PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A
MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

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Phil. Trans. R. Soc. Lond. A 1981 **301**, 293-303
doi: 10.1098/rsta.1981.0112

Phil. Trans. R. Soc. Lond. A **301**, 293–303 (1981)
Printed in Great Britain

Tonalites in crustal evolution

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Tonalites, including trondhjemite as a variety, played three roles through geological time in the generation of Earth's crust. Before about 2.9 Ga ago they were produced largely by simple partial melting of metabasalt to give the dominant part of Archaean grey gneiss terranes. These terranes are notably bimodal; andesitic rocks are rare. Tonalites played a crucial role in the generation of this protocontinental and oldest crust 3.7–2.9 Ga ago in that they were the only low-density, high-SiO_2 rocks produced directly from basaltic crust.

In the enormous event giving the greenstone-granite terranes, mostly 2.8–2.6 Ga ago, tonalites formed in lesser but still important proportions by partial melting of metabasalt in the lower regions of down-buckled greenstone belts and by remobilization of older grey gneisses.

Tectonism in the Archaean (3.9–2.5 Ga ago) perhaps was controlled by small-cell convection (McKenzie & Weiss 1975). Little or no ophiolite or eclogite formed, and only minor andesite. Plate tectonics of modern type (involving large, rigid plates) commenced in the early Proterozoic. Uniformitarianism thus goes back one-half of the age of the earth.

Tonalites compose about 5–10% of crust generated in Proterozoic and Phanerozoic time at convergent oceanic–continental margins. They occur here as minor to prominent members of the compositionally continuous continental-margin batholiths. A simple model of generation of these batholiths is offered: mantle-derived mafic magma pools in the lower crust above a subduction zone reacts with and incorporates wall-rock components (Bowen 1922), and breaches its roof rocks as an initial diapir. This mantle magma also develops a gradient of partial melting in its wall rocks. This wall-rock melt accretes in the collapsed chamber and moves up the conduit broached by the initial diapir, the higher, less siliceous fractions of melting first, the lower, more siliceous (and further removed) fractions of melting last. The process gives in the optimum case a mafic-to-siliceous sequence of diorite or quartz diorite through tonalite or quartz monzodiorite to granodiorite and granite. The model implies that great masses of cumulate phases and refractory wall rock form the roots of continental-margin batholiths, and that migmatites overlie that residuum and underlie the batholiths.

1. Introduction

Tonalites (as defined by the I.U.G.S. system (Streckeisen 1976; figure 4)) are a major rock type of the granitic (s.l.) group. Tonalites and quartz diorites are the least potassic of common granitic rocks and show the widest range of SiO_2 contents (*ca.* 65–75%). In much of the older geological literature tonalites were not treated rigorously; foliated tonalites, especially, commonly were termed 'granitic gneiss'. Recent research, however, has clarified the geological occurrence and many characteristics of tonalites. This paper considers the role of tonalites in crustal evolution. Its major conclusion is that tonalites have played distinctive but rather different roles in each of the three major tectonomagmatic intervals of Earth's recorded history.

2. Three stages of crustal evolution

In considering tonalites in the last 3800 Ma of Earth history our approach is first to postulate that several modes of mantle and crustal tectonism and magmatism have operated during this period, and then to consider the role of tonalites and the associated granitic rocks in each mode. Three stages of crustal evolution are examined and summarized in table 1. The protocontinental stage (3.9 to 2.8 Ga ago) was characterized by production of sialic crust dominated by tonalite and trondhjemite, and containing both minor granodiorite, granite and other intrusives, and ultramafic to tholeiitic basalts; the greenstone belt stage (typically 2.8 to 2.6 Ga ago but as

