Emergent Basalt and Submergent Carbonate–Clastic Sequences Including the Upper Triassic Dilleri and Welleri Zones on Vancouver Island

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Rates, depths, and some conditions of accumulation of epipelagic siliceous lime-shale, coarse bioclastic limestone, oceanic basalt flows, pillow lava, and pillow breccia are inferred from new biostratigraphic data, geometry of occurrence, and analogies between major and minor cyclical sequences. Sporadic thin assemblages of intravolcanic beds, pillowed basalt, and flows within the Karmutsen Subgroup reproduce on a minor scale the major sequence of pre-Karmutsen sediments followed by pillow lava, pillow breccia, and flows of the whole Karmutsen. The Dilleri Zone, documented in several new occurrences, includes not only the uppermost Karmutsen, but also almost the entire Quatsino Limestone, an aggregate thickness of 610-920 m, whereas eight ammonoid zones, from Welleri to Lower Suessi, are encompassed by 300-365 m of epipelagic lower Parson Bay Formation. Although recurrent extinction of benthos and increase in proportions of volcanic components, manifest both in microfacies of intravolcanic beds and in the major post-Karmutsen succession (Quatsino Limestone, Parson Bay, and Harbledown Formations), correlate with increasing depth of sedimentation, maximum depths even during deposition of epipelagic lime-shale were probably never more than 60 m. 'Porosity' of pillowed volcanics and aquagene tuff is discussed. The pillow–pillow breccia–flow sequence represents emergent, the sediments, submergent deposition.

Les vitesses, profondeurs et quelques conditions d'accumulation de calcshiste siliceux épipélagique, de calcaire bioclastique grossier, de coulées de basalte océanique, de "pillow lava" et de brèches à coussinets sont déduites de nouvelles données biostratigraphiques, de la géométrie des affleurements et d'analogies entre les cycles majeurs et mineurs. Des assemblages sporadique et minces de lits intravolcaniques, de basalte à coussinets et de coulées dans le Sous-groupe de Karmutsen reproduit, à plus petite échelle, la séquence majeure suivante: les sédiments pré-Karmutsen suivis de 'pillow lava', de brèches à coussinets et des coulées du Karmutsen. La zone de Dilleri, mieux connue par de nouveaux affleurement inclue non seulement le Karmutsen le plus élevé mais aussi presque tout le Calcaire de Quatsino, une épaisseur totale de 610-920 m, tandis que huit zones d'ammonoïdes de Welleri au Suessi inférieur, sont renfermées par les 300-365 m de la Formation épipelagique inférieure de Parson Bay. Bien que la disparition périodique du benthos et l'augmentation de la proportion de composants volcaniques d'une part se manifestent à la fois dans les microfaciès des lits intravolcaniques et dans la succession majeure post-Karmutsen (Calcaire de Quatsino, Formations de Parson Bay et de Harbledown) et d'autre part correspondent à une augmentation de la profondeur de sédimentation, les profondeurs maxima, même durant le dépôt du calcschiste épipélagique, n'ont probablement jamais excédé 60 m. On discute la 'porosité' des volcaniques à coussinets et du tuff aquagène. La séquence coussinets–brèche à coussinets–coulée représente un dépôt d'émersion, les sédiments un dépôts d'immersion.

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Introduction and Stratigraphic Setting of the Problem

The Upper Triassic submarine basalt assemblages on Vancouver Island include extraordinarily thick and extensive units of tholeiitic pillow lava, pillow breccia, and aquagene tuff and they are associated with, and in part contain, limestones and unusual siliceous sediments. Some of the most instructive rock units fall within the *Tropites dilleri* and *Tropites welleri* Zones of the Upper Karnian Substage (Tozer 1967, pp. 33, 34, 82; Silberling and Tozer 1968). Data recently gathered provide new insights into biostratigraphic and lithostratigraphic relationships and also into a little understood volcanic-marine environment.

Four major, quite distinct lithostratigraphic units are involved (Fig. 1, Vancouver Group), of which the oldest discussed in any detail here and by far the thickest and areally most dominant is the Karmutsen Subgroup. Overwhelmingly volcanic and marine, the Karmutsen consists typically of 8000 to 9000 ft (2400–2800 m) of continuous, closely packed pillow lava, overlain by 2000 to 4000 ft (600–1200 m) of equally continuous pillow breccia and

aquagene tuff, overlain in turn by 9000 to 10 000 ft (2800–3050 m) of massive and amygdaloidal basalt flows. It occurs over almost the whole of central and northern Vancouver Island. Its composition is that of K-poor, so called 'ocean floor' tholeiite (Kuniyoshi 1971), its structure sheetlike. Intercalated within the flows of the upper Karmutsen, particularly in the upper 3500 ft (1100 m) are sporadic and discontinuous sequences of marine bioclastic limestone, siliceous-calcareous *Halobia*-rich shale, and feldspathic wacke ('intravolcanic beds') each from only a fraction of an inch to several feet thick and each nearly invariably overlain by pillow lava or sequences of pillow lava, pillow breccia, and aquagene tuff a few feet to a few hundreds of feet thick (Carlisle 1963). The pillow breccia, both in the minor sequences within the upper Karmutsen and in its major occurrence as the middle unit of the Karmutsen Subgroup is crudely to very well bedded and contains structures indicative of slumping and turbid flow.

The second major unit in the succession, resting directly on the flows without evidence of diastrophism or erosion, is the coarse and richly bioclastic Quatsino Limestone (Dolmage 1919, p. 32B). Sparry cement, oolite layers commonly cross-bedded, an abundance of broken unsorted pelecypod and gastropod shells through most of the unit, fragmental corals, crinoid ossicles, and some preserved algal structures imply a shallow-water relatively high-energy environment. The limestone is light gray and rather pure (Table 1), silica ranging from only 1 to about 13%, mainly in the form of very sparse chalcedony and, in some horizons, as cherty replacement nodules. Stylolitic bedding is common. In a few areas, tongues of pillow lava intercalated within the limestone indicate resurgent volcanism. The thickness of the Quatsino Limestone is commonly in the order of 1000 ft (300 m) although Jeletzky (1970, p. 3) has pointed out that thicknesses may be as little as 100 to 550 ft (30–170 m) and as great as 2500 ft (760 m) or more on western and north-central Vancouver Island.

Thin, dark interbeds appear in the upper 100 ft (30 m) or so of the Quatsino Limestone, and then in abrupt gradation, the lithology changes entirely to the third major unit in the succession beginning with the peculiar black siliceous and carbonaceous lime-shales, calcareous arenites, and fine-grained wackes referred here to the lower Parson Bay Formation of Crickmay (1928). For nearly 1000 ft (300 m) on northeastern Vancouver Island, very fissile laminated beds, packed with flat-lying *Halobia*, in the lower part, or *Monotis*, above the *Halobia* beds, alternate with nonfissile laminated beds. A black to brown microscopically impalpable siliceous matrix is a minor to major constituent of nearly every rock type in the Parson Bay Formation and is responsible for the black color and for other distinguishing physical and chemical characteristics. Microprobe analyses reveal that silica ranging from 6 to 60% and alumina up to 14% are localized largely, and in some calcareous shales, entirely, in this matrix. Noncarbonate carbon is detectable by microprobe and by gravimetric analysis. Nothing in the lower Parson Bay Formation suggests very strong currents, turbulent water or continuous wave action. Graded bedding is absent, with very rare exceptions. The fauna is pelagic and dominated by a single genus in contrast with the abundant and varied benthos of the Quatsino Limestone. Although texturally immature and similar to rocks elsewhere that have been described as deep water, off-shelf deposits, these, as will be shown, are not typical graywackes of the sort that accumulate rapidly in a trench or deep basin. Nor are the thin bedded nonskeletal calcarenites and very fine-grained, pyritic feldspathic wackes of the upper Parson Bay Formation, although cross-bedding and occasional slump features, all rare in the east but more common in the north and west of Vancouver Island, are increasingly apparent upward.

