic continental margin and the continental margin of eastern Canada.

3.1. The seismicity of the Canadian Arctic continental margin

Fig. 3.1 shows the general features of this area. Before seismic stations were installed at Mould Bay and Alert in 1961, it was impossible to relate the observed seismic activity to geological features. Stevens (1974) has summarized earthquake patterns in the area up to 1967. Basham *et al.* (1977) have presented a more comprehensive treatment of the seismicity of northern Canada based on a much larger data base obtained mainly from 1962 to 1974.

The landward extension of the Canadian Arctic continental shelf consists mainly of lower Paleozoic rocks overlying Precambrian

Fig. 3.1. Geological provinces and geographic locations of the Arctic Archipelago.
The key for geographic location is:

1. Prince Patrick Island	10. Cornwall Island
2. Melville Island	11. Bathurst Island
3. Mackenzie King Island	12. Cornwallis Island
4. Borden Island	13. Prince of Wales Island
5. Lougheed Island	14. Somerset Island
6. Ellef Ringnes Island	15. Boothia Peninsula
7. Axel Heiberg Island	16. Barrow Strait
8. Amund Ringnes Island	17. Roes Welcome Sound
9. Ellesmere Island	18. Southamoton Island
A. Byam Martin Channel	

(revised from Douglas, 1970).

basement and is referred to as the Arctic Lowlands. The geological structure in the Arctic continental margin consists of thick sedimentary piles, but many diapiric structures are located in the Arctic islands (see Fig. 3.1). In the Late Paleozoic, folding produced the Franklinian Geosyncline with a total sedimentary thickness of about 16 km. In the Late Paleozoic and Mesozoic, 12 km of sediments were deposited in the Sverdrup Basin that was superimposed on the folded Franklinian Geosyncline. In the Mesozoic and Cenozoic, the Prince Patrick Uplift occurred. The Arctic coastal plain composed of gently-dipping late Cenozoic strata consists of a narrow strip of uncompensated sediments that extend from Banks Island to Meighen Island.

From free-air gravity anomalies, Basham et al. (1977) recognized the continental margin break of the Arctic Ocean (Fig. 3.2). These positive gravity anomalies at the marginal break are related to prograded wedges of uncompensated Recent to Tertiary sediments more than 2 km thick along the continental break (see Fig. 3.3). Seismicity patterns in the Canadian Arctic continental margin can be divided into (i) the Yukon-Mackenzie Valley, (ii) the Arctic continental shelf and slope, and (iii) the eastern Canadian Arctic Island (see Fig. 3.4). Up to 1971, nine more stations were installed after MBL and ALT in 1961 and show a more comprehensive seismic pattern. Therefore it is now possible to detect and locate lower magnitude earthquakes in this area.

Fig. 3.2. Free-air gravity anomalies and seismicity for the period 1962-1974 along the Canadian Arctic margin. Bathymetry (in black) is in 500 m contours and free-air gravity (in red) in 20 mGal contours (from Basham *et al.*, 1977).

Fig. 3.3. Explanations for the positive free-air gravity anomaly of passive continental margins. Concept 1 shows the combined effect of crustal thinning and a locally uncompensated ridge in the basement below low density sediments. Concept 2 accounts for the gravity high on the basis of uncompensated dense masses within the basement. Concept 3 postulates accumulation of uncompensated sediments along a prograding continental slope. Isostatic equilibrium is assumed prior to the deposition of sediments; the locally uncompensated recent sediments cause the positive free-air anomaly.

3.1.1. The Yukon-Mackenzie Valley

Fig. 3.4 shows a northwest trending distribution of epicentres in the Yukon Territory and Mackenzie Valley. The epicentres appear to cluster in a rather dense grouping with the Peel River at its centre and a more diffuse grouping to the southeast. Most offshore earthquakes are confined to the continental slope between the 200 and 2,400 m bathymetric contours (Hasegawa et al., 1979). Scattered activity extends westward into Alaska and from the Alaskan coast of the Beaufort Sea to the more highly active region of southeastern Alaska. The horizontal component of the stress field calculated from sedimentary loading in the Beaufort Sea shows that the deviatoric tension is normal to the continental margin, in the same direction as the tensional vector of earthquakes at a depth of 40 km in this area (Leblanc and Wetmiller, 1974; Hasegawa et al., 1979).

3.1.2. The Arctic continental shelf and slope

Earthquakes in this area are associated with large elliptically-shaped free-air gravity anomalies (see Fig. 3.2). They coincide with clusters of epicentres on the continental slope south of Banks Island, west of Ellef Ringnes Islands, and also on the continental shelf off Martin Point on the mainland. Numerous recent earthquakes in this area have magnitudes less than 5 whereas for earlier earthquakes (1920 10 16 off the west coast of Banks Island, 1937 07 05 and 1943 09 07 both north of the Mackenzie Delta) magnitudes greater than 5

have been recorded. Such earthquakes are related to stress which may be built below the thick continental shelf sediments.

3.1.3. The eastern Canadian Arctic Islands

Seismicity is somewhat higher here than that of the shelf-slope area. The seismicity in this area with the Sverdrup Basin as a centre (see Fig. 3.4) is discussed in this section. Two seismic swarms have been recorded in the Arctic Islands. In a 30-day period during March-April 1965 a swarm of about 2,000 microearthquakes were recorded from the southeastern coast of Prince Patrick Island on the west of Sverdrup Basin. The foci lay 15 km east-southeast of Mould Bay at depths of less than 10 km with the magnitudes of -1 to 3 (Smith et al., 1968). The other swarm occurred between November, 1972 to January, 1973 between Melville and Lougheed Islands. These earthquakes are located in the southern Sverdrup Basin. During this period, 67 earthquakes occurred in 55 days. Four of these had recorded magnitudes of between 5 and 6. One of the large earthquakes appeared to have a shallow (9-14 km) focal depth whereas the other three were of an intermediate (27-31 km) depth (Hasegawa, 1977). The activity continued at a reasonably high rate during 1973, was reduced somewhat during 1974 and increased again in December 1974, and was still continuing in 1977 (Basham et al., 1977). The seismic data show that the western edge of the Sverdrup Basin appears to be seismically more

active than the eastern (Fig. 3.4). The seismic data in the central Sverdrup Basin, show that a minor concentration of epicentres exists between Ellef Ringnes and Amund Ringnes Islands and Western Axel Heiberg Islands. The largest, most recent earthquake in the area was of magnitude 5 on 1975 03 08 and occurred on the western coast of Axel Heiberg Island (Basham et al., 1977).