Table 1. Tonalites in crustal evolution

stage and duration	protocontinental stage	greenstone belt stage	plate tectonic stage
	3800–2900 Ma, present N Hemisphere 3800?–3400 Ma, present S Hemisphere	2800–2600 Ma, present N Hemisphere 3500–2600 Ma, present S Hemisphere	*ca.* 2200 Ma – present
mode of tectonism	small-cell mantle convection; small crustal plates; no eclogite, no ophiolite	small-cell convection(?), small crustal plates; down-folding of volcanic piles, no ophiolite	plate tectonics (large rigid plates) of modern type; eclogite necessary
associated mantle-derived volcanism	komatiitic to tholeiitic basaltic; small greenstone belts formed?	ultramafic to komatiitic to tholeiitic basaltic; commonly bimodal tholeiitic–dacitic; locally basaltic to andesitic (calc-alkaline)	basaltic–andesitic–dacitic–rhyolitic, tholeiitic, calc-alkaline, alkaline
coeval andesitic magmatism?	traces	locally found in stratigraphically high calc-alkaline suites	abundant
generation of siliceous magmas			
(1) source rocks, composition	basaltic; secondarily tonalite and trondhjemite	basalt to dacite; greywacke and other sediments; pre-existing sialic crust	accreted oceanic rocks; continental crustal rocks; continental marginal sediments; mantle magmas
(2) source rocks, metamorphic facies	amphibolite to granulite	greenstone to amphibolite; granulite	greenstone to amphibolite; locally blueschist to eclogite
(3) magma-forming process	batch partial melting; local c.–l. (crystal–liquid) fractionation	batch partial melting; local c.–l. fractionation	mantle-derived magma and lower to intermediate crust interact; local c.–l. fractionation
(4) rocks formed	tonalite–trondhjemite suite; secondary granite–granodiorite	mixed tonalite, trondhjemite, granodiorite, granite, other rocks	± gabbro ± diorite–quartz diorite–tonalite; or quartz monzodiorite–granodiorite–granite; locally trondhjemite
preservation	poor	partly good, partly poor (as in later mobile belts)	good to excellent
role of tonalitic rocks	generation of protocontinents; first sialic rocks formed from basalts	major involvement in generation of 20–60 % of present continents	forms *ca.* 10–55 % of continental-margin batholiths

early as 3.5 Ga ago in parts of Africa and Australia) commenced with ultramafic to tholeiitic basalts, dacites and other extrusives in elongate basins, followed by down-buckling of the basins and up-welling of tonalite, trondhjemite, granodiorite, granite and minor syenite; and the plate tectonic stage (about 2.2 Ga ago to the present day) involved large, rigid plates; magmatic arcs formed above subduction zones at convergent oceanic–continental plate margins and produced suites of ± gabbro ± diorite–quartz diorite–tonalite (or quartz monzodiorite)–granodiorite–granite, or separate members thereof, by interaction of mantle-derived mafic magma and crustal rocks.

3. The protocontinental stage

The greater part of Earth's oldest (older than 2.9 Ga) rocks are the so-called grey gneiss terranes, which consist largely of intensely deformed plagioclase–quartz gneisses of tonalitic to trondhjemitic composition. They include widely varying proportions of inclusions or dykes of amphibolite or granulite of ultramafic to tholeiitic composition. Intrusives of granitic gneisses commonly are found in grey gneiss terranes; many are of or close to minimum-melt compositions and probably were derived from nearby migmatitic tonalitic–trondhjemitic gneisses (e.g. as in the Ancient Gneiss Complex of Swaziland (Hunter *et al.* 1978). Metasedimentary rocks, predominantly quartzite, metachert and ironstone, occur as thin, scattered layers in most grey gneiss terranes.

In spite of early assertions that the grey gneisses are 'granitized' sedimentary rocks, most of these rocks show elemental abundances like those of less metamorphosed or deformed igneous rocks and most modern workers believe they largely are metaigneous. Many probably formed from plutonic protoliths, whereas some that show closely interlayered siliceous and mafic rocks may be either metavolcanic or structurally attenuated siliceous intrusives and mafic inclusions or dykes.