The fourth major unit, the Harbledown Formation is largely, if not entirely, Jurassic, and is included in this discussion because of the rather remarkable analogy that will be drawn between the whole post-Karmutsen sedimentary succession and the sequence of lithologies that one finds within many of the very thin Karmutsen intravolcanic occurrences. The base of the Harbledown Formation is taken where limestone beds no longer occur and the rock becomes light to medium gray color-laminated, pyritic, very fine-grained feldspathic wacke

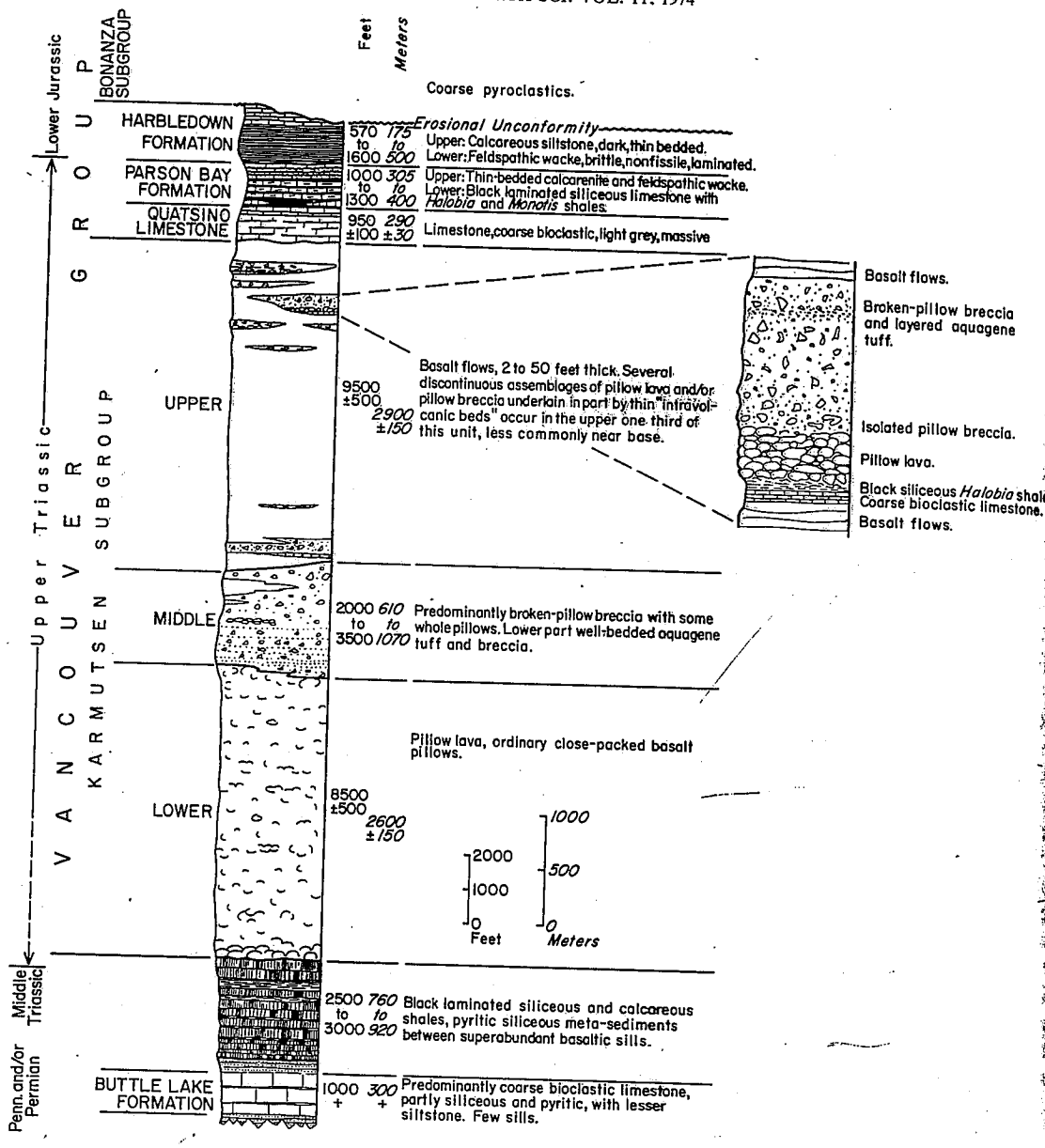


FIG. 1. Stratigraphic column showing analogous minor sequences of sediments and volcanics within the major sequence. Thicknesses are from measurements on northeastern Vancouver Island. Expanded column offset to the right shows a typical upper Karmutsen intravolcanic sediment–pillow assemblage. Several sets of names have been used for post-Karmutsen sedimentary rocks and, in the absence of a type section, the definition of Quatsino Limestone (Dolmage 1919, p. 32B; Gunning 1933, pp. 33–34A; Hoadley 1953, pp. 17–21) has been variously interpreted. Surdam *et* al. (1964), attempting to apply the lithologic definition of Gunning and Hoadley, tentatively included the overlying *Halobia* and *Monotis*-bearing siliceous and carbonaceous lime-shale (lower Parson Bay Formation: Welleri to Lower Suessi Zones of this report) at Iron River in the Quatsino Formation. Surdam (1968) subsequently adopted the inclusion as did Jeletzky (1970, p. 5). Muller and Carson (1969, p. 16), however, placed these same rocks in the Sedimentary Division of the Bonanza Group. On the other hand, Jeletzky (1970, p. 3) places the essentially identical black *Halobia* and

TABLE 1. Compositional ranges of sedimentary facies

	Upper Karmutsen intravolcanic beds			Post-Karmutsen sedimentary units		
	Microfacies 1 Coarse Bioclastic Limestone	Microfacies 2 *Halobia* Lime-shale	Microfacies 3 Feldspathic Wacke	Quatsino Limestone	Parson Bay Formation	Harbledown Formation
			Principal minerals by X-ray diffraction (%)			
Quartz	0–6	3–60	20–30	0.5–13	5–60	5–44
Plagioclase	0	2–30	50–60	0	0–53	36–80
Calcite	92–100	20–93	5–15	88–98	5–93	2–15
			Partial chemical analysis by X-ray fluorescence (%)			
SiO_2	0.5–6.0	3.5–62	50–70	1–13	6–61	45–75
Al_2O_3	0.3–0.8	0.5–8.7	12	0.1–0.5	0.2–14	10–17
CaO	51–55	15–53	4–13	49–55	14–50	2–13

(Crickmay 1928; Carlisle 1972, p. 22). Impure, calcareous beds appear again higher in the formation and it is in some of these that the Sinemurian ammonite *Arniotites* is found (Crickmay 1928). As shown by Jeletzky (1970, pp. 10–20) the Harbledown Formation is replaced laterally on western Vancouver Island by coeval volcanic rocks of the Bonanza Subgroup.

The cyclical repetition of conformable volcanic and sedimentary rock types on the very small scale of the upper Karmutsen intravolcanic beds and on the very large scale of the whole Vancouver Group[1] is striking. Is there an interaction of volcanic and marine environments which could explain the orderly sequence of benthic biogenic carbonate, pelagic lime-shale and wacke, pillowed volcanics and flows, and could this interaction have occurred with such a widely varying periodicity, over a span of at least 20, perhaps 80 m.y.? The trend of each of the sedimentary successions is toward increasing proportions of volcanic components. Is the extinction of benthonic forms a consequence of this volcanic trend or does it simply reflect changes in submarine topography?