The seismicity of the Arctic Islands archipelago is connected to the south by a concentration of epicenters in Barrow Strait. A more diffuse trend passes through Somerset Island, down the Boothia Peninsula, and thence southeast to a major cluster in Wager Bay and Roes Welcome Sound. This trend, with magnitudes ranging from about 2.5 to 4.5, also goes eastward from Cornwallis Island and Barrow Strait to northern Baffin Island. Seismicity is concentrated on the north coast, as well as in northwestern Baffin Bay and in the Labrador Sea (Fig. 3.4). The seismicity in the margins off Baffin Island is also located at the slope. It appears to match the gravity high anomaly which is caused by the uncompensated sedimentary loading (Keen and Hyndman, 1979).

3.2. The seismicity at the eastern Canadian margin

According to the plate tectonic synthesis of Williams (1979), there was a complete Early Paleozoic tectonic cycle near the present Atlantic continental margin. Williams showed that the Taconic orogeny was the response to a geotectonic cycle through the generation and destruction of a Late Precambrian - Early

Paleozoic Iapetus Ocean. The western part of the orogen consists of the Humber zone that is interpreted as the western side of the ancient Iapetus Ocean; this is followed eastward by the Dunnage zone (vestiges of Iapetus), the Gander zone (eastern margin of Iapetus) and Avalon zone (stable platform or marine shelf). The present Atlantic continental margin started stretching and thinning in the Triassic (about 225 Ma) and opened in the Jurassic (about 200 Ma) in the area adjacent to Nova Scotia with concomitant transform motion along the southern margin of the Grand Banks as Africa and North America separated. The margins off Labrador and Baffin Island were formed by the separation of Greenland and North America during the late Cretaceous and Tertiary (about 130 - 65 Ma). Thus, the margins off eastern Canada exhibit different stages of development of the two main structural types of passive continental margins: transform and rifted. They show typical features of passive continental margins such as zones of attenuated continental crust with possibly faults in the basement, thick piles of sediments and diapirs (Keen and Hyndman, 1979). The margin consists of a miogeocline encompassing the Mesozoic - Cenozoic succession underlying the continental shelf, and a eugeocline comprising strata of similar age overlying the continental rise and abyssal plain (Haworth and Keen, 1979).

As shown in Fig. 3.5, and with reference to the discussion of Section 3.1, there are elliptically-shaped positive free-air gravity anomalies also at the continental break in the margins off

Fig. 3.5. Free-air gravity map of offshore eastern Canada. The dashed line is the 2,000 m contour. Stippled areas denote negative anomalies. Contour interval is 10 mGal. Elliptically-shaped positive anomalies lie along the continental break (from Keen and Hyndman, 1979).

eastern Canada. Fig. 3.6 shows the free-air gravity anomaly profiles from the margins of west Baffin Bay, central Labrador, N.E. Newfoundland, and Nova Scotia. Keen and Hyndman (1979) described positive free-air gravity anomalies that may correspond to the most recent uncompensated sediments of the wedge, as shown in Sobczak's model (1975); these anomalies can also be related to a magnetic anomaly high (see Fig. 3.6).

Fig. 3.7 shows the seismicity along the Canadian passive continental margin from Baffin Island to Newfoundland (Stein et al., 1979). These earthquakes occurred landward of the 3,000 meter contour. There are three seismic clusters in the area. The northernmost, with seismicity on the continental shelf and slope extending to 3,000 m isobath, is the largest both in number of events and magnitudes. It is located along Baffin Bay. Smaller earthquakes in the Labrador Sea generally occur beyond the 1,000 m contour. The third seismic region is located on the Grand Banks, with a smaller number of earthquakes on the continental slope.

During this century, large earthquakes have occurred along the eastern Canadian coast. One of the largest was on 1933 11 20 in Baffin Bay with magnitude 7.3. It is probably the largest intraplate earthquake ever recorded in the eastern North America (Stein et al., 1979). Another of magnitude 7.2 occurred on 1929 11 18 off the Grand Banks. An earthquakes with magnitude 6.2 occurred on 1963 09 04 in Baffin Island and another on 1976 11 12 at Baffin Bay with the magnitude 5.1. All quakes are shallow (less than 33 km) except

Fig. 3.6. Cross-sections along the continental margins of west Baffin Bay, central Labrador, N.E. Newfoundland, and Nova Scotia. The free-air gravity (solid) and magnetic (dashed) anomalies are shown above each section. Velocities in the sections are in km/s. The key for the various shadings shown opposite Profile D is as follows: 1: post-rifting sediments, 2: sediments of Early Jurassic age and older on the Nova Scotian shelf, 3: continental basement rocks, 4: continental crust, 5: oceanic crust, 6: mantle (from Keen and Hyndman, 1979).

Fig. 3.7. Seismicity and focal mechanisms of the eastern Canadian margin. Earthquakes seaward of the 1,000 m contour show thrust faulting; landward events show primarily normal faulting (from Stein *et al.*, 1979).

the large 1933 one with 65 km focal depth.

3.3. The causative forces

Some of the potential causes that produce the stress difference which generates intraplate earthquakes at passive continental margins are plate tectonic forces, isostatic postglacial uplift, sedimentary loads, pre-existing faults, and diapiric structures.

3.3.1. Plate tectonic forces

Since sublithospheric earthquakes are absent, Basham et al. (1977) concluded that tectonic forces characteristic of plate margins are not a direct cause of seismicity in the Canadian Arctic although they may affect it. Fault-plane solution studies show that the principal horizontal stress for the Yukon earthquakes is compressional in the northeast and tensional to the southeast (see Fig. 3.8). This stress might be related to the compressive force exerted by the Nansen-Gakkel spreading ridge (Fig. 3.9). The three seismic clusters in the Beaufort Sea may also be related to compressive plate tectonic forces transmitted from the spreading ridge (Hasegawa et al., 1979). A large earthquake in the Beaufort Sea on 1975 06 14; for example, had magnitude greater than 5. This earthquake shows a deviatoric compression in a north-south direction with a focal depth of 40 km, that could correspond to the ridge

push force. Forces related to tectonism, when transmitted through the Arctic Ocean lithosphere cause seismic activity along specific zones at the passive continental margin off north Yukon. From the intraplate seismic observations along eastern north America, Sykes (1978) proposed that the earthquakes were related to reactivation of pre-existing zones of weakness due to the motion of the plate. A similar mechanism may be operative in the Canadian Arctic.

Fig. 3.8. Minimum horizontal stress direction inferred from mechanism solutions for earthquake in the Yukon Territory (Leblanc and Wetmiller, 1974) (from Basham et al., 1977).

Fig. 3.9. Deviatoric tension (T) and compression (P) vectors for the earthquake (occurred on 1975 06 14) in the Beaufort Sea. The stress vector is related to the northern active spreading Nansen-Gakkel ridge (from Hasegawa *et al.*, 1979).