The dark portions of the grey gneisses are compositionally like the tholeiites and komatiites that are common in late Archaean volcanic greenstone belts. However, the extent of mafic–ultramafic volcanism in the protocontinental stage is poorly known. The metaplutonic Amitsoq gneisses of Greenland (McGregor 1979), for example, which are 3.6 Ga old, contain scattered metabasaltic and metasedimentary enclaves, termed the Akilia association by McGregor & Mason (1977). Lithologies of the Akilia suite range from ultramafic to Fe-rich tholeiitic basalts, laminated quartzite or metachert, ironstone and mica schists. McGregor & Mason note that these rocks may represent part of an older greenstone belt. In contrast, the older parts of the Ancient Gneiss Complex of Swaziland (Hunter *et al.* 1978) consist of closely interlayered komatiitic to tholeiitic metabasalts, quartzo-feldspathic gneisses of trondhjemitic composition and minor metasedimentary rocks. These rocks may represent an original volcanic succession, now deformed and metamorphosed to upper amphibolite facies, of bimodal basaltic–dacitic type (Barker & Peterman 1974). Thus in terranes dominated by either plutonic or volcanic rocks a record of early mafic volcanism is preserved. As various workers have suggested, mafic rocks were the dominant constituent of an early crust before the protocontinental stage. Much of this very early mafic crust was destroyed or severely modified during the protocontinental stage, when it acted as the principal source of tonalitic–trondhjemitic and dacitic magmas, as discussed below.

It is now widely accepted that trondhjemitic and tonalitic magmas are derived from basaltic sources, either as the product of partial melting of amphibolite or eclogite, or through fractional crystallization of basaltic liquid. This conclusion is based on the common association of tonalite and trondhjemite with basalt, amphibolite or gabbro, the success of trace element models (Arth & Hanson 1975), the primitive initial Sr isotopic composition found in trondhjemites of all geologic eras (Peterman 1979), the oxygen isotope ratios of trondhjemites that are similar to those of basalts but lower than those of most other granitic rocks (Barker *et al.* 1976), and laboratory experiments in which amphibolite partially melted to trondhjemitic liquid (Helz 1976). Students of grey gneiss terranes tend to prefer partial-melting mechanisms to differentiation of mafic magma because of the scarcity of intermediate magmas.

Our view of the environment of generation of early, largely sialic crust of protocontinental type is summarized in table 1. From the many dissimilarities of these rocks to those formed in plate-tectonic environments of modern type we do not believe that the Archaean tectonic regime was dominated by strong, overriding plates. Ophiolites, for example, are rare or non-existent in the Archaean. Rather, we agree with McKenzie & Weiss (1975) that small-cell convection in the mantle easily fractured and fragmented basaltic or sialic crust. McGregor's model (1979) for generation of the Nûk gneisses, which are 2.8 Ga old, may generally apply to Archaean grey gneiss terranes. It postulates convection of small cells of perhaps 100–200 km breadth, basaltic volcanism at the centre of each cell over rising mantle material, sweeping of this mafic crust and of any older sialic crust into areas of sinking between the mantle cells and thickening of as much as several tens of kilometres in these areas, and attendant rise of geotherms in the lower regions of this thickened crust. Metamorphism to amphibolite and granulite facies and magma production by partial melting also occur. In McGregor's model the thickened doubly facing wedges above the down-flowing mantle currents would grade from volcanic rocks at the surface, down through recumbently folded and thrust-faulted volcanic and plutonic rocks, to melt-depleted granulites just above the mantle. Partial melting of 10–20 % of meta-basalt in this column at or near the amphibolite–granulite transition produced mostly tonalitic and trondhjemitic liquids (Barker & Arth 1976), the tonalite–trondhjemite suite. These magmas then rose as diapirs.

Several aspects of this small-cell model deserve comment.

(i) This type of tectonomagmatic event did not occur continuously over long periods of time, but was sporadic. A spacing of major events, producing either mafic–ultramafic volcanic rocks and/or tonalitic–trondhjemitic gneisses 200 to 700 Ma apart apparently is typical of grey gneiss terranes.