In Part I below, new lithostratigraphic and paleontologic data are presented documenting the cycles more fully and showing that whereas much of the upper Karmutsen and almost the entire Quatsino Limestone—an aggregate stratigraphic thickness of at least 2000 to 3000 ft (610–920 m)—falls within the single Dilleri Zone, fully eight ammonoid zones, from the Karnian Welleri to the Norian Lower Suessi, are encompassed by roughly 1000 to 1200 ft

[1]Less is known about the details of pre-Karmutsen sedimentary rocks, but the sequence of lithologies in their upper part, from coarse bioclastic limestone (Buttle Lake Formation, Fig. 1) to dark siliceous and calcareous shales with at least occasional layers rich in flat pelecypods (see *Daonella* in Fig. 2) and to Harbledown-like feldspathic wacke and bedded chert, also parallels quite closely that of the intravolcanic and post-Karmutsen beds.

Monotis-bearing siliceous lime-shale, wacke and calcarenite of the 'thinly bedded member' on the west coast and the Parson Bay Formation on the northeast coast of Vancouver Island in the Sedimentary Division of the Bonanza Group. All of these predominantly pelagic interbedded black calcareous shales, laminated nonfissile limestone, nonskeletal calcarenites and wackes with ubiquitous impalpable black siliceous matrix are strikingly different from the coarsely bioclastic light gray Quatsino Limestone of Dolmage (1919) and are now recognized (Carlisle 1972; Muller 1972) as lithological equivalents of the Parson Bay Formation of Crickmay (1928). Harbledown Formation is distinguishable from Parson Bay on the lithological criteria of Crickmay. The threefold formational subdivision adopted here is entirely consistent with original descriptions, is unambiguous and, except for Harbledown Formation, which is replaced by coeval Bonanza volcanic rocks in the west, is readily recognizable over the whole of northern and central Vancouver Island. Implicit in the correlations is an appreciable modification of the concept of an early to middle Norian depositional trough extending from Sproat Lake through Iron River and eastern Quatsino Sound to Graham Island (Jeletzky, 1970). The term Bonanza Subgroup is used here only for the coarse pyroclastics and other volcanic rocks of intermediate composition (with or without intercalated thin sedimentary units) formerly assigned to the Bonanza Group, Volcanic Division.

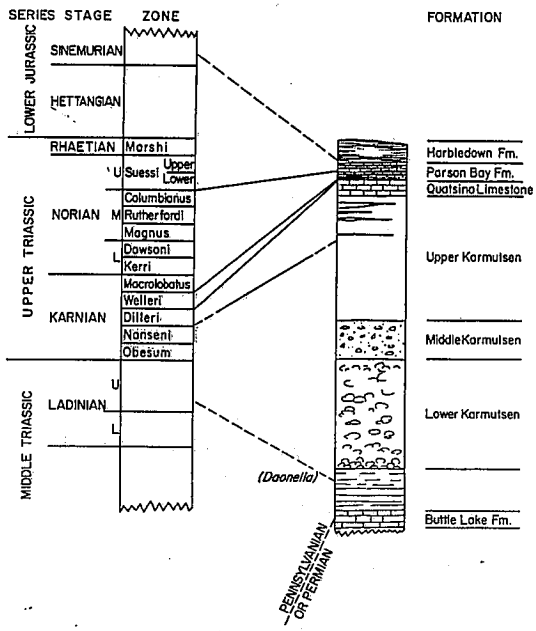


FIG. 2. Biostratigraphic–lithostratigraphic correlation chart. The singular occurrence of a Kerri Zone fauna in limestone of Quatsino lithology at Kenny Point on eastern Quatsino Sound (Jeletzky 1970, p. 3; text herein) is not represented on the chart.

(300–365 m) of the lower Parson Bay Formation (Fig. 2). Part II is an interpretation of these data in terms of rates and environments of deposition of coarse bioclastic limestone, pelagic siliceous lime-shale, and what we have called an emergent basalt sequence.

Part I Lithostratigraphic and Faunal Data

(A) Occurrence, microfacies, composition, and fauna of the intravolcanic beds (Dilleri Zone) in the Upper Karmutsen Subgroup

Geometry in relation to flows and pillows

Thirty to forty occurrences of intravolcanic sedimentary beds have been mapped or noted by the writers at many different horizons through more than 3000 ft (1000 m) of upper Karmutsen massive flows. They are typically only a fraction of an inch to a few feet thick, though in one case over 80 ft (24 m) thick, and less continuous along strike, even, than the pillowed volcanics, which almost invariably overlie them (Fig. 1, offset column, also Fig. 6*e*). Exposures are from a few feet to a few hundred or thousand feet in length.

Contacts of intravolcanic beds with the underlying massive flows are perfectly concordant. Nor are there recognizable unconformities within the sequences. In each occurrence sedimentation terminated with the outpouring of pillow lava, rarely flows, causing the upper layers of wet sediment to be deformed and commonly to be squeezed upward for several feet between pillows. This, and the fact that the beds pinch out gradationally along strike without any evidence of erosion or of disruption other than that caused by the pillows indicate that the present size and shape of each occurrence is not a result of deposition over a wide area followed by a period of erosion and partial removal, but is, in fact, the size and shape of the depositional basin.

Usually the pillow lava is overlain by isolated-pillow breccia with a matrix of aquagene tuff and this, in turn, by broken-pillow breccia with the same matrix in an orderly sequence, though any component of the sequence may be missing in a particular occurrence (Carlisle 1963). In essence, then, each intravolcanic sedimentary assemblage is merely the basal member, and typically the thinnest and least extensive member of a sediment–pillow cycle from a few feet to several hundred feet thick, isolated geographically and stratigraphically in a terrane consisting overwhelmingly of massive basalt flows. The pillowed components of the cycle, like the sediments, also pinch out or grade into flows, along strike, in this case over a few hundred feet to a few miles, also without evidence of structural discordance or erosion prior to deposition of the overlying flows. As with the sediments, therefore, the pillowed assemblages, too, can be said to have retained a size and shape identical with that of their depositional site.

All of these observations indicate that the intravolcanic beds represent local rather than regional interruptions in the accumulation of massive basalt flows; that they developed recurrently but in isolated short-lived sites and in each case as part of an orderly sediment-pillow cycle.

But were these sites merely open places on the ocean floor temporarily free from volcanic accumulation; were they near the ocean surface; or did they change appreciably in depth through their life?

Microfacies and fauna

The lithology of the intravolcanic sediments and the nature of the fauna suggest three fairly distinct microfacies (Table 1), each of which can be represented by a few inches to a few feet of strata or it may be absent. Where more than one facies is represented, they are in the following order from base to top:

Microfacies 1 (Fig. 3*a*) Massive bioclastic limestone; sparry grain-supported skeletal calciruddite, nonfissile, light to dark gray, relatively pure ($CaCO_3$ 92–100; SiO_2 0.5–6.0%), except at the basal contact. In most occurrences this limestone does not contain recognizable ammonites, but is packed with a great variety of unsorted mostly fragmental pelecypods—including *Halobia*—gastropods, brachiopods (*Spiriferina* sp.), crinoids (*Isocrinus* sp. cf. *I. californicus*), corals, and other benthonic organisms as well as pellets and intraclasts. On Hisnit Island in Ououkinsh Inlet (Lat. 50°10′05″, Long. 127°28′20″), however, intravolcanic limestone of this facies contains *Tropites* sp. indet., *Paratropites sellai* Mojsisovics (of Smith), *Paratropites* cf. *P. sulcatus* (Calcara of Gemmellaro), *Trachysagenites* cf. *T. herbichi* Mojsisovics, *Pleuronautilus* cf. *P. alaskensis* Kummel, evidently representing the Dilleri Zone (Tozer 1967, p. 82; Jeletzky 1970, p. 2; Muller and Rahmani 1970, p. 16).

