SUBSIDENCE ANALYSIS AND TECTONIC MODELING OF THE SVERDRUP BASIN

R.A. Stephenson and J. Boerstoel, Institute of Earth Sciences, Vrije Universiteit, De Boelelaan 1085, 1081 HV Amsterdam, Netherlands
A.F. Embry, Geological Survey of Canada, Institute of Sedimentary and Petroleum Geology, 3303-33 Street Northwest, Calgary, Alberta T2L 2A7, Canada
B.D. Ricketts, Geological Survey of Canada, Cordilleran Division, 100 West Pender Street, Vancouver, British Columbia V6B 1R8, Canada

ABSTRACT

The Sverdrup Basin is a pericratonic trough in the northern Canadian Arctic Archipelago containing more than 13 km of Carboniferous to Tertiary strata. Quantitative subsidence analysis (backstripping) and forward modeling of the stratigraphy at 12 locations within the Sverdrup Basin are presented here. Initial basin formation was a result of crustal extension ("stretching factor" $\beta_1 < 2$) in the Permo-Carboniferous. An active extensional regime ("stretching factor" $\beta_2 \leq 1.3$) appears to also have influenced basin evolution from the Oxfordian until the end of the Albian. The results demonstrate thereafter a transition from gradual thermal subsidence to accelerated subsidence in the Paleocene. The calculated rates of the crustal tectonic subsidence in the eastern Sverdrup Basin during this phase are as great as those seen during the Permo-Carboniferous syn-rift phase of basin development. The accelerated subsidence occurs in an increasingly compressional setting related to the eventual onset of the Eurekan.

INTRODUCTION

Initial (Carboniferous-Early Permian) deposition in the Sverdrup Basin (Fig. 1) took place in a restricted marine environment (e.g., Davies and Nassichuk, 1975) with contemporaneous extensional deformations (Meneley et al., 1975) related to the development of a major continental rift zone (cf., Balkwill and Fox, 1982). Syn-rift strata comprise a circumferential belt of massive carbonate rocks surrounding an axial basin of mainly deep-water shale overlying several hundreds of meters of evaporitic rock (Nassichuk and Davies, 1980). Postrift crustal subsidence, due to thermal contraction and sediment loading, continued essentially uninterrupted until the Early Cretaceous. During this tectonic phase, a large, deep (>2000 m) pericontinental basin in existence at the close of the initial rifting phase (Early Permian) was filled with sediment by the Early Jurassic. The present-day (compacted) thickness of the postrift Permian to lowest Lower Cretaceous strata is approximately 7 km at the basin center.

The style of evolution of the Sverdrup Basin in the Cretaceous was affected by the development of the adjacent Amerasia Basin of the present-day Arctic Ocean. Widespread igneous activity, related to a mantle hot spot centered in the adjacent Arctic Ocean basin (Ricketts et al., 1985; Embry and Osadetz, 1988), characterizes the basin during this time. In the northeast, basaltic flows are interlayered with upper Lower and lowermost Upper Cretaceous strata. Diabase dikes and sills of the same age (Jackson and Halls, 1988) are more widespread.

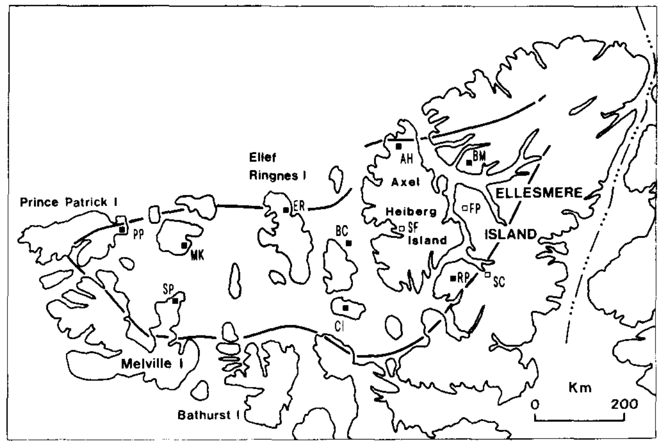

Fig.1. The Sverdrup Basin, Arctic Canada (with approximate limits indicated by the intermittent solid line), showing locations of stratigraphic sections used in the tectonic subsidence analysis. Key: AH, northern Axel Heiberg Island; BC, Basin Center; BM, Blue Mountains, Ellesmere Island; CI, Cornwall Island; ER, northern Ellef Ringnes Island; FP, Fosheim Peninsula (Late Cretaceous, Tertiary section merged with BM); MK, Mackenzie King Island; PP, eastern Prince Patrick Island; RP, Raanes Peninsula; SC, Strathcona Fiord (Late Cretaceous, Tertiary section merged with RP); SF, Strand Fiord (Late Cretaceous, Tertiary section merged with BC); SP, Sabine Peninsula, Melville Island.