(ii) Sinking of basaltic crust in the down-flowing parts of the cells remains a question. Green (1975) has pointed out that steep Archaean geothermal gradients probably would preclude formation of eclogite at shallow depth (< 80 km), and that granulite-facies basaltic crust that formed would not be sufficiently dense to sink into the mantle and would accumulate as crust (McGregor 1979). However, geothermal gradients probably varied geographically in Archaean time as they do today, and the possibility remains that basaltic crust rapidly carried down could pass through an eclogite stability field. Secondly, volcanic piles of ultramafic or peridotitic komatiites at granulite-facies mineralogy may be sufficiently dense to sink into the mantle.

(iii) The scarcity of andesite (SiO_2, 55–63 %) or its plutonic equivalents characterizes the protocontinental stage (Barker & Peterman 1974). Most workers in Cainozoic island arcs and continental margins suggest a strong to dominant role of the mantle wedge overlying the subducted oceanic plate in the production of andesites. Thus, vertical subduction involving no mantle wedge and producing little andesitic magma may have characterized the Archaean.

Preservation of most Archaean grey gneiss terranes is poor. They occur as remnants dispersed among younger granitic rocks and near late Archaean greenstone belts. Their destruction is summarized by the flow diagram (figure 1). The role of tonalitic–trondhjemitic magmatism, however, was of crucial importance in generating protocontinental crust from basaltic material and in providing the first abundant, low-density crustal rocks.

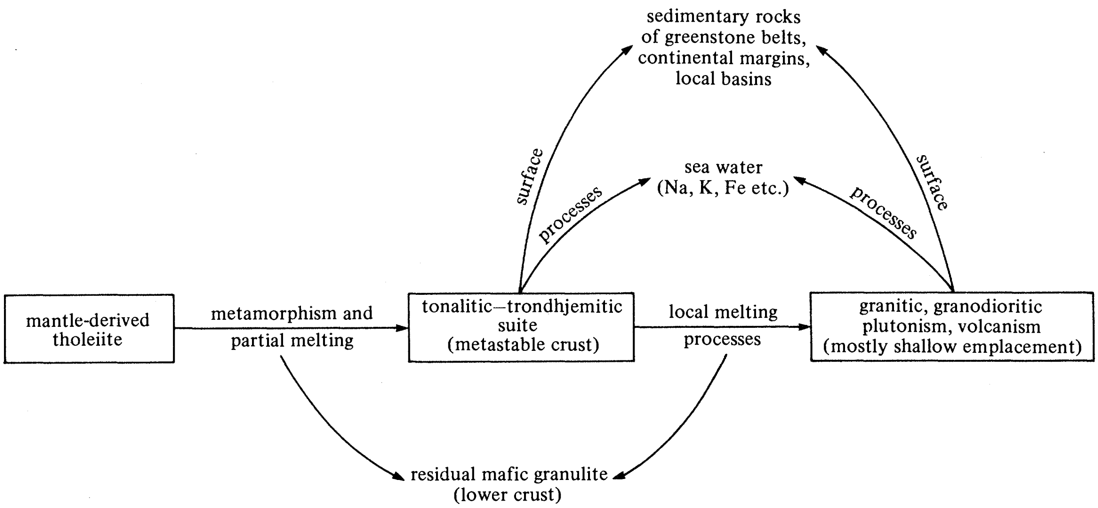

FIGURE 1. Flow diagram to illustrate major processes during the protocontinental stage.

4. THE GREENSTONE BELT STAGE

Formation of the greenstone-granite terranes in the late Archaean (but locally commencing as early as 3.5 Ga in the present southern hemisphere) probably contributed more to the development of Earth's continental crust than did any other event. About 30 % of present-day North America is composed of greenstone-granite terranes, largely well preserved but partly reconstituted in mobile belts by tectonic and metamorphic processes. Tonalites compose at least 10 %, and perhaps 15 to 25 %, of greenstone-granite terranes, but reliable determinations await future mapping.