This is a very shallow-water, relatively high-energy facies with minor pelagic components and is essentially identical with much of the Quatsino Limestone, particularly its lower part.

Microfacies 2 (Fig. 3*b*) *Halobia* lime-shale, finely laminated, siliceous, matrix supported, or matrix-rich, dark gray carbonaceous limestone. Microscopically, this facies commonly consists of a cumulate of extremely abundant, flat-lying delicate *Halobia* shells, slightly deformed but commonly unbroken, resting in a black to brown impalpable siliceous matrix forming 50 to 60% of the rock and identical with the matrix of the lower Parson Bay Formation. Calcic spheres approximately 0.1 mm in diameter (algal spores?) and angular to foliate calcite intraclasts up to 2 mm are almost invariably disseminated through the cumulate.

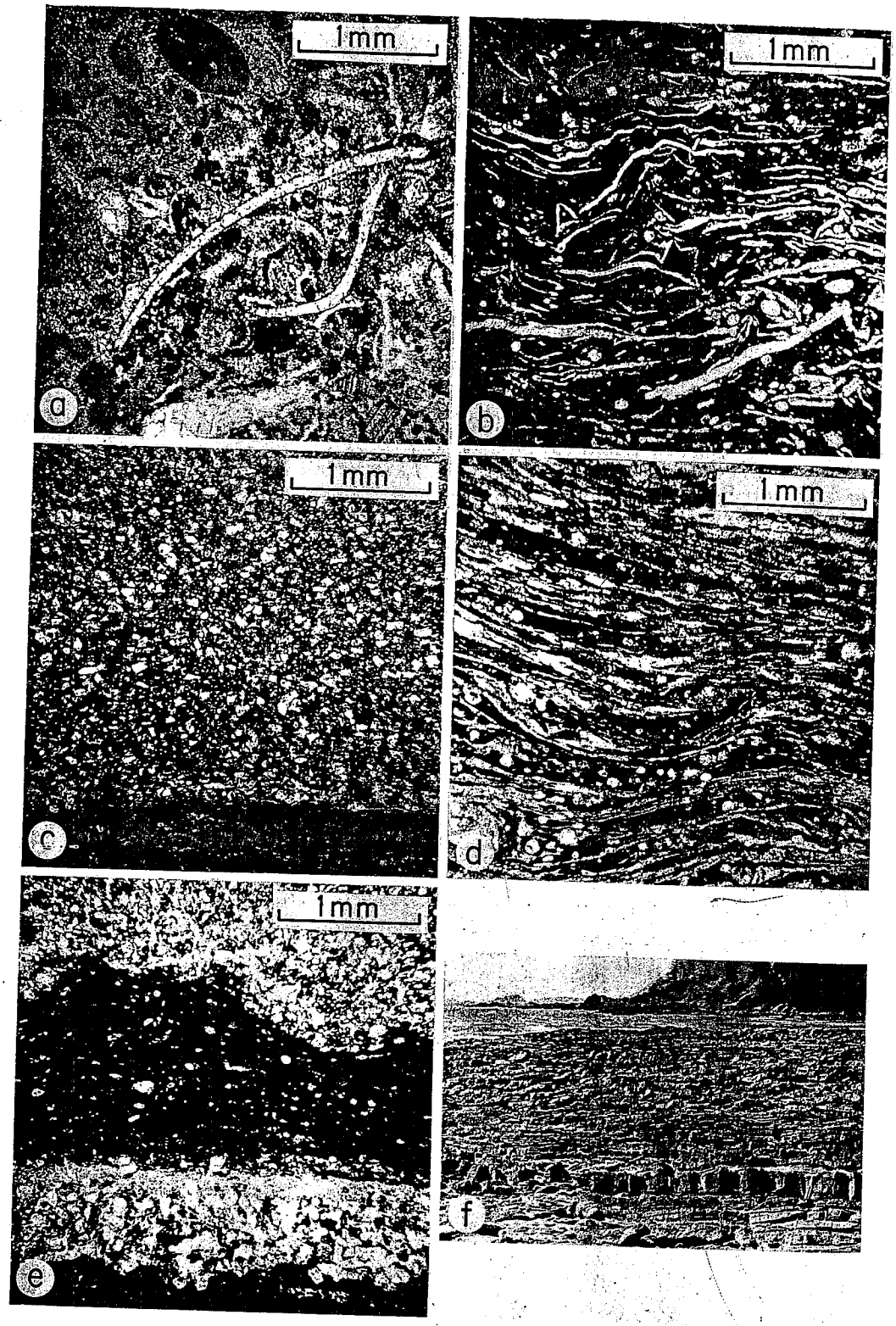
Quartz shown as constituting from approximately 3 to 60% of the rock and plagioclase from 2 to 30% in Table 1, are, with very rare exceptions, present only as cryptocrystalline components of the black matrix identifiable by X-ray or, in the case of quartz, as secondary chalcedony. Noncarbonate carbon is roughly 20% of total carbon. The lamination is a consequence of varying proportions of components; fissility results simply from the abundance of flat-lying *Halobia* shells.

As in the lower Parson Bay Formation, the ubiquity of extremely fine-grained matrix, the siliceous and carbonaceous components, the scattered grains of diagenetic pyrite, the well developed lamination combined with a lack of current structures, grain gradation or bioturbation, and the delicacy of *Halobia* shells indicate a quiet organic-rich environment, probably deficient in oxygen at the water–sediment interface.

Textural features of this microfacies are not uncommonly masked by recrystallization and local redistribution of both silica and calcium carbonate. In some laminae or patches, SiO_2 from the black matrix has replaced carbonate or has recrystallized as spherulitic chalcedony giving the rock a cherty appearance; in others calcite has replaced much of the black matrix. Strongly recrystallized beds are especially susceptible to leaching of the calcite, particularly in the weathered zone and in the presence of pyrite, resulting in a very punky, blocky-fracturing, siliceous rock.

An ammonoid fauna abundant in some of the thicker intravolcanic occurrences, sparse or absent in others, includes *Spirogmoceras shastense* (Smith), *Traskites* sp. *Sagenites* (*Trachysagenites*) cf. *S.* (*T.*) *herbichi* (Mojsisovics), and *Paratropites* sp., and is indicative of the Dilleri Zone (Givens and Susuki 1964; see also Mathews and McCammon 1957, p. 36; Muller and Carson 1969, p. 13, for occurrences on Texada Island; and Tozer 1967, p. 82 for an intravolcanic collection made by Jeletzky on Union Island).

Microfacies 3 (Fig. 3*c*) Feldspathic wacke or microwacke; light brown nonfissile, matrix-supported, very fine-grained fragmental crystal wacke or wackestone. The grains and clasts are: (*1*) microlites, serrate grains, and angular fragments of unzoned intermediate plagioclase up to 0.2 mm diameter making up 10 to 40% of the rock; (*2*) similarly small but highly irregular grains of turbid calcite constituting up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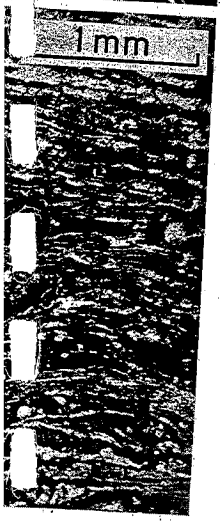
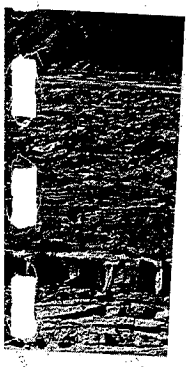




to 15%; and (*3*) chalcedony as minute grains, felted masses, and rounded aggregates up to 0.1 mm in diameter. The chalcedony has apparently resulted from postdepositional recrystallization from the dark impalpable matrix, which, as in microfacies 2, makes up one half to three quarters of the volume. Pyrite grains and clusters, commonly very large, may constitute 2 to 3%. Fossils are absent.