The final tectonic phase in the development of the Sverdrup Basin involved the uplift and deformation of the northeastern portion of the Arctic Islands by the early Tertiary Eurekan Orogeny, as Greenland impinged upon Ellesmere Island (Balkwill, 1978; Miall, 1984; de Paor et al., 1989). Stratigraphic evidence indicates that Eurekan tectonism may have been active in the late Paleocene but that it climaxed during the Eocene with the clear development of intramontane sedimentary basins (Ricketts, 1987, 1993) associated with large-scale compressional structures (Stephenson et al., 1990).

In this paper, the history of vertical tectonic motions of the crust in the Sverdrup Basin is quantified by a backstripping- and tectonic-subsidence-modeling analysis of a

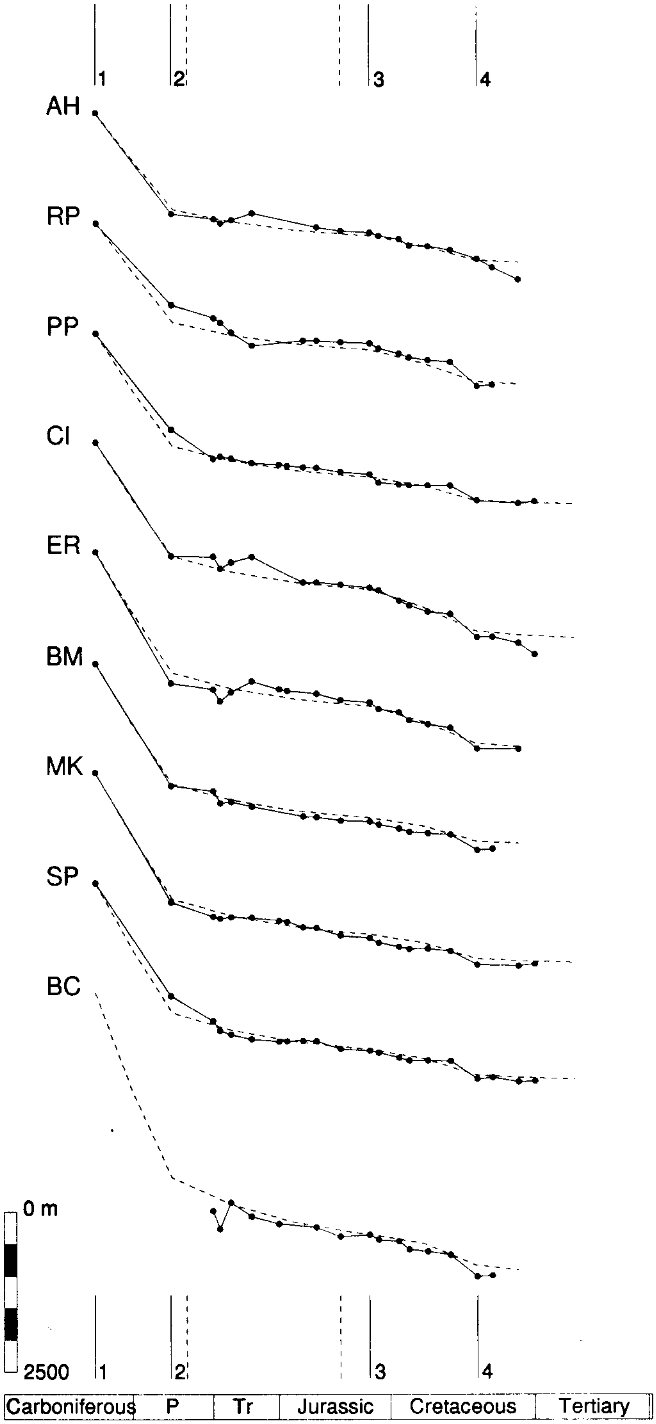

Fig.2. Observed (solid lines with data points) and modeled (dashed lines) tectonic subsidence curves for nine locations indicated in Fig.1 (filled symbols). The timespan covered in Carboniferous to Cretaceous. Vertical solid lines delimit modeled tectonic phases: 1, extension; 2, thermal subsidence; 3, renewed extension (vertical dashed lines indicate limits based on geological evidence, as noted in the text); 4, thermal subsidence. Relevant model parameters are: crustal thickness of 40 km; lithospheric thickness of 125 km; asthenosphere temperature of 1333°C; coefficient of thermal expansion of 3.4 x 10^{-5}°C^{-1}; thermal diffusivity of 7.8 x $10^{-7}m^2s^{-1}$.