Individual greenstone belts typically consist of a lower, largely mafic volcanic pile and an upper, sedimentary or sedimentary–volcanic sequence, of 5 to 17 km overall thickness; they are complexly folded, giving arcuate, Y-shaped, or more complexly bifurcating outlines in plan, features commonly interpreted as buckling due to vertical instability. Intrusives, commonly ovoid in plan, of tonalite, trondhjemite, granodiorite, granite and minor syenite are disposed along the margins of the volcanic–sedimentary belts and locally are found in the interiors. The belts mostly are at amphibolite facies along their margins (i.e. at the interfaces with the mixed plutonic rocks) and at greenschist facies in their interiors. The lower volcanic sequence usually commenced with pillowed peridotitic to basaltic komatiites and tholeiites, and ranged upward to tholeiites and dacites, giving an essentially bimodal or trimodal suite. A few belts commence with ultramafic to tholeiitic lavas and give way to calc-alkaline andesite and dacite (see, for example: Goodwin 1977; Hawkesworth & O'Nions 1977). The sedimentary rocks include tuff, agglomerate, volcanogenic greywacke, ironstone, jaspilite, chert, pelite, quartzite and conglomerate.

Hypotheses of origin of greenstone belts (see Windley 1976) include deposition as island arcs on mafic crust, in basins formed as rifts in sialic crust, in basins underlain by sialic crust, and in marginal basins. Our favoured model (table 1) follows that of McKenzie & Weiss (1975) and Sun & Hanson (1975) and involves breaking or rifting of protocontinental sialic to basaltic crust by the action of relatively small convection cells. The general lack of geological asymmetry or sequential temporal relations across or along greenstone-granite terranes precludes, we believe, modern analogues of plate tectonic type.

Assessing the importance of tonalites in greenstone-granite terranes is difficult because plutonic rocks of these regions have received less study than have the volcanic rocks. Recent work (see, for example, Ermanovics *et al.* 1979; Smith & Williams 1978) is providing much new information. We tentatively distinguish two types of tonalite in greenstone–granite terranes.

(i) Well foliated, commonly migmatitic and bearing scattered enclaves of amphibolite, and streaked with pegmatite; found locally along margins of down-buckled greenstone sequences, but mostly in the extensive batholithic or gneiss belts that lie between the volcanic belts. For example, the Berens batholithic belt of the Superior Province, which measures 250 by 600 km, consists of 40 to 50% tonalite of this type and a remainder of younger granitic intrusives (Ermanovics *et al.* 1979). Large portions of the English River gneiss belt also are migmatitic tonalite and trondhjemite (Breaks & Bond 1977; Ermanovics *et al.* 1979), and the remainder metasedimentary gneisses and other rocks. This type apparently is remobilized older tonalite.

(ii) Relatively homogeneous rock intruded during folding of a greenstone belt that is compositionally equivalent to stratigraphically high siliceous dacites of that greenstone belt. The 2.65 Ga old extrusive Northern Light Gneiss and the intrusive Saganaga Tonalite of northeastern Minnesota are an excellent example (Arth & Hanson 1975). Partial melting of metabasalt, which petrographically is amphibolite, at the base of the thick volcanic pile generates both dacite and tonalite (Barker & Arth 1976).

5. The plate tectonic stage

(*a*) *Commencement of plate tectonics in early Proterozoic time*

The operation in the Precambrian of plate tectonics of modern type, involving large, rigid plates forming at oceanic spreading centres and being partially transformed to eclogite during subduction, has been a major debate (see, for example, Dewey & Spall 1975). Most agree that this type of tectonics occurred in the late Precambrian in the North Atlantic region; a few (see, for example, Burke & Kidd 1979) contend that it operated more vigorously in the Archaean than now. We believe that modern-type plate tectonics commenced in early Proterozoic time and we agree with the suggestion of Hietanen (1975) that the 2.0–1.8 Ga old Proterozoic Svecokarelian orogeny of Finland involved convergence of an oceanic plate on the southwest and an Archaean continental plate on the northeast. Plutonic rocks that range from gabbro to granite intrude the Svecokarelian volcanic–sedimentary terrane (Simonen 1960); tonalite composes 15 to 25% of these.