This is the least abundant of the three microfacies. It occurs commonly as laminae intercalated within the upper parts of microfacies 2 in which case the assemblage resembles the upper Parson Bay Formation more than the Harbledown. Near the head of Ououkinsh Inlet on western Vancouver Island, this facies is interlayered with both microfacies 2 and 3.

Graded bedding is manifest in many laminae by a decrease in the size and abundance of plagioclase grains upward and commonly there are minute load casts on underlying layers. In some cases, the laminae are irregular, lenticular, or contorted rather like the composite turbidites described by Carozzi and Frost (1966, pp. 567–568).

Nothing in these subpillow sediments requires proximity to a continent. Terrigenous rock fragments or minerals such as muscovite, zircon, rutile, or tourmaline are completely absent. All of the components could have been derived from the basalts, from marine organisms or from chemical or colloidal solutions or exhalations. Although it cannot be shown from the sediments alone whether some part of the Karmutsen flows were at or above sea level immediately prior to or during sedimentation, the plagioclase grains and much of the matrix in microfacies 3 could well have been derived by weathering of basalt prior to transportation by turbid flow. Ash fallout is a less probable source in this entirely basaltic terrane. Radiolaria have not been recognized with certainty in any of the microfacies but whether this is owing to absence or to destruction during recrystallization is not clear.[2]

(*B*) *Occurrence, Microfacies and Fauna of the Dilleri Zone in the Quatsino Limestone*

Because the existence of Dilleri Zone fauna in post-Karmutsen rocks (Surdam *et al.* 1964; Carlisle and Susuki 1965; Tozer 1967; Table 1) has been questioned or denied (Muller and Carson 1969, p. 15; Jeletzky 1970, p. 2) and because this bears on the matter of depositional rates, it is appropriate to at least outline the principal known and new occurrences in the Quatsino Limestone (Fig. 4) and to demonstrate their indigenous character. Fauna are listed in Appendix A.

Open Bay, Quadra Island

The 'Open Bay Formation' on Quadra Island is a metamorphic tectonite along the margin of the Coast Range batholith, marbleized, highly deformed and intruded by sill-dikes, but nevertheless retaining the characteristic light and dark banding of the uppermost Quatsino Lime-

[2]It is interesting that *Tropites dilleri* itself has not yet been reported from the intravolcanic beds. An assemblage containing *Spirogmoceras shastense* Smith and *Paratropites* cf. *P. sulcatus* (Calcara of Gemmellaro) has been reported by Silberling (1961, p. 538) in the upper part of the Natchez Pass Formation in northeastern Nevada where it likewise is not associated with *Tropites*. Givens and Susuki (1964), referring to the intravolcanic beds, and Silberling referring to the Upper Natchez Pass Formation have suggested that the *Spirogmoceras*-bearing fauna may be indicative of the earliest part of the Late Karnian. In view of the great stratigraphic thickness of the Dilleri Zone, including the *Tropites*-bearing Quatsino Limestone on Vancouver Island and the apparent lack of *Tropites dilleri* itself in the intravolcanic beds, it is possible that an informal subdivision of the zone may be useful in this region.

FIG. 3. Sedimentary facies. (*a*) Intravolcanic microfacies 1. Coarse bioclastic limestone. Occurrence between Iron River and Upper Campbell Lake. (*b*) Intravolcanic microfacies 2. *Halobia* lime-shale. Black background is the siliceous matrix discussed in text. Occurrence between Iron River and Upper Campbell Lake. (*c*) Intravolcanic microfacies 3. Feldspathic wacke resting on microfacies 2. Sharp angular clasts are plagioclase, irregular grains are calcite. Occurrence on Quadra Island. (*d*) Lower Parson Bay Formation. *Halobia* lime-shale in the Welleri Zone at Iron River. Calcite spheres are variably replaced by chalcedony. (*e*) Lower Parson Bay Formation. Lamella of feldspathic wacke in the Columbianus Zone at Iron River. (*f*) Dilleri Zone in foreground (hammer) in uppermost Quatsino Limestone. The overlying Welleri Zone with black fissile limestone and lenticular ammonite-bearing nonfissile limestone beds is midway in the photograph. Occurrence is at Klaskino Inlet on the northwest coast of Vancouver Island.

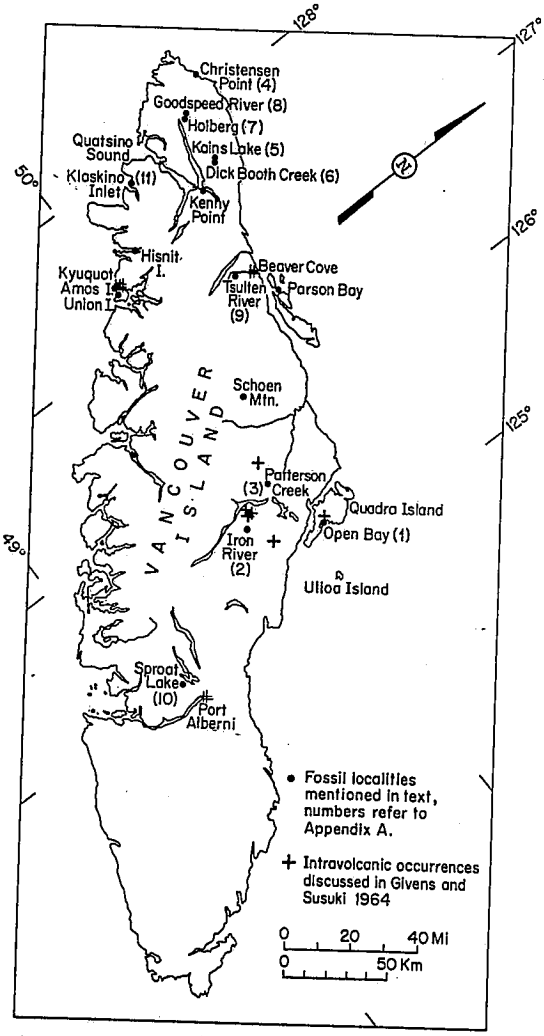


FIG. 4. Fossil localities mentioned in the text including many of the areas from which stratigraphic sections have been described.

stone (Carlisle and Susuki 1965). Contrary to a view once expressed (Muller and Carson 1969, p. 13), no similar limestone, nor so extensive a limestone has ever been found among Karmutsen intravolcanic beds. Unfolded, the thickness would be in the order of one hundred to a few hundred feet (30 m). On Quadra Island the unit can be followed for over 9 mi (14.5 km) in a belt up to 1 mi (1.6 km) across. Karmutsen flows lie to the west across a zone of shearing; folded upper Parson Bay calcarenites and siliceous beds lie to the east. Thin units of pillow lava, folded with the limestone, are examples of resurgent Karmutsen volcanism.