number of stratigraphic sections throughout the basin. The results elucidate the timing and character of thermo-extensional and other tectonic events in the Sverdrup Basin and thus help to establish constraints on the plate tectonic evolution of the surrounding areas.

TECTONIC SUBSIDENCE CURVES

Stratigraphy records vertical motions of the crust during the development of sedimentary basins. The back-stripping method (or geohistory analysis) separates isostatic effects of sediment and water loading from those of tectonic subsidence (Steckler and Watts, 1978). Mathematical procedures to calculate tectonic subsidence through time from stratigraphic columns have been discussed, for instance, by Steckler and Watts (1978), Sclater and Christie (1980), and Bond and Kominz (1984). The necessary input consists of stratigraphic time-depth information (thicknesses and ages, lithology, compaction/porosity, and depositional waterdepths).

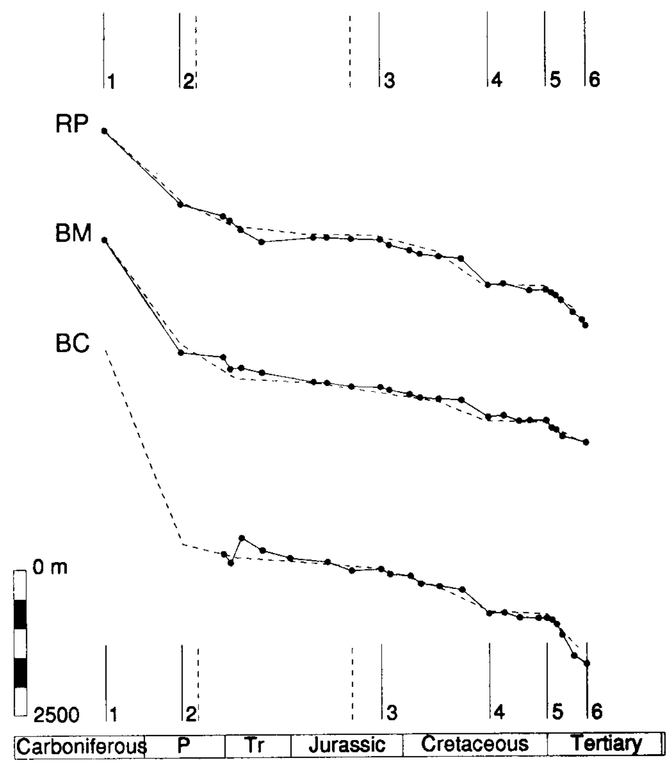

Fig.3. Observed (solid lines with data points) and modeled (dashed lines) tectonic subsidence curves for three locations indicated in Fig.1. The timespan covered is Carboniferous to Tertiary. These curves are composites; data for Cenomanian and younger strata were taken from nearby exposures (locations SC, FP, and SF, Fig.1). Vertical lines delimit tectonic phases as in Fig.2 appended by: 5, onset of horizontal compression (increasing to 600 MPa at 43 Ma); 6, culmination of the Eurekan Orogeny and demise of the Sverdrup Basin. Base of the elastic layer in the flexural model coincides with the 250°C isotherm.