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(b) *Continental-margin batholiths*

Continental growth in Proterozoic and Phanerozoic time has been largely at the margins of cratons by tectonic and magmatic processes that accompany convergence of oceanic and continental plates. Subsidiary processes, such as rifting extending inward from a continental margin and the attendant potassic magmatism, contribute to continental growth only in a minor way. Rocks added to continental margins in convergent processes include accreted oceanic rocks, especially island arcs, ophiolites, pelagic sediments, and various basin-fill

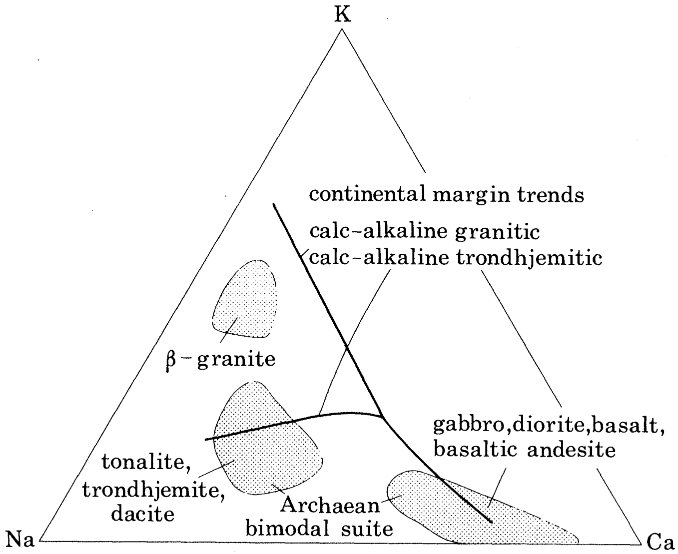

FIGURE 2. Ternary K–Na–Ca diagram (percentage by mass), showing: granitic and trondhjemitic trends, both calc-alkaline; fields of the Archaean bimodal suite; and the field of β-granite (Streckeisen 1976), which is at or close to minimum-melt composition. Actually, trends of some continental-margin batholiths extend from the granitic–trondjhemitic 'fork' to the β-granite field.

sedimentary and volcanic rocks, as well as continental-margin rocks of sedimentary and igneous origin. These igneous rocks include magmatic arcs, the subduction-related, plutonic-volcanic belts of mafic to siliceous rocks that form parallel or at low angles to the convergent plate boundaries. The plutonic parts of magmatic arcs form the so-called Cordilleran or continental-margin batholiths. Tonalite is a major rock type in them. In contrast to the bimodal tonalite–trondjhemite and metabasalt suites of Archaean terranes the post-Archaean continental-margin batholiths are compositionally continuous. Though both Archaean bimodal and most continental-margin suites are calc-alkaline, they show important chemical differences, like the K–Na–Ca relations of figure 2.

Some of the best known continental-margin batholiths are the Mesozoic–Cainozoic circum-Pacific ones of western America, which extend discontinuously from Alaska to Patagonia. These share many of the following characteristics:

(i) a compositional range from quartz diorite to granite and of SiO_2 percentages from the mid-50s to low- or mid-70s; diorite and gabbro also are found (Abbott & Todd 1979; Bateman & Dodge 1970; Cobbing *et al.* 1977; Pitcher 1978);

(ii) intrusion in large composite pulses, or sequences, in order of mafic to siliceous magmas every ten million or several tens of millions of years (Pitcher 1978; Bateman & Chappell 1979; Bateman & Nokleberg 1979);

(iii) a transverse change in composition from less siliceous and more calcic and femic on the oceanic side to more siliceous and potassic on the continental side (Buddington & Chapin