3.3.2. Postglacial uplift

The uplift results from the disappearance of the Pleistocene ice sheet. Based on the data from 58 sites in the Canadian Arctic, Andrews (1970) found that there are at least three centres of postglacial uplift in northern and eastern North America. They are located over southeastern Hudson Bay, over northern Keewatin, and a smaller centre over northern Queen Elizabeth Islands (see Fig. 3.10). The figure shows also that postglacial uplift rates produce large and rapidly varying crustal stress gradients which may explain the observed seismic patterns in the Sverdrup Basin (Basham et al., 1977). The uplift curves (see Fig. 3.10) suggest that the Baffin Island-Foxe Basin area (northern Keewatin) is rebounding more slowly than Hudson Bay. Tilting to the northeast along Baffin Bay is caused by this doming of the central region. Isostatic rebound, therefore, in addition to the uncompensated sedimentary loading, may influence the pattern of seismic activity in this area.

Most of the east coast of Canada from Baffin Island to the Newfoundland shelf and extending southward to New York was glaciated (Loken, 1971; Stein et al., 1979). Stein and others proposed that the earthquakes are induced by the removal of glacial load. Upon unloading, horizontal extension (normal faulting) is caused by de-flexing in the upper lithosphere in areas formerly covered by ice, while areas not covered by ice undergo a compressional regime and thrust faulting. Fig. 3.11 shows the model used by Stein et al. (1979) to explain the stress patterns for the ice flexure. These

Fig. 3.10. Average rates of postglacial uplift for northern North America (m/100 a). There are three centres of postglacial uplift: southeastern Hudson Bay, northern Keewatin, and northern Queen Elizabeth (from Andrews, 1970).

stresses reactivate basement faults present in the plate margin since the original rifting. The area of horizontal extension matches that of Baffin Island earthquakes. Earthquakes in the northern part of the area show normal faulting on an east-west plane, whereas thrust fault mechanisms are found in the oceanward part of the Labrador Sea.

The flexural stress induced by ice loading and unloading can be calculated, and it is found that the maximum stress occurs at ± 180 km from the ice margin and has a magnitude of 10 MPa at a depth of 20 km, and 15 MPa at the surface (Stein *et al.*, 1979). Although these kinds of magnitudes of stresses are not enough to fracture unfaulted lithosphere, earthquakes could be induced on pre-existing faults remaining from the rifting (e.g. Turcotte *et al.*, 1977a; Keen and Hyndman, 1979; Stein *et al.*, 1979). For a discussion of pre-existing faults, see Section 3.3.4.

Fig. 3.11. Geometry of the lithospheric flexural stresses induced by deglaciation (from Stein *et al.*, 1979)

3.3.3. Sedimentary loads

The maximum sediment thickness in the eastern continental margins of Canada is usually about 6 - 10 km. These sediments correspond to the free-air gravity anomaly high. Keen and Hyndman (1979) pointed out that at least some of this load is uncompensated. Thus, sedimentary loading could contribute to stress on the eastern continental margin, as well as on the Arctic margin.

Figure 3.2 shows the earthquake epicenters along the Arctic Coastal Plain. They are also associated with large free-air gravity anomalies which are caused by the presence of uncompensated sedimentary loads (see Fig. 3.3). The epicentral clusters west of Ellef Ringnes Island, near the southern end of Banks Island and off Martin Point correspond to three large elliptically-shaped free-air anomalies. One free air gravity anomaly that is coincident with the earthquake cluster west of Ellef Ringnes Island is interpreted as a major sediment-filled trough by Bhattacharryya (1968) on the basis of aeromagnetic data.

The detailed seismicity pattern depends upon the particular relationship of the load to the underlying structure. If pre-existing faults that have not experienced movement for a long time approach a critical state of stress, sedimentary loading might trigger earthquakes in underlying basement structures. A sedimentary load stress of 10 - 15 MPa occurs at the depth about 40 km, as calculated by Hasegawa et al. (1979, see Chapter 2). This could be sufficient to induce the earthquakes, if the basement structure is characterized

by pre-existing faults.

An elliptically-shaped positive free-air gravity anomaly denoting a sediment-filled trough is also found in the Beaufort Sea, and it partly corresponds to Beaufort Sea seismicity. Hasegawa et al. (1979) have inferred that the stress regime in the Beaufort Sea consists of horizontal deviatoric tension normal to the strike of continental margin. This stress system is caused by a combination of current (uncompensated loading of Quaternary sediments), remanent (tectonic) stresses and north-south compression generated from the Nansen-Gakkel spreading ridge push transmitted through the Arctic Ocean lithosphere (see Fig. 3.9).

If the basement has no appropriate structures such as pre-existing faults, the occurrence of shocks cannot be explained easily. Basham et al. (1977) suggested that the lack of seismicity in the continental shelf and slope west of Banks Island in the south and Borden Island in the north is due to an absence of appropriate structures or an insufficient load to activate the structures.

3.3.4. Pre-existing faults and other features

Rifting of continents causes faults along plate margins. Due to the differential thermal contraction between oceanic and continental lithosphere, there might be initial vertical decoupling.

Turcotte et al. (1977a) postulate a normal fault with several kilometres offset at the continental-oceanic lithosphere boundary. They also concluded that the evolution of Atlantic-type margins might be controlled dominantly by the amount of vertical displacement along this fault since the fault collects sediments on the oceanic side and also reduces the strength of the passive plate. However, as pointed out by Bott (1979), differential thermal contraction can not account for regional subsidence of more than 4 km. In order to account for thick sedimentary accumulations of 5 to 15 km, some other processes are required.

No matter what their origin is, passive continental margins are usually characterized by fractures, since these faults are preserved in the plate during rifting stage (Turcotte et al., 1977a; Keen and Hyndman, 1979). Earthquakes will be induced more likely where crustal strength is reduced by pre-existing faults. Service and Douglas (1973) described how stresses will be concentrated around the tips of such cracks. Due to crack-tip stress concentration, the structure can be fractured at lower average stresses. Moreover, natural faults often contain gouge of alteration minerals, and their coefficient of friction is strongly dependent on the composition of gouge. If the sliding surfaces are separated by gouge of montmorillonite or vermiculite, friction is very low because water molecules included in the clay lattice act as "pseudo pore pressure" (Byerlee, 1978). Thus the strength of the structure is reduced. In addition, water in fault systems provides lubrication (Scholz, 1977).

that reduces the stress required to cause earthquakes.

Finally, there are widespread diapiric features in the Canadian Arctic Archipelago, on the islands and in the channels. Immature topography indicates recent or ongoing uplift (Tozer, in Douglas, 1970). Because earthquake swarms are commonly related to highly fractured zones and intrusive bodies, earthquake swarms might be induced by these intrusive uplift bodies. Earthquake tremors lasting from November 1972 to January 1973 in the Byam Martin Channel have been attributed to reactivation of pre-existing faults by diapiric structures (Basham et al., 1977).