Fig.2 shows late Carboniferous-Late Cretaceous backstripped-tectonic-subsidence curves for nine locations throughout the Sverdrup Basin (located in Fig.1). The stratigraphic input data were constructed from measurements of exposed strata (more so in the eastern part of the basin affected by Tertiary deformation and uplift) and from seismic stratigraphy in conjunction with subsurface stratigraphy from exploration wells. Published sources include Tozer (1963), Tozer and Thorsteinsson (1964), Balkwill and Roy (1977), and Balkwill (1983). Measured horizons in the basin center (BC) do not exceed Triassic in age. Detailed stratigraphic sections of Cenomanian and younger strata from three different locations in the eastern part of the Sverdrup Basin (cf., Fig.1) derived from Ricketts (1991, 1993) provide information on tectonic subsidence into the Tertiary. These data have been merged with those of the respective closest older section to construct the composite tectonic subsidence curves shown in Fig.3. Also included were the effects of overlying strata now eroded with thicknesses estimated from compaction and vitrinite-reflectance studies (e.g., Allen, 1986; Brooks et al., 1992). These estimates are necessarily imprecise and are intended to illustrate the general pattern of tectonic subsidence through to the end of sedimentation in the Eocene.

In the backstripping calculations, the chronostratigraphic timescale of Harland et al. (1989) has been used. Porosity-depth corrections were calculated on the basis of the observed average lithologies using standard exponential relations and material parameters (cf., Sclater and Christie, 1980). Different (de)lithification processes can be adopted (e.g., Bond and Kominz, 1984) with the resulting calculated minima and maxima providing an estimate of vertical uncertainty in the tectonic-subsidence estimates. These are illustrated for one location (MK) in Fig.4; this figure also shows the basement subsidence through time, corresponding to the calculated tectonic subsidence. No specific paleobathymetric data are available; estimates of paleowaterdepth have been taken from the paleogeographical modeling of Stephenson et al. (1987). These are consistent with a deep (>2000 m) Early Permian basin that becomes filled with sediment by the Early Jurassic (cf., waterdepth plot in Fig.4). Sea-level changes were not included in the backstripping. A local (Airy) isostatic mechanism with a mantle-compensation density of 3,300 kgm^{-3} was adopted.

SUBSIDENCE MODELS

The late Paleozoic-Cretaceous tectonic subsidence of the Sverdrup Basin (Fig.2) displays a stepwise pattern with periods of rapid tectonic subsidence followed by phases of slower subsidence. The former correspond to times of active tectonic influence, viz., the initial Permo-Carboniferous "syn-rift" extensional phase and a subsequent episode of extensional rejuvenation in the Late Jurassic and Cretaceous. These tectonic events are followed by phases of slower, exponentially decaying, thermal "postrift" subsidence during which the extensional mechanism is inactive. Fig.3 shows that the second of these was terminated at the beginning of the Paleocene by an abrupt and marked increase in the rate of tectonic subsidence.

Thermo-Extensional Phases

Two different theoretical syn-rift and postrift thermal subsidence curves are plotted in Fig.4 (dashed lines) for comparison with the observed tectonic subsidence. They are based on the lithospheric-stretching model of McKenzie (1978), whereby the lithosphere is heated during rifting and subsequently cools through vertical heat conduction. The implementation of the model allows incorporation of finite (e.g., Jarvis and McKenzie, 1980) and multiple stretching phases (Kooi et al., 1989). The