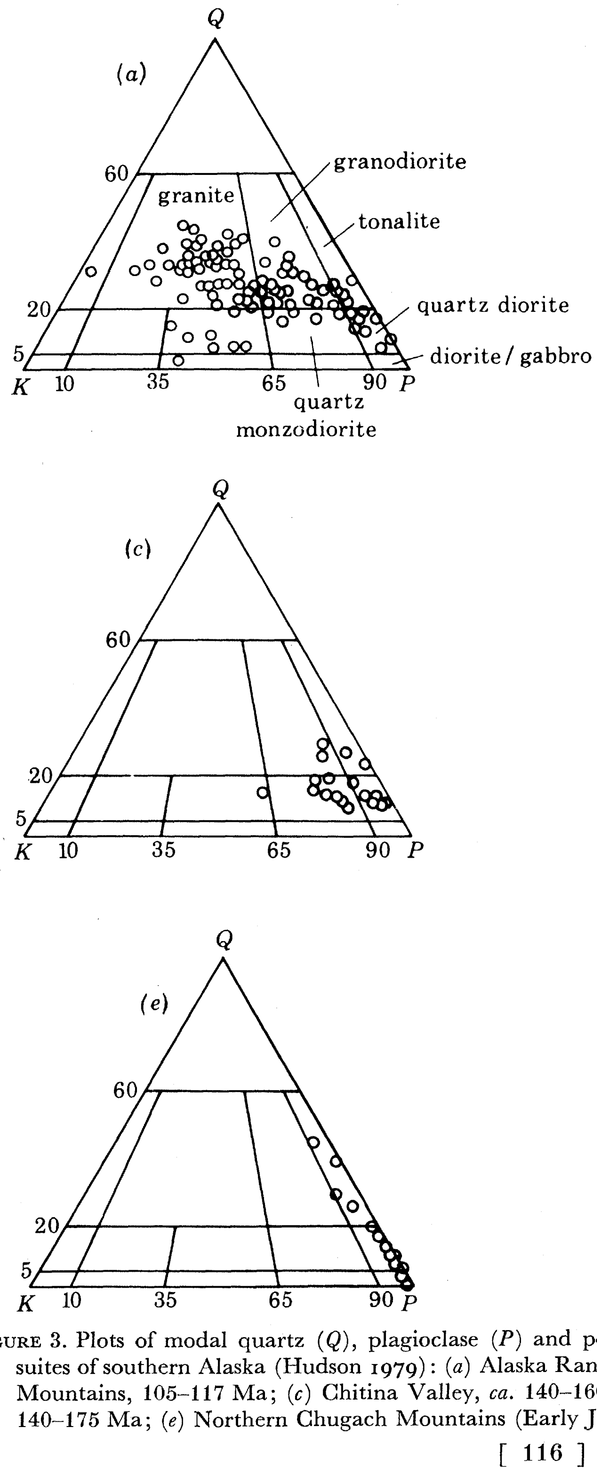

FIGURE 3. Plots of modal quartz (*Q*), plagioclase (*P*) and potassic feldspar (*K*) of six tonalite-bearing plutonic suites of southern Alaska (Hudson 1979): (*a*) Alaska Range, Talkeetna Mountains, 50–75 Ma; (*b*) Nutzotin Mountains, 105–117 Ma; (*c*) Chitina Valley, *ca*. 140–160 Ma; (*d*) Aleutian Range, Talkeetna Mountains, 140–175 Ma; (*e*) Northern Chugach Mountains (Early Jurassic?); (*f*) Kodiak Island (185–195 Ma).

1929; Bateman & Dodge 1970) and an accompanying increase in initial $^{87}Sr/^{86}Sr$ ratios (Kistler & Peterman 1978);

(iv) a proportion of tonalite ranging from about 10% to as much as 55% (latter figure from Pitcher (1978)); and

(v) individual plutons or masses ranging from migmatitic rocks, which usually are quartz dioritic or tonalitic in composition (Hutchison 1970), to plutons containing many mafic inclusions, to homogeneous, inclusion-poor plutons of grandioritic or granitic composition (Bateman *et al.* 1963).