4. CONVERSION OF PASSIVE INTO ACTIVE CONTINENTAL MARGINS

Following the plate tectonic hypothesis first formulated by Wilson (1965), Dewey (1969) has shown a model for the evolution of a continental margin from a passive into an active type. When subduction occurs at a passive continental margin, the margin becomes active due to the creation of a new plate boundary (see Fig. 4.1). One important mechanism causing the initiation of subduction at a passive continental margin may be gravitational instability (denser material above lighter material) combined with a large flexural stress, and zones of weakness associated with pre-existing faults. However, the mere existence of an unstable vertical density distribution is not a sufficient condition for subduction (Ringwood, 1975; McKenzie, 1977, England and Wortel, 1980). There still needs to be a compressive force to cause the subduction to be self-sustaining (McKenzie, 1977). Thus, subduction can start at the passive continental margin if the gravitational instability is accompanied by a compressive force acting from the ridge and a spreading force (this study) from the continental side.

4.1. Gravitational Instability

There are two possible processes capable of producing the gravitational instability of the lithosphere. One is the penetration of hot denser asthenosphere material through the lithosphere so

Fig. 4.1. Schematic sections along a great circle showing successive relationships between plates, oceans, continents, and orogenic belts. An evolution for a passive plate (Atlantic type) is shown. Passive continental margin evolves into active type where the plate boundary (trench) occurs (from Dewey, 1969).

that it overlies the lighter material when the sea floor is lower than the mantle geoid. This may happen at an early stage of sea-floor spreading due to thermal contraction; the sea floor continuously subsides to a level below the mantle geoid. Besides the increase in density due to temperature decrease of the lithosphere, the other possibility is the phase transition of basalt to eclogite in the lower crust, depending upon the overburden pressure, loading and appropriate temperature for the stability of eclogite. Because eclogite is stable at high pressure (see Fig. 4.2), the transition might occur at a later stage of the evolution of the oceanic plate when the overburden pressure (increasing with age mainly because of the thickening of the sedimentary pile overlying the lithosphere) is sufficiently high (e.g. about 1,000 MPa for 30 km depth).

These two possible mechanisms are examined in the following subsections.

4.1.1. Penetration of asthenospheric material

In terms of mantle convection, Turcotte et al. (1977b) treated the initiation of subduction as a fluid instability when the Rayleigh number $R = \frac{g\alpha\beta\delta^4}{\kappa\nu}$ is greater than 657.5, where g = acceleration of gravity, α = coefficient of thermal expansion, β = temperature gradient, δ = thickness of the layer, κ = thermal diffusivity, and ν = kinematic viscosity. The fluid layer, heated from within or below, becomes unstable and flows when the critical Rayleigh number

Fig. 4.2. The stability fields for rocks having the composition of the quartz-tholeiite variety of basalt (Ringwood and Green, 1966), showing also the typical continental geothermal gradient of Birch (1958) (from, Bott, 1971).

is exceeded. During sea-floor spreading, the upper thermal boundary layer (i.e. the lithosphere) subsides with time because it cools. The mantle geoid (defined as the height to which hot mantle rock from asthenosphere can rise) is roughly 3.25 km below the hydrogeoid (sea level) and at a depth of 0.75 km below the oceanic ridge (Turcotte *et al.*, 1977b). Once the subsiding sea floor lies below the mantle geoid, hot denser asthenospheric material ($\rho_m = 3,300 \text{ kg/m}^3$) can penetrate through pre-existing faults in the lithosphere to overlie the lithosphere (lighter, $\rho_l = 3,000 \text{ kg/m}^3$). Such a process would considerably increase the crustal weight, favour the basalt-eclogite transition, and possibly initiate subduction if pre-existing faults are present at the passive margin. Using equation 2.10 (Turcotte and Ahern, 1977) for the subsidence of the sea floor, Turcotte *et al.* (1977b) found that sea floor older than about 6 Ma is below the mantle geoid. This implies that the weight of the lithosphere begins to increase due to asthenospheric penetration only 6 Ma after initial rifting. However, the density increase depends on the number of fractures in the lithosphere; thus, it might not be an important mechanism for the initiation of subduction if the lithosphere is not highly fractured. Furthermore, no seismic evidence has been found to support this idea.

4.1.2. Basalt-eclogite transition

Ringwood (1975) has proposed a model for the evolution of the lithosphere (Fig. 4.3).

Fig. 4.3. Petrological model for the evolution of a passive continental margin based on Ringwood and Green (1966). Plate consumption and boundary generation are caused by the phase change (basalt-eclogite).

- A. Initial state - e.g. east coast of North America (Drake et al., 1959); rifting cracks preserved.
- B. Anhydrous mafic lower crust transforms to eclogite.
- C. Eclogitic lower crust slowly subsides into mantle causing eugeocline to deepen.

After Phase C, subduction could be initiated due to the denser sinking eclogite depressing the fractured passive continental margin. A Benioff zone is therefore established (from Ringwood, 1975).

This involves a phase transition in the lower crust at a passive continental margin beneath the continental rise, and was developed from the studies of Ringwood and Green (1966) who suggested that the initial sinking of the lithosphere might be caused by an instability generated by the transformation of basalt in the lower crust to eclogite (see Fig. 4.2). Under the P-T conditions existing in the normal crust, dry basaltic rocks, with density about $3,000 \text{ kg/m}^3$ would not be stable and would be transformed into eclogite, possessing a density of $3,450$ to $3,600 \text{ kg/m}^3$. The reaction rate for the phase change is very low (Fyfe and Leonardos, 1977) and conditions of $3,000 \text{ MPa}$ and $1,000$ - $1,250^\circ \text{C}$ were estimated by Green and Ringwood (1967). However, Bott (1971) considered that lower crustal conditions of 300 - 600°C at $1,000 \text{ MPa}$ were sufficient for the transition.

Fyfe and Leonardos (1977) emphasized the point that sediment loading can induce the basalt-eclogite transition in oceanic crust by increasing the overburden pressure. They proposed that if old, cool oceanic lower crust becomes loaded by an overburden of about 20 km (upper crust + sediment + water), then the transition will be initiated. Once initiated, the resulting contraction will promote further loading and cause further sinking. The mechanism is controlled by the slow build-up of sediments depressing the basalt (or gabbro) into the eclogite stability field. Spohn and Neugebauer (1978) concluded that the minimum thickness of sediments necessary to cause the transformation is of approximately 3 km , and that the

rate of the transformation is a function of the rate of sedimentation. Neugebauer and Spohn (1978) calculated that the subsidence of the continental rise increases rapidly in response to the phase transition. Thus, if a large volume of basalt-eclogite is involved, the sinking of the lower crust beneath a passive continental margin becomes significant, but this mechanism does not account for the initiation of subduction itself, unless some form of decoupling between continental and oceanic lithosphere is achieved.