The tropitid assemblage at Open Bay (Appendix A, *1*), which is unusually abundant and varied, displays an interesting microstratigraphic arrangement in association with very continuous, thin pyritic chert lamellae. *Hannaoceras*, the most abundant genus overall, is restricted to a single group of limestone beds totalling 1 to 3 ft (0.3–1 m) in apparent thickness. *Arcestes*, the next most abundant genus, is found with *Hannaoceras* and in beds for several feet above it, where locally it may constitute 70% of the forms present, but is found nowhere below *Hannaoceras*. *Tropites dilleri* Smith is very common and is present throughout the fossiliferous section but is most abundant in beds where *Hannaoceras* is most abundant. The authors have suggested (1965, p. 458) that this microstratigraphic distribution, in association with undisrupted chert lamellae, which themselves may have been organically controlled, may reflect local environmental factors, perhaps food supply. This, in conjunction with the continuity of thin beds, the lack of any evidence for slump or turbidites, and the abundance and variety of the fauna, suggests that the Dilleri Zone assemblage at Open Bay, if not actually *in situ*, is not far removed from the environment in which the animals died.

Northeastern Vancouver Island

At Iron River (Appendix A, *2*), 24 mi (39 km) southwest of Open Bay, 405 ft (123 m) of exposed, faulted Quatsino Limestone is thought to represent approximately one half of the total Quatsino. Abundant *Tropites dilleri* Smith occur in the typically layered uppermost part only 30 to 80 ft (9–24 m) stratigraphically below the Parson Bay Formation (Surdam *et al.* 1964, Lower Quatsino Formation of these authors is equivalent to Quatsino Limestone herein). *Halobia* is omnipresent, most notably in the thin, black, fissile interbeds. The ammonites, however, occur mainly in the nonfissile beds and particularly in nonlaminated layers within these, where *Halobia* shells are jumbled and fragmented. Small gastropods are locally abundant. Intraclasts of reworked limestone and eroded echinoid ossicles are present, but there is nothing to suggest strong current action. The

jumbling and fragmentation is probably a result of burrowing or browsing. Chert lamellae are absent though redistributed silica in the form of highly amorphous nodules, crudely lenticular bodies often with fossil debris, and in stilolitic sutures is common.

The microfacies of these Dilleri beds at Iron River, viewed in the context of overlying and underlying strata, appear to represent shallow intermediate depths, but with no evidence of currents competent enough to erode and transport the fauna once it had accumulated.

A Dilleri Zone fauna, considerably more complete than at Iron River occurs at Patterson Creek (Appendix A, *3*) 16 mi (26 km) to the northwest in lithologically identical upper Quatsino beds. A few hundred feet to the west, tongues of pillow lava and pillow breccia take the place of part of the lower and middle Quatsino Limestone, much as they take the place of upper Quatsino beds at Open Bay.

Medial Vancouver Island

Along the central axis of Vancouver Island, both in the extreme north and in the south near Port Alberni, fauna characteristic of the Dilleri Zone occur in Quatsino Limestone, again commonly only a few feet below Parson Bay Beds. On the north coast, east of Christensen Point in a somewhat faulted section, first described by Dawson (1887, p. 76B), *Tropites* cf. *T. dilleri* Smith was recognized by the authors in typical light and dark bedded upper Quatsino Limestone not less than 600 ft (180 m) above Karmutsen flows. In the same belt of limestone southeast of Kains Lake *Paratropites*, a characteristic member of the Dilleri Zone, occurs in light gray, bioclastic and cherty, middle Quatsino Limestone along with indeterminate colonial corals, questionably *in situ*. The rock has none of the dark interbeds that characterize the uppermost Quasino Limestone. Along Dick Booth Creek, 3 mi (4.8 km) to the southeast, where the Quatsino–Parson Bay contact is well exposed, a Dilleri Zone fauna can only be inferred from ammonite cross-sections on eroded surfaces of typical uppermost Quatsino beds. *Tropites* cf. *T. dilleri* Smith occurs in upper Quatsino Limestone in a quarry 2 mi (3.2 km) northeast from Holberg (Appendix A, *4*, *5*, *7*).

At Sproat Lake (Appendix A, *10*) on southern Vancouver Island, 185 mi (300 km) to the southeast from Christensen Point and 120 mi (193 km) southerly from Iron River, but still along the central axis of the Island, *Tropites* cf. *T. dilleri* Smith occurs in the lower part of the upper Quatsino Limestone 650 ft (200 m) above Karmutsen flows in continuous section and 200 ft (60 m) below Parson Bay beds, with a Welleri Zone fauna. The Dilleri rock is packed with completely unsorted broken shells, pellets and intraclasts in sparry cement and is quite clearly the product of a high energy environment.

Western Vancouver Island

The most complete and instructive new occurrence of Dilleri Zone fauna, discovered as the result of a suggestion by J. E. Muller and examined in part with T. E. Tozer in 1970, is on Klaskino Inlet (Appendix A, *11*). It establishes for the first time in western British Columbia, the presence of well documented Dilleri and Welleri Zones in an unbroken conformable succession of strata.

The Dilleri Zone is represented in three distinctive beds in uppermost Quatsino Limestone. The main bed, 1 to 3 ft (0.3–1 m) thick, is light pinkish-gray, recrystallized, coarse, bioclastic limestone much like other Dilleri Zone occurrences but with larger—1 to 2 mm—and more numerous intraclasts and with ammonites at random. A diagenetic nodular or cobbley structure distinguishes it from beds beneath it. Unlike other occurrences, tropitids also occur abundantly in the laminated, slightly fissile interbeds for 6 to 10 in. (15–25 cm) above the massive bed. However, one does not find the characteristic microsequence, which was observed at Iron River, where finely laminated nonfissile *Halobia* cumulates become disturbed only in the ammonite—and gastropod—bearing layers. All of the rock is predominantly bioclastic, ill sorted and arenaceous in texture. The impression gained by microscope is of appreciable disruption and probably some transportation of the fauna, but not so violently as to cause recognizable erosion at this site.

(*C*) *Lithofacies and Fauna of the Lower Parson Bay Formation (Welleri to Lower Suessi Zones)*

With one recorded exception, discussed be-

low, the Dilleri–Welleri boundary on Vancouver Island corresponds fairly closely with the transition from coarse bioclastic, predominantly benthonic limestone to predominantly pelagic black lime-shale. The lithologic change, analogous to the transition from microfacies 1 to microfacies 2 observed in some intravolcanic beds, marks the base of the Parson Bay Formation. It also marks an order of magnitude decrease in the stratigraphic thicknesses of ammonoid zones (Fig. 2).

Quadra Island, Ulloa Island, and Parson Bay

Neither the lowermost Parson Bay Formation nor a separate Welleri Zone have been recognized on Quadra Island, though *Tropites* cf. *T. welleri* does occur in the same beds as *Tropites dilleri*. Highly metamorphosed *Monotis* beds (Lower Suessi) form a granitic roof pendant on Ulloa Island. On Parson Island in Parson Bay discussed first by Crickmay (1928), the Welleri Zone may have been omitted by faulting. *Juvavites* sp. occurs in *Halobia* lime-shale with sparse plagioclase clasts and locally with convolute bedding and slump structure. Nonskeletal calcarenites increase upward in a discontinuous section and the Lower Suessi *Monotis* beds include coarse feldspathic wacke and grit with graded bedding, scour and fill, load casts and intraformational conglomerate with clasts up to 8 in. (20 cm) long.

Northeastern Vancouver Island

At Iron River (Appendix A, *2*) the Welleri Zone is reasonably inferred from the presence of *Discotropites* sp. in Parson Bay Beds some 90 ft above the *Tropites dilleri* bearing beds. The ammonite bearing layers are nonfissile limestone, quite black, subordinate to the very fissile *Halobia* lime-shales and in the form of lenticular beds up to a foot or so thick and a few feet long. These lenses occur both in single layers and, unlike boudins, as isolated bodies at several horizons. The lime-shale is compacted differentially around them. They show no erosive contacts, are ill sorted and only vaguely laminated in outcrop. On one exposed bedding surface the lenses were observed to be elongate with a well developed north–south preferred orientation. They resemble large load-casted ripples or dunes (Dzulynski and Walton 1965, pp. 146–149, 157).