A new example of the diversity of continental-margin plutonism is taken from southern Alaska. Complex convergent and transform plate movements here, involving accretion of Palaeozoic, Mesozoic and Tertiary terranes to the North American craton, caused numerous plutonic episodes. At least six of these produced tonalite-bearing suites over a span of about 150 Ma, as shown in figure 3; some of these suites consist only of quartz diorite and tonalite, whereas others range from diorite to granite.

Origin of continental-margin batholiths is a major problem. We disagree with hypotheses of simple melting of crustal rocks, but rather agree with those that have suggested processes involving interaction of mantle-derived magmas and crustal rocks (see, for example: Presnall & Bateman 1973; Wyllie *et al.* 1976; Brown & Hennessy 1978).

We offer a model for the origin of both single and mafic-to-siliceous multiple magma pulses of continental-margin batholiths. It commences with general models, by Bowen (1922), Presnall & Bateman (1973), Hodge (1974), and Barker *et al.* (1975), of magma–wall rock interactions, and then considers partial melting of the wall rocks. Its locale is above a subducting oceanic plate at a continental margin. The major features of the model are as follows.

(i) Basaltic magma rises from the mantle and pools in the lower to intermediate crust in large chambers (5–30 km long and 10–300 km^3 in volume). The crustal rocks may be of accreted oceanic, various continental-margin or continental-craton types.

(ii) Heat flows from the magma (initially at 1150 to 1250 °C) into the crustal rocks, which start at only 300–600 °C. The magma starts convecting and liquidus phases (e.g. clinopyroxene, olivine, plagioclase) crystallize and tend to settle out. The wall rocks at the magma–wall interface either melt or react with the liquid and precipitate, tending to make the liquid more siliceous.

(iii) Prolonged transfer of heat into the wall rocks causes a gradient of partial melting to be established in them.

(iv) With continued separation of dense phases and assimilation of crustal rocks the magma decreases in density to a critical value, at which it breaches its weakened roof rocks, opens a conduit and ascends buoyantly as a diapir. The SiO_2 percentage content of the magma may be anywhere from the low 50s to the mid-60s, depending on the extent of assimilation and concomitant precipitation of liquidus phases.

(v) As the wall rocks flow inwards and are severely deformed to displace the original magma chamber, their low-density partial melt phase is separated from the residuum by density-driven accretionary processes. The melt progressively collects near the roof of the former chamber and at the lower end of the conduit; the first melt to collect and move up the conduit is the highest fraction of melting and least siliceous composition generated close to the original magma–wall rock interface. More siliceous melts, formed by lower fractions of melting further from that original interface, in the ideal case successively move up the conduit and produce a mafic-to-siliceous sequence of magmas at the site of consolidation that may range from diorite or tonalite to granite.

In our model the first or initial diapir may be 50 to more than 90% of juvenile mantle material. Successive diapirs are largely of crustal derivation, unless diluted with additional magma from the mantle.

Formation of a large continental-margin batholith by this process, involving many tens to perhaps several hundreds of individual reaction cells in the original lower to intermediate crust, will produce four major crustal layers. From the mantle–crust boundary upwards these are: (i) accumulations of liquidus phases, refractory minerals and rock fragments that collect in the lower regions of each cell; (ii) migmatite, gneiss and schist of granulite to amphibolite facies developed from crustal rocks at and above the region of reaction cells and depleted of its low-melting fraction; (iii) plutons of the batholith and minor screens and inclusions of wall rocks; and (iv) volcanic rocks. The roots of continental batholiths thus consist largely of cumulate and refractory material of layer (i), and, because of heat-balance requirements of the interaction process (Taylor 1980), the mass of this layer may be one to five times that of the overlying batholith.

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