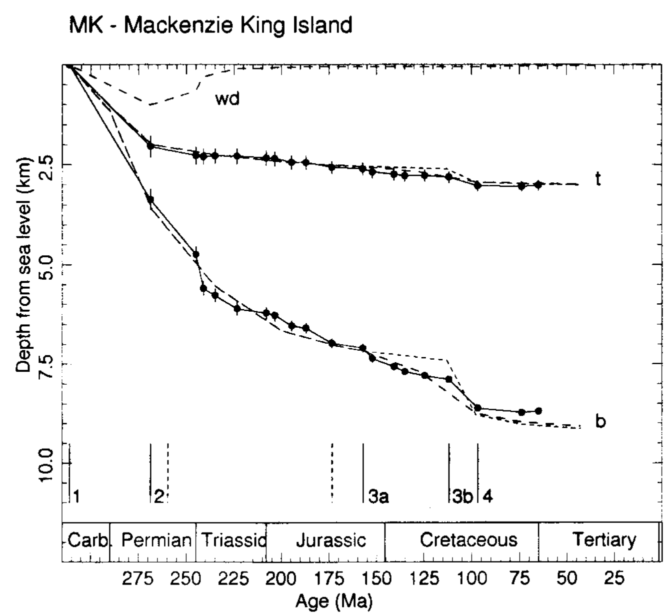

Fig.4. Observed (solid lines with data points) and modeled (dashed lines) basement (b) and tectonic (t) subsidence curves for location MK (Mackenzie King Island) in Fig.1. The numbered vertical bars refer to the temporal framework of the model: 1, active lithospheric extension in the period 311.1-268.8 Ma followed by 2, thermal subsidence interrupted by renewed extension either at 3a, 157 Ma (longer dashed line), or at 3b, 112 Ma (shorter dashed line), in both cases ceasing activity at 4, 97 Ma when passive thermal subsidence is re-established. (Vertical dashed lines indicate these limits based on geological evidence, as noted in the text.) The "error" bars on the data points refer to decompaction extrema. Uncertainties of similar magnitude also should be associated with the mean curves displayed in Figs.2 and 3. The water-depth profile used in the backstripping calculations is plotted as a dashed line (wd).

model assumes Airy isostatic compensation and disregards the potential effects of lateral heat flow. One of the most sensitive modeling parameters is the initial crustal thickness; a value of 40 km was adopted in accordance with seismic refraction and gravity data (Forsyth et al., 1979; Sobczak and Overton, 1984).

The modeled initial "syn-rift" phase of subsidence ends at the end of the Sakmarian (~269 Ma), in general agreement with the geological evidence (e.g., Meneley et al., 1975). Beauchamp et al. (1989) indicate that rifting may have continued until the end of the Artinskian (~260 Ma). The temporal limits of the Mesozoic rejuvenation event, especially its commencement, are more difficult to place given the uncertainty in the calculated tectonic subsidence. The two sets of calculated curves in Fig.4 correspond to models with this event occurring from 157-97 Ma (Oxfordian-Albian) and 112 to 97 Ma (Albian). The end of this phase coincides with a widespread (Albian-Cenomanian) unconformity that has been interpreted to coincide with either the onset (Embry and Dixon, 1990) or cessation (Embry and Dixon, this volume) of seafloor spreading in the Amerasia Basin. Strata underlying the unconformity display normal faulting. In the Banks Island area, southeast of the present study area along the continental margin, unpublished seismic data indicate that faults were active since the Late Jurassic (Embry and Dixon, 1990). The subsidence model with active extensional tectonics beginning at 157 Ma, therefore, has been used in modeling the other subsidence data, although the backstripped data clearly do not provide unequivocal evidence.

The dashed lines in Fig.2 represent the best-fitting, two-phase thermo-extensional model subsidence curves for the Sverdrup Basin backstripped curves. The "stretching" factors used to calculate these curves, β_1 and β_2, are shown in Fig.5. The β_1 values are slightly lower than those of the local isostatic model of Stephenson et al. (1987), easily explained by the different modeling methodology employed in the latter, in which thermal subsidence is less efficient. β represents a kind of effective lithospheric temperature-perturbation parameter that should not be literally interpreted in terms of crustal or lithospheric thinning. Relative changes in β throughout a basin may be of tectonic interest, however. For instance, total β_T ($=\beta_1 \bullet \beta_2$) compared to β_1 or to β_1/β_2 indicates that the second thermo-extensional event was relatively more important in the eastern part of the basin, coinciding with the distribution of Cretaceous igneous activity. This suggests that the development of the Sverdrup Basin during this time perhaps reflects not only incipient rifting of the polar continental margin but also more localized effects such as those due to the postulated adjacent offshore hot spot (Embry and Osadetz, 1988).

Paleocene-Eocene Compressional Phase

The phase of rapid subsidence evident in the eastern Sverdrup Basin in the early Tertiary (Fig.3) occurs in a setting of gradually increasing compression related to seafloor spreading in the Labrador Sea and Baffin Bay and the consequent impingement of Greenland against the islands of eastern Arctic Canada (e.g., Scotese et al., 1988). The effects of the changing tectonic regime can be recognized in the sequence stratigraphy of the Late Cretaceous-Paleogene succession (Ricketts, 1993). The geological record indicates the short-lived development of intramontane basins just prior to the overall cessation of regional basin formation at the middle Eocene culmination of the Eurekan Orogeny. This has been modeled as a whole-lithosphere compressional failure (Stephenson et al., 1990).