Microscopically, the lenses are composed of pelecypod and ammonoid shells, unidentified microorganisms at random, small intraclasts, some of which may be echinoid in origin, and very abundant recrystallized organic (?) calcite spheres roughly 0.1 mm in diameter, all supported by the abundant characteristic black carbonaceous and impalpable siliceous matrix. The surrounding lime-shales are essentially delicately preserved cumulates of *Halobia* with variously abundant calcite spheres, commonly silicified, and sparse intraclasts (organic?) supported by the same black carbonaceous siliceous matrix (Fig. 3*d*). Identical textures have been described by Sutherland Brown (1968 pp. 51–54) in the black limestone member of the Kunga Formation on Queen Charlotte Island, unquestionably correlative with the lower Parson Bay Formation. Grain gradation and bioturbation both are absent at Iron River. Pyrite is present mainly in the shale. The similarity between this rock and the *Halobia* lime-shale facies of the intravolcanic beds, not only texturally, but also mineralogically and chemically (Table 1) is striking. The sediment undoubtedly accumulated in very quiet water with only occasional surficial or perhaps internal waves if indeed the lenses are wave-formed.

Above the inferred Welleri Zone at Iron River very sparse graded laminae of feldspathic wacke appear in the Columbianus Zone (Fig. 3*e*) and there is one intraformational conglomerate bed 2 ft (0.6 m) thick within the *Monotis* beds containing reefoid limestones, but there are no coarse or cross-bedded sandstone or other structures indicating persistent strong currents as at Parson Bay. An isolated reef limestone rests on *Monotis* beds in the upper Memekey River area 24 mi (39 km) to the northwest (Carlisle 1972, p. 22). Above the *Monotis* beds at Iron River, through 300 ft (92 m) of upper Parson Bay Formation, fossils become extremely sparse, fissility decreases, nonskeletal calcarenites become interlayered with extremely fine-grained feldspathic beds, commonly with interrupted laminations and occasionally with convolute bedding, but the impalpable siliceous matrix and the black color remains. These beds are succeeded by

nearly 600 ft (180 m) of Harbledown color-laminated feldspathic wacke.

Medial Vancouver Island

East of Christensen Point on the northern tip of Vancouver Island where the Welleri Zone (Appendix A, *4*) is only 30 ft (10 m) stratigraphically above the Dilleri; on Goodspeed River, six mi (15 km) by road north from Holberg (Appendix A, *8*) and near Beaver Cove on a tributary of the Tsulten River (Appendix A, *9*), the Welleri fauna are again concentrated in lenses of bioclastic nonfissile dark limestone with *Halobia* bearing lime-shales, as at Iron River, over a stratigraphic interval of 15 ft (4.5 m) or less. At Dick Booth Creek (Appendix A, *6*) *Tropites* cf. *T. welleri* Smith occurs mainly in thin continuous nonfissile layers within wavy-laminated *Halobia*-packed lime-shales in a 25 ft (7.6 m) thick section transitional between Quatsino and Parson Bay Lithology. Nonfissile lenses, just like those at Iron River and Beaver Cove, appearing a few feet higher in the section have concentrations of arcestid ammonites in the lower few inches suggesting differential settling within an agitated medium. Sedimentary structures indicative of persistent strong currents are again absent however. At Sproat Lake, near Port Alberni, the lithology of the Welleri Zone (Appendix A, *10*) is essentially identical with that at Iron River except for the fact that primary textures are particularly well preserved.

In most of these areas, the typical succession of lower Parson Bay beds up to the *Monotis subcircularis* bearing Lower Suessi Zone is visible, though almost invariably faulted. At Sproat Lake the Monotis beds are replaced by a reefoid limestone previously unrecognized, but apparently correlative with the Sutton Limestone.

Western Vancouver Island

The Welleri Zone at Klaskino Inlet (Appendix A, *11*) occurs over a 10 ft (3 m) interval beginning only 35 ft (11 m) above the Dilleri Zone in a transitional lithology, which the present authors prefer to place in the basal Parson Bay Formation, but which Muller and Rahmani (1970, p. 16) and Jeletzky (1970, Fig. 2) have placed in the uppermost Quatsino Limestone (see subtitle to Fig. 1 herein). Ammonites are especially plentiful, though small in size, and here again they are concentrated randomly in lenses of limestone rich in the black impalpable matrix with abundant *Halobia* fragments quite like the Welleri lithology described several times above, and surrounded by *Halobia*-packed lime-shales (Fig. 3*f*). The lenses are unusually abundant, however, constituting about 70% of the rock. Unlike those described elsewhere or any Quatsino Limestone, they contain numerous slump-folded structures, including intraformational blocks of deformed dark gray limestone up to 2 × 16 in. (5 × 40 cm), and they show load casts on underlying beds. Here, even more clearly than at Parson Bay, significant gravity transport and proximity to a slope is indicated. Higher in the Parson Bay Formation, distinctly arenaceous beds appear, becoming coarse and cross-bedded in the Lower Suessi Zone. Muller and Rahmani (1970, p. 15) have drawn attention to intraformational slump conglomerates which occur below *Monotis* in Middle (?) Norian beds and involve what the present authors identify as reefoid limestone.

Jeletzky has obtained representatives of the Welleri Zone from four localities in the general vicinity of Kyuquot (Tozer 1967, p. 82). Those on the islet south of Amos Island and at Racoon Point on Union Island, examined by the senior author, are within the very uppermost Quatsino Limestone or in the transition from Quatsino Limestone to the Parson Bay Formation. In agreement with Jeletzky (1970, p. 3) we find the Quatsino Limestone to be extremely thin, perhaps only 100 ft (30 m) here, but we suggest that this is due to Karmutsen volcanism having continued much later here than elsewhere on Vancouver Island. The upper part of the section is very similar to that at Klaskino except that reef limestone referred to the Sutton Limestone by Jeletzky (1950; 1970), and by the present authors, commonly rests upon or partly replaces the *Monotis* beds.

Eastern Quatsino Sound

Unlike all of the other known occurrences on Vancouver Island, including those south of Kains Lake, which is only 10 mi (16 km) to the northwest, and those near Holberg only 25 mi (40 km) to the west-northwest (both of which areas are separated from Quatsino Sound by a major fault), the Dilleri–Welleri Zone boundary in the eastern Quatsino Sound

area does not appear to coincide, even roughly, with the change from coarse bioclastic limestone to siliceous *Halobia* lime-shale. *Tropites* sp. indet. does occur in typical Quatsino Limestone on the east side of Holberg Inlet about one third mile (0.5 km) south of Stewart Point (J. E. Muller, personal communication) fixing *Tropites* beds at least in this lithology. Two and one half mi (4 km) to the southeast, however, along the coast northeast of Kenny Point (Lat. 50°34′, Long. 127°31′40″), Jeletzky (1970, p. 3) has reported Quatsino Limestone of an age much younger than any other known occurrence. *Mojsisovicsites kerri* (McLearn), the index species of the lowermost Norian zone, above Welleri and separated from it by the Macrolobatus Zone, occurs in a coarsely bioclastic light gray limestone. The limestone at Jeletzky's localities, examined by the writers, is light gray, coarsely bioclastic and is bedded in the manner of the Upper Quatsino save only that the interbeds are cherty. The occurrence is bounded by faults, but no rock of Parson Bay lithology or rock containing a Welleri Zone fauna is known closer than those already described. Nor has a Welleri Zone fauna yet been found in the faulted belt of limestone extending southeastward from this area to Tahsis Inlet.