4.2. *The initiation of subduction*

From the arguments developed in the previous section, it is clear that it is not possible to attribute the initiation of subduction to a single mechanism. It seems that the initiation of subduction can be considered as a possible but not necessary result of the aging of the oceanic lithosphere, in which two physical phenomena are particularly important. They are temperature decrease and density increase, which induce an instability roughly proportional to the time elapsed since rifting. For example, Wortel (1980, quoted by Cloetingh et al. in press) calculated that oceanic lithosphere becomes unstable for ages older than 30 Ma. As the lithosphere cools, the temperature decrease causes thermal contraction and subsidence which can accommodate thicker sediments at a passive

continental margin. The negative downward buoyancy force is also increased by the thermal contraction. In addition, sea floor subsidence below the mantle geoid after about 6 Ma might cause asthenospheric material to penetrate and overlie the lithosphere. This instability not only tends to depress the lower crust, but also increases the loading (pressure) to enhance the basalt-eclogite phase change. The tendency to depress the lower crust would increase the reaction rate of the phase change because of higher temperatures. Assuming the minimum thickness of sediments necessary to cause the phase transition is 3 km (Spohn and Neugebauer, 1978), and that sedimentation has kept up with the subsidence of the cooling oceanic lithosphere (Turcotte and Ahern, 1977), the timing of the initiation of the phase change can be roughly estimated. Based on equation 2.11 ($Z = (300 \pm 80)t^{1/2}$, where Z is depth in metres, and t is age in Ma) for the subsidence of the Nova Scotia margin, this critical age ranges from about 80 to 130 Ma. However, the thickness of the sediments can also be increased due to flexure, and the amplitude of flexure is greater in a thinner lithosphere; therefore, the phase transition might begin earlier, depending on the degree of vertical deflection and loading.

Due to the thick sediments, high stresses at the base of the elastic lithosphere (up to 300 MPa, Cloetingh et al., in press) are induced by the flexural effects. Such an accumulation of thick sediments (e.g. a river delta) on young oceanic lithosphere would cause larger flexural stresses than in the case of older lithosphere, since a

younger thinner lithosphere increases stress concentration. These tensional stresses are probably high enough to reactivate the pre-existing marginal fault system and even to fracture lithosphere with a strength ranging from 100 to 1,000 MPa. Probably the decoupling vertical faults between oceanic and continental lithosphere (Turcotte et al., 1977a) play a leading role in the activation of passive continental margins.

It appears, therefore, that subduction could be initiated as a result of the depression of a pre-fractured passive continental margin to a critical degree, due to the combined effects of aging of the oceanic lithosphere, the basalt-eclogite transition, sedimentary loading, and possibly penetration by denser asthenospheric material. The initiation of subduction might occur at any time during the evolution of the oceanic plate, as long as the critical conditions have been reached. It can occur at a very early stage if the still thin passive continental margin lithosphere is sufficiently fractured, flexured, and loaded by sediments. However, under normal conditions the possibility of subduction should increase with the aging of the oceanic lithosphere.

4.3. The maintenance of subduction

After subduction is initiated, it may not be maintained if there is insufficient force to push the subducting plate (McKenzie,

1977; England and Wortel, 1980). The reason is that there is a frictional stress (probably about 10 MPa, McKenzie, 1977) between the sinking slab and the overlying plate, and the resistance of the asthenosphere to lithospheric penetration must be overcome.

McKenzie (1977), considering subduction as a fluid instability, concluded that two factors are needed to maintain a newly formed subduction zone until it becomes self-maintaining. These are (i) the presence of a compressive stress of at least 80 MPa pushing down the slab, and (ii) a subduction rate larger than 1.8 cm/a. If these two factors are successful in producing a slab longer than 180 km and dipping at an angle $\gamma \geq 45^\circ$, the subduction becomes self-maintaining.

The required balance of forces is illustrated in Fig. 4.4 (from McKenzie, 1977). The forces involved are ridge push (F_r), slab pull (F_s), and resisting forces, the latter consisting of frictional force (F_e) between sinking slab and overlying plate, and a force expressing the resistance of the plate to shape changes (F_b). The greatest stress separation between driving stress (σ_d) and resisting stress (σ_r) has a magnitude of about 80 MPa, and at least 130 km of slab must have been subducted (here a dip angle of 90° is assumed) before driving and resisting stress are equal. For longer slabs, the instability is increased with the depth increase of the subducted slab and the subduction becomes self-maintaining. This occurs because the resisting force stops increasing once the friction becomes steady-state and the slab shape has changed whereas the

Fig. 4.4. Sinking slab self-maintaining requirements. X is the length of subducted slab. Stresses are calculated according to the inequality relationship $\sigma_d > \sigma_r$, where $\sigma_d = (F_r + F_s)/T_e$, $\sigma_r = (F_e + F_b)/T_e$, and T_e is lithospheric thickness causing frictional stress (from McKenzie, 1977).

driving force continues to increase as a consequence of the thermal instability of the slab. Thus the instability of the sinking slab grows as the plate driving force exceeds the resisting force. The actual magnitude of the forces involved is, of course, also a function of the age of the slab. Furthermore, to drive subduction, the slab must sink sufficiently rapidly to maintain its temperature contrast with the mantle and therefore its negative buoyancy. The faster the subduction, the larger the instability. According to McKenzie's (1977) calculation, the rate of subduction must exceed 1.8 cm/a.

From the viewpoint of this thesis, it should be noted that the compressive stress can also be induced by the spreading force due to topography. In a Cordilleran region, where the difference in elevation between the mountain chain and the adjoining trench can be 7 km, a compressive stress of 85 MPa can be generated (McKenzie, 1977). According to the calculation in Chapter 2, the differential loading at a passive margin (no relief on the continental side), generates a compressive stress of about 20 MPa in the uppermost 20 km oceanward of the rise, and therefore contributes significantly to the compressive state of stress in the oceanic plate.

5. CONCLUSIONS

The main arguments developed in this thesis are summarized in this Chapter.

5.1. Stress-generating mechanisms at passive continental margins

5.1.1. Stresses induced by differential loading

The analysis by the finite-element model suggests significant stress contributions induced by differential topographic loading. A spreading force per unit width at a continental shelf of about 4×10^{11} N/m is obtained from the spreading stresses caused by the differential loading (about 20 MPa) in the elastic crust (to a depth of 20 km). The spreading force (ridge push) varying from 1×10^{12} to 2×10^{12} N/m (Jacoby, 1970; Richter and McKenzie, 1978) from the other side, originates a compressive state of stress in the oceanic lithosphere, while there may be relative tension on the continental side. This state of stress corresponds to normal faulting as found in the shelf (Bott and Dean, 1972), and gives good agreement with seismological observations in the oceanic lithosphere (Sykes and Sbar, 1973; Sykes, 1978).