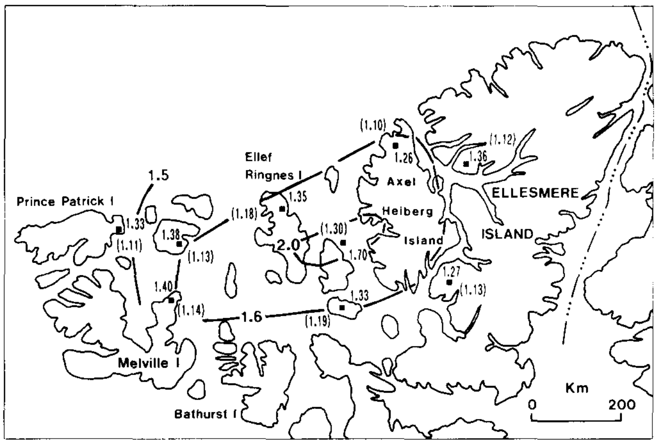

Fig.5. Data locations with model values of β_1 (for the first syn-rift period, 311-269 Ma) and β_2 (for the second syn-rift period, 157-97 Ma, in parentheses) and β_T (product of β_1 and β_2) contoured.

The model curves in Fig.3 through the end of the Cretaceous are based on the same extensional tectonic framework discussed above, although adopting a flexural isostatic-compensation mechanism. The flexural assumption allows the incorporation of the isostatic effects of changing tectonic stresses (e.g., Cloetingh et al., 1985). As such, the Paleogene subsidence of the Sverdrup Basin has been modeled analogously to the Plio-Pleistocene acceleration of subsidence observed in the North Sea (e.g., Kooi and Cloetingh, 1989; Kooi et al., 1991). In the North Sea, this has been referred to as "late-stage compression," although the Sverdrup data indicate that it results not from some process intrinsic to the preceding thermo-mechanical subsidence but rather to significant changes in the plate tectonic setting. It is important to note that there is no geological evidence of a definitive overthrust/foredeep association in the Paleogene strata of the eastern Sverdrup Basin (Ricketts, 1993). Thus, subsidence during this phase was not a response to tectonically emplaced vertical loads as in a typical foreland basin (e.g., Beaumont, 1981).

In the model, the applied horizontal compressional stress increased from nil at the beginning of the Paleocene to 600 MPa (6 kbar) at the culmination of the Eurekan Orogeny in the middle Eocene (43 Ma). This is very probably an unrealistically high magnitude of horizontal stress in a lithospheric plate and, like ß, should be viewed as a kind of model-dependent integrated-stress/rheology parameter. Similar values have been calculated from models of intraplate stress due to plate boundary forces (Cloetingh and Wortel, 1986) and of large-scale intraplate compressional deformation (e.g., Lambeck, 1983; McAdoo and Sandwell, 1985).

SUMMARY AND CONCLUSIONS

There are three tectonic "stages" in the evolution of the Sverdrup Basin identifiable in backstripped regional subsidence curves: the first two are extensional and the third is compressional.

The first extensional stage is modeled as a Moscovian-Sakmarian rifting phase (311-269 Ma) followed by generally subdued thermal subsidence. Poor resolution of depositional waterdepths during the rifting phase degrades the reliability of calculated stretching parameters. Maximum β_1 for a local isostatic model is less than 2.

The pattern of post-Paleozoic-rift thermal subsidence is interrupted by renewed extensional tectonism in the Late Jurassic. The tectonic subsidence data can be satisfactorily modeled by a phase of continuous stretching beginning in the Oxfordian and ending with the Albian (157-97 Ma), giving an indication of the geological timespan during which active extensional tectonics affected the Canadian Arctic in the Mesozoic. There is no evidence for significant diachroneity of the Mesozoic extensional event within the Sverdrup Basin, although it is not well constrained. In terms of the overall development of the Sverdrup Basin, this event is relatively more important in the eastern part of the basin affected by Cretaceous igneous activity.

Cenomanian-Maastrichtian basin development continued in the style that characterized much of the earlier development of the Sverdrup Basin, being controlled by passive thermal subsidence (induced by both earlier thermo-extensional events).

The rate of tectonic subsidence in the Sverdrup Basin from the Paleocene until the culmination of the Eurekan Orogeny in the mid-Eocene (65-43 Ma) equaled the rate inferred during Permo-Carboniferous rifting. The Tertiary acceleration of subsidence can be modeled by the compressional amplification of thermal subsidence in the developing Eurekan Orogen, which ultimately destroyed the contiguous Sverdrup Basin, leaving in its place several small, intramontane sedimentary basins.

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