5.1.2. Stresses induced by thermal effects

If the temperature of the material drops through a yield temperature T_p ($T_p \leq 450^\circ\text{C}$), the thermal stresses will be frozen in the lithosphere. On the basis of plane stress elasticity, thermal stresses were calculated by Turcotte *et al.* (1977a). Their magnitude is of the order of several hundred MPa, almost the range of crustal strength. This high magnitude stress is probably relieved in the early stages of lithospheric evolution, mainly by fracturing (transform faults).

5.1.3. Stresses induced by flexure due to sediments or ice

The elastic lithosphere at passive continental margins shows flexure under loading by sediments and/or ice. From estimations of the flexural rigidity and the thickness of the elastic layer, flexural stress profiles due to loading were calculated by Beaumont (1979) and horizontal flexural stress variations were computed by Walcott (1972) and Turcotte *et al.* (1977a). The tensile stresses at the base of the lithosphere range from as low as 100 MPa up to a maximum estimate of 1,000 MPa. These induced stresses might be reduced by stress relaxation (Beaumont, 1979). Taking into account the horizontal stress variation, the maximum tensile stress occurring at the edge of a continental shelf (Walcott, 1972) amounts to

50 MPa (Turcotte et al., 1977a). Glacial rebound stresses from 15 MPa at the surface to 10 MPa at 20 km depth were calculated by Stein et al. (1979).

5.1.4. Plate tectonic forces

Stresses generated by the age-dependent ridge push contribute to the evolution of a passive continental margin. They not only compress the passive continental margin in the horizontal direction, but also increase the tensile stress at the base of the lithosphere by increasing the flexure generated by vertical loads. When and if subduction begins, ridge forces contribute to the mechanism by which subduction is maintained.

5.2. State of stress and earthquakes

The stress calculations from the model show that the crustal strength is greater than shear stresses induced by the differential loading within the elastic layer. However, many earthquakes occur at shallow depths due to pre-existing faults and the superposition of stresses of different origin. Pre-existing faults characterized by gouge not only decrease the crustal strength but also concentrate stresses at their tips. These combined effects increase the stress

difference at the faults and decrease the effective strength.

5.2.1. Arctic continental margin

The relatively rapid change in postglacial uplift rate (Andrews, 1970) could be the main mechanism responsible for the greater seismicity occurring in the eastern part of the Canadian Arctic Islands. The rapid changes in postglacial uplift rate imply large and rapidly varying crustal stress gradients in the area (Basham et al., 1977). The seismicity at the Arctic continental shelf and slope can be attributed to wedges of uncompensated sediments as indicated by elliptically-shaped positive free-air gravity anomalies. The load induced by these uncompensated wedges of sediments has possibly acted as a triggering mechanisms for movement on underlying faults which were close to a critical state of stress. Plate tectonic forces transmitted from the Nansen-Gakkel ridge may contribute stresses affecting the scattered northwest-southeast trending seismicity in the Yukon-Mackenzie Valley area.

5.2.2. Eastern Canada continental margin

As in the Arctic, possible causes of the seismicity off eastern Canada include postglacial uplift (Stein et al., 1979), plate tectonic forces (Sykes, 1978), and uncompensated sedimentary loads

(Keen and Hyndman, 1979). A range of stress of 10-15 MPa caused by postglacial uplift has been calculated by Stein et al. (1979). If the lithosphere possesses pre-existing faults, these may induce earthquakes. A similar earthquake-triggering stress may be induced by uncompensated sedimentary loads. On the other hand, on the basis of fault plane solutions of earthquakes occurring at continental margins of eastern north America, Sykes and Sbar (1973) and Sykes (1978) have emphasized that plate tectonic forces (i.e. ridge push) are primarily responsible for the reactivation of pre-existing faults. These mechanisms, of course, are not mutually exclusive.

5.3. *Initiation of subduction*

At present, it is difficult to evaluate subduction-initiating processes. In most cases, initiation of subduction might be the possible but not necessary result of the aging of the oceanic plate. Aging of the lithosphere is accompanied by temperature decrease and density increase, both important factors to gravitational instability at passive continental margins. Subsidence of the sea floor allows the accumulation of thick sedimentary loads. Denser asthenospheric material might penetrate the lighter lithosphere when the sea floor is below the mantle geoid (Turcotte et al., 1977b); this might increase the instability necessary to initiate subduction. The thickening sediment wedge not only favours the basalt-eclogite phase

change, but also produces high magnitude flexural stresses which can reactivate the pre-existing faults at the margin (Ringwood, 1975; Turcotte et al., 1977a). Subduction, then, would begin because of the gravitational instability of the lithosphere and/or the presence of supercrustal loads. If compressive forces at the margin and the convergence rate are above a critical point (McKenzie, 1977), subduction becomes self-sustaining and the passive margin evolves into an active plate boundary.

6. REFERENCES

- Abramowitz, M and Stegun, I.A. (1965) Handbook of Mathematical Functions, pp.295-329, Dover, New York.
- Andrews, J.T. (1970) Present and postglacial rates of uplift for glaciated northern and eastern North America derived from postglacial uplift curves. Canadian Journal of Earth Sciences, V. 7, pp. 703-715.
- Ashby, M.F. and Merrall, R.A. (1977) Micromechanisms of flow and fracture, and their relevance to the rheology of the upper mantle. Philos. Trans. R. Soc. London, Ser. A, V. 288, pp. 59-95.
- Barazangi, M., and Dorman, J (1969) World seismicity map of ESSA Coast and Geodetic Survey epicentre data for 1961-1967. Bull. Seism. Soc. Am., V. 59, p. 369.
- Basham, P.W., Forsyth, D.A. and Wetmiller, R.J. (1977) The seismicity of northern Canada. Canadian Journal of Earth Sciences, V. 14, pp. 1646-1667.
- Beaumont, C. (1979) On the rheological zonation of the lithosphere during flexure. Tectonophysics, V. 59, pp. 347-366.
- Bhattacharryya, B.C. (1968) Analysis of aeromagnetic data over the Arctic Islands and continental shelf of Canada. Geological Survey of Canada, Paper 68-44.
- Birch, F. (1958) Interpretation of the seismic structure of the crust in the light of experimental studies of wave velocities in rocks, In: Benioff, H., Ewing, M., Howell Jr., B.F., and Press, F., eds., Contributions in geophysics in honor of Beno Gutenberg, Pergamon Press, pp. 158-170.
- Bott, M.H.P. (1971) The Interior of the Earth. Edward Arnold, 316p.
- Bott, M.H.P. (1973) Shelf subsidence in relation to the evolution of young continental margins, In: Tarling, D.H. and Runcorn, S.K., eds., Implications of Continental Drift to the Earth Sciences, V.2.
- Bott, M.H.P. (1979) Subsidence Mechanisms at passive continental margins, In: Watkins, J.S., Montadert, L. and Dickerson, P.W., eds., Geological and Geophysical Investigations of Continental Margins. American Association of Petroleum Geologists Memoir, V. 29, pp. 3-9.
- Bott, M.H.P., and Dean, D.S. (1972) Stress systems at young continental margins. Nature, V. 235, No. 54, pp. 23-25.

- Byerlee, J. (1978) Friction of rocks. *Pure and Applied Geophys.*, V. 116, pp. 615-626.
- Carslaw, H.S. and Jaeger, J.C. (1959) *Conduction of Heat in Solids*. 2nd ed., Oxford at the Clarendon Press, London, pps. 6-15, 58-62, 148, 282-291.
- Cloetingh, S.A.P.L., Wortel, M.J.R., and Vlaar, N.J. (1981) Implications of age-dependent properties of the oceanic lithosphere for the state of stress at a passive continental margin. In press.
- Dean, D.S. (1972) *Stress Analysis of the Lithosphere*. Ph.D. thesis, University of Durham, Durham, 289p.
- Dewey, J.F. (1969) Continental margins a model for conversion of Atlantic type to Andean type. *Earth and Planet. Sci. Let.*, V. 6, pp. 189-197.
- Dewey, J.F. and Kidd, W.S.F. (1977) Geometry of plate accretion. *Geological Society of America Bull.*, V. 88, pp. 960-968.
- Douglas, R.J.W., Ed., (1970) *Geology and Economic Minerals of Canada*. Economic Geology Report No. 1.
- Drake, C.L., Ewing, M., and Sutton, G.H. (1959) Continental margins and geosynclines: the east coast of North America north of Cape Hatteras, *Phys. and Chem. of the Earth*, V. 3, pp. 110-198.
- England, P and Wortel, R. (1980) Some consequences of the subduction of young slabs. *Earth and Planet. Sci. Let.*, V. 47, pp. 403-415.
- Fyfe, W.S. and Leonardos Jr., O.H., (1977) Speculations on the causes of crustal rifting and subduction, with applications to the Atlantic Margin of Brasil. *Tectonophysics*, V. 42, pp. 29-36.
- Green, D.H. and Ringwood, A.E. (1967) An experimental investigation of the gabbro to eclogite transformation and its petrological applications. *Geochim. Cosmochim. Acta*, V. 31, pp. 767-833.
- Griggs, D.T., Turner, F.J. and Heard, N.C. (1960) Deformation of rocks at 500° to 800° C, *Geol. Soc. Am. Memoir*, V. 79, pp. 39-104.
- Harper, J.F. (1975) On the driving forces of plate tectonics. *Geophys. J. R. Astr. Soc.*, V. 40, pp. 465-474.
- Hasegawa, H.S. (1977) Focal parameters of four Sverdrup Basin, Arctic Canada, earthquakes in November and December of 1972. *Canadian Journal of Earth Sciences*, V. 14, pp. 2481-2494.

- Hasegawa, H.S., Chou, C.W., and Basham, P.W. (1979) Seismotectonics of Beaufort Sea. *Canadian Journal of Earth Science*, V. 16, pp. 816-830.
- Haworth, R.T. and Keen, C.E. (1979) The Canadian Atlantic Margin: a passive continental margin encompassing an active past. *Tectonophysics*, V. 59, pp. 83-126.
- Haxby, W.F. and Turcotte, D.L. (1976) Stresses induced by the addition or removal of overburden and associated thermal effects. *Geology*, V. 4, pp. 181-184.
- Jacoby, W.R., (1970). Instability in the upper mantle and global plate movements. *J. Geophys. Res.*, V. 75, pp. 5671-5680.
- Jaeger, J.C. and Cook, N.G.W. (1976) *Fundamentals of Rock Mechanics*. Chapman and Hall, John Wiley and Sons, Inc., New York, 585 p.
- Keen, C.E. (1979) Thermal history and subsidence of rifted continental margins, evidence from wells on the Nova Scotian and Labrador Shelves, *Canadian Journal of Earth Science*, V. 16, pp. 505-522.
- Keen, C.E., Keen, M.J., Barrett, D.L., and Heffler, D.E. (1975) Some aspects of the ocean-continental transition at the continental margin of eastern North America. *Geol. Surv. Canada*, paper 74-30 (2); pp. 189-197.
- Keen, C.E. and Hyndman, R.D. (1979) Geophysical review of the continental margins of eastern and western Canada. *Canadian Journal of Earth Science*, V. 16, pp. 712-747.
- Keen, C.E., Beaumont, C., and Boutilier, R (1981) Preliminary results from a thermo-mechanical model for the evolution of Atlantic-type continental margins. In press.
- King, E.R., Zeitz, I., and Alldredge, L.R. (1966) Magnetic data and the structure of the central Arctic region. *Bulletin of the Geological Society of America*, V. 77, pp. 619-646.
- King, L.H., Hyndman, R.D., and Keen, C.E. (1975) Geological development of the continental margin of Atlantic Canada. *Geosci. Canada*, V. 2, pp. 26-35.
- Kusznir, N.J. and Bott, M.H.P. (1977) Stress concentration in the upper lithosphere caused by underlying visco-elastic creep. *Tectonophysics*, V. 43, pp. 247-256.

- Leblanc, G. and Wetmiller, R.J. (1974) An evolution of seismological data available for the Yukon Territory and Mackenzie Valley. *Canadian Journal of Earth Sciences*, V. 11, pp. 1435-1454.
- Loken, O.H. (1971) Bathymetric observations along the east coast of Baffin Island: submarine moraines and iceberg distribution, Earth Science Symposium on Offshore Eastern Canada. *Geol. Surv. Can.*, Paper 71-23, pp. 509-519.
- McKenzie, D.P. (1977) The initiation of trenches: a finite amplitude instability, In: Talwani, M. and Pitman III, W., eds., *Island Arcs, Deep Sea Trenches and Back-Arc Basins*. Am. Geophys. Union, Maurice Ewing Ser. 1, pp. 57-61.
- McKenzie, D.P. (1978) Some remarks on the development of sedimentary basins. *Earth Planet. Sci. Lett.*, V. 40, pp. 25-32.
- Nadai, A.L. (1963) *Theory of Flow and Fracture of Solid*, McGraw-Hill Book Company, Inc., pp. 283-348 (Chapter 10).
- Neugebauer, H.J. and Spohn, T. (1978) Late stage development of mature Atlantic-type continental margins. *Tectonophysics*, V. 50, pp. 275-305.
- Parsons, B and Sclater, J.G. (1977) An analysis of the variation of ocean floor bathymetry and heat flow with age. *J. Geophys. Res.*, V. 82, pp. 803-827.
- Richter, F.M. and McKenzie, D.P. (1978) Simple plate models of mantle convection. *J. Geophys.*, V. 44, pp. 441-471.
- Ringwood, A.E. (1975) *Composition and Petrology of the Earth's Mantle*. McGraw-Hill, New York, pp. 298-303.
- Ringwood, A.E. and Green, D.H. (1966) An experimental investigation of the Gabbro-eclogite transformation and some geophysical implications. *Tectonophysics*, V. 3, pp. 383-427.
- Schlolz, C.H. (1977) Transform fault systems of California and New Zealand: similarities in their tectonic and seismic styles. *J. Geol. Soc. London*, V. 133, pp. 215-229.
- Service, K.G. and Douglas, A. (1973) Boundaries and fractures in finite element models of geological structures. *Geophys. J. R. Astr. Soc.*, V. 32, pp. 1-14.

- Sleep, N. (1969) Sensitivity of heat flow and gravity to the mechanism of sea-floor spreading. *J. Geophys. Res.*, V. 74, pp. 542-549.
- Sleep, N. (1971) Thermal effects of the formation of Atlantic continental break-up. *Geophys. J. R. Astr. Soc.*, V. 24, pp. 325-350.
- Sleep, N. and Snell, N.S. (1976) Thermal contraction and flexure of mid-continent and Atlantic marginal basins. *Geophys. J. R. Astr. Soc.*, V. 45, pp. 125-154.
- Smith, W.E.T., Whitham, K., and Piche, W.T. (1968) A microearthquake swarm in 1965 near Mould Bay, N.W.T.. *Bulletin of the Seismological Society of America*, V. 58, pp. 1991-2011.
- Sobczak, L.W. (1975) Gravity and deep structure of the continental margin of Banks Island and Mackenzie Delta. *Canadian Journal of Earth Sciences*, V. 12, pp. 373-384.
- Sobczak, L.W. (1977) The tectonic development of the southern Beaufort Sea and its relationship to the origin of the Arctic Ocean Basin, by Yorath and Norris (1975), discussion. *Bulletin of Canadian Petroleum Geology*, V. 25, pp. 698-703.
- Spohn, T. and Neugebauer, H.J. (1978) Metastable phase transition models and their bearing on the development of Atlantic-type geosynclines. *Tectonophysics*, V. 50, pp. 387-412.
- Stein, S., Sleep, N., Geller, R.J., Wang, S.-C., and Kroeger, G.C. (1979) Earthquakes along the passive margin of eastern Canada. *Geophys. Res. Letters*, V. 6, pp. 537-540.
- Stephenson, R. (1977) Stresses Developed at an Aseismic Continental Margin, M. Sc. Thesis, Department of Geology, Carleton University, Ottawa, Canada.
- Stevens, A.E. (1974) Seismicity of northern Canada. *Bulletin of Canadian Petroleum Geology*, V. 22, pp. 387-404.
- Sykes, L.R. (1978) Intraplate seismicity, reactivation of pre-existing zones of weakness, Alkaline Magmatism, and other tectonism postdating continental fragmentation. *Reviews of Geophys. and Space Phys.*, V. 16, No. 4, pp. 621-688.
- Sykes, L.R. and Sbar, M.L. (1973) Intraplate earthquakes, lithospheric stresses and the driving mechanism of plate tectonics. *Nature*, V. 245, pp. 298-302.

- Talwani, M. and Eldholm, O. (1972) Continental margin off Norway: a geophysical study. Bull: Geol. Soc. Am., V. 83, pp. 3575-3608.
- Turcotte, D.L. (1974) Are transform faults thermal contraction cracks? J. Geophys. Res., V. 79, pp. 2573-2577.
- Turcotte, D.L. and Oxburgh, E.R. (1976) Stress accumulation in the lithosphere. Tectonophysics, V. 35, pp. 183-199.
- Turcotte, D.L. and Ahern, J.L. (1977) On the thermal and subsidence history of sedimentary basins. J. Geophys. Res., V. 82, No. 26, pp. 3762-3766.
- Turcotte, D.L., Ahern, J.L., and Bird, J.H. (1977a) The state of stress at continental margins. Tectonophysics, V. 42, pp. 1-28.
- Turcotte, D.L., Haxby, W.F., and Ockendon, J.R. (1977b) Lithospheric instabilities, In: Talwani, M. and Pitman III, W., eds., Island Arcs, Deep-Sea Trenches and Back-Arc Basins. American Geophysical Union, Maurice Ewing Series, V. 1, pp. 63-69.
- Walcott, R.I. (1970a) Isostatic response to loading of the crust in Canada. Canadian Journal of Earth Sciences, V. 7, pp. 716-727.
- Walcott, R.I. (1970b) Flexural rigidity, thickness, and viscosity of the lithosphere. J. Geophys. Res., V. 75, pp. 3941-3954.
- Walcott, R.I. (1972) Gravity, flexure and the growth of sedimentary basins at a continental edge. Geol. Soc. Amer., V. 83, pp. 1845-1848.
- Watts, A.B. and Talwani, M. (1974) Gravity anomalies seaward of deep-sea trenches and their tectonic implications. Geophys. J. R. Astro. Soc., V. 36, pp. 57-90.
- Watts, A.B. (1978) An analysis of isostasy in the world's ocean: part 1 - Hawaiian-Emperor seamount chains. J. Geophys. Res., V. 83, pp. 5989-6004.
- Williams, H. (1979) Appalachian Orogen in Canada. Canadian Journal of Earth Sciences, V. 16, pp. 792-807.
- Wilson, J.T. (1965) A new class of faults and their bearing on continental drift. Nature, V. 207, pp. 343-347.

Worzel, J.L. and Shurbet, G.L. (1955) Gravity anomalies at continental margins. Proc. Nat. Acad. Sci., V. 41, pp. 963-983.

Zienkiewicz, O.C. (1971) The Finite Element Method in Engineering Science. McGraw-Hill, London, 521 p.

NO INITIAL STRESS CONDITION, NON-UN
THERMAL STRAIN

DISTRIBUTION OF STRESSES

| 20F |

NON-UNIFORM POISSON'S RATIO WITHOUT THE CONSIDERATION OF

22

OF / DE

OT

60F6

END

0	1	H	0	9	H	8	2

FIN