Jeletzky's occurrence at Kenny Point demands the conclusion that a shallow-water, high-energy environment continued to produce coarse bioclastic Quatsino Limestone in eastern Quatsino Sound well into the Early Norian, while all of the other areas described above from mid-Late Karnian on, were accumulating only the pelagic dark siliceous lime-shales of the lower Parson Bay Formation. Except for the possibility of major tectonic separation—which must be kept in mind—the eastern Quatsino Sound area in Late Karnian–Early Norian probably should be looked upon as one of continued vigorous shallow-water accumulation within a larger region that elsewhere had already become markedly deprived of biogenic sediments. This late coarse bioclastic facies in eastern Quatsino Sound can not be identified, however, with upper Karnian–Lower Norian, predominantly pelagic, siliceous lime-shale lithology (lower Parson Bay Formation) in the Iron River area as suggested by Jeletzky (1970, p. 3, Fig. 2).

Summation of Part I

The sporadic and thin sequences of intravolcanic sediments, pillow lava, pillow breccia and basalt flows in the upper Karmutsen Subgroup reproduce on a small scale the major sequence from pre-Karmutsen limestone and siliceous sediments to pillow lava, pillow breccia and finally flows of the whole Karmutsen Subgroup (Fig. 1). Likewise, sequences of sedimentary microfacies within the upper Karmutsen intravolcanic beds, even though they occupy only a very small part of a single ammonoid zone, mimic part or all of the whole post-Karmutsen sedimentary succession up to 3000 ft (920 m) thick and ranging in age from Karnian to at least Sinemurian on northeastern Vancouver Island, from Karnian to Upper Norian on western Vancouver Island. An apparently similar but less well documented major sequence of facies is found in the upper pre-Karmutsen sediments.

Contrasting compositions, textures, faunas and durations of sedimentation all indicate much quieter, more persistent, but biologically much less diversified *in situ* environments for the *Halobia* lime-shale than for the coarse and richly bioclastic limestone whether these are found in the major succession of Quatsino and Parson Bay formations or in the minor sequences within the intravolcanic beds. The question that has been raised is whether the facies differences simply reflect changes in depths of deposition or in topography or whether they have involved some more complex interaction of volcanic and marine processes repeated with a widely varying periodicity over a span of at least 20 million years.

Usually one has little more than texture and composition from which to judge depth of deposition and these properties may not be conclusive, especially for pelitic rocks, because, while there are numerous positive and reasonably reliable criteria for sedimentation in shallow water, a great many of which are illustrated by the coarse bioclastic limestone of the Quatsino and of microfacies 1 in the intravolcanic beds, there are very few criteria for deep water sedimentation and many of these are based directly or indirectly on negative evidence. Fineness of grain, matrix support, dominance of pelagic components or of silica, or persistence of lithology through time do not necessarily

imply great depth. In the present case, however, an additional line of evidence is provided by the geometrical relations between the sediments and the basalts taken in conjunction with the paleontological data.

Part II Rates and Conditions of Accumulation of Volcanic and Sedimentary Facies

Aggregate rates of deposition of coarse bioclastic limestone and Halobia lime-shale

If it is assumed that the time represented by each ammonoid zone is in the order of 1 m.y. (Tozer 1967, p. 11; Harland *et al.* 1964, p. 261) and that the Quatsino Limestone, roughly 300 m thick on northeastern Vancouver Island, represents the upper one half approximately of the Dilleri Zone (the lower half being represented by the upper 3500 ft (1070 m) approximately of Karmutsen, an assumption for which there is some support below), then the apparent aggregate rate of Quatsino accumulation can be estimated very roughly as about 0.06 cm yr^{-1}. Similarly, the apparent aggregate rate of accumulation of the lower Parson Bay Formation—also roughly 300 m thick on northeastern Vancouver Island, but representing eight ammonoid zones spanning perhaps 10 m.y. (viz. Tozer 1971, p. 993)—might be estimated as approximately 0.003 cm yr^{-1}; about one twentieth that of the Quatsino Limestone. These are rough estimates of aggregate rates for the rocks in their present state. In order to obtain initial depositional rates, corrections of unknown magnitude must be made to the apparent aggregate rates of accumulation to allow: (*1*) for discrete periods of nondeposition (perhaps more common in the Parson Bay Formation), (*2*) for intraformational erosion and sedimentary cannibalism and (*3*) for dehydration and recrystallization compaction, and intrastratal solution (all perhaps more common in the Quatsino Formation judging from the abundances of reworked intraclasts, recrystallized calcite, and of stylolitic bedding). A semiquantitative confirmation of the estimated disparity in rates is provided by the pelagic components of the rocks, if these can be assumed to have been contributed to both formations at essentially similar rates. *Halobia*, for example, which may have drifted freely or have been attached to algae, is at least 10 times and possibly 20 or 30 times more abundant in the lower Parson Bay Formation than in the Quatsino Limestone. The contrasting silica content is similarly in agreement though not indicative in itself.

The paucity of fossils in the upper Parson Bay and Harbledown Formations, has precluded zonation of these rocks, but from the occurrence of *Arniotites vancouverensis* (Sinemurian) about 230 ft (70 m) above the Lower Suessi Zone (Upper Norian) at Iron River and about 525 ft (160 m) above the top of the Parson Bay Formation (questionably within 30 m of the top) at Parson Bay (Crickmay 1928, p. 59) and using 7 m.y. as the extrapolated time interval between Upper Norian and Sinemurian (Harland *et al.* 1964, p. 261), the apparent aggregate rate of accumulation for these feldspathic rocks (70 m in 7 m.y.) is of the same order of magnitude as for the lower Parson Bay Formation or perhaps as little as one third of that rate.

By way of comparison, Cloud (1962) has estimated Holocene accumulation of wet aragonite in shallow water west of the Bahamas to have taken place at 0.08 cm yr^{-1} corresponding to a rate of 0.038 cm yr^{-1} for compacted and lithified aphanitic and pelletal limestone. Fisher (1964, p. 143) has estimated the rate of deposition for the Triassic Dachstein Formation, a shallow-water, slightly dolomitic limestone, at 0.007 to 0.01 cm yr^{-1}, without corrections for postdepositional thinning. Pelagic sedimentation rates of carbonate and also of diatomaceous and clay oozes in deeper parts of the ocean, but above the level of carbonate corrosion and apparently undistorted by turbidity current deposition, are typically in the range of 0.0005 to 0.005 cm yr^{-1} as reported by several investigators. In view of these data the estimates of 0.06, 0.003 and 0.001 (?) cm yr^{-1}, respectively, for lithified Quatsino Limestone, lower Parson Bay and Harbledown Formations, uncorrected for postdepositional thinning, appear reasonable and will be used as approximate accumulation rates in the succeeding discussion. Because of the likelihood of unrecognizable or unmeasurable periods of nondeposition within a unit as thick as the lower Parson Bay Formation, however, allowance must be made for the fact that individual beds or thin sequences of beds, probably

accumulated more rapidly than the aggregate unit.

Aggregate rate of accumulation of the Karmutsen Subgroup

In the summer of 1971, geologists with a mineral exploration group (Vangulf Exploration Co.) discovered *Daonella* in siliceous and carbonaceous black limeshales, intruded by basalt sills, 100 to 200 ft (30–60 m) stratigraphically below the basal pillows of the Karmutsen Subgroup on Schoen Mountain. E. T. Tozer has provided the following preliminary report on collections by the senior author:

Daonella sp; group of *D. tyrolensis*. Age: Middle Triassic, probably early Upper Ladinian.

Contemporaneous deformation of the sediments beneath the basal Karmutsen pillows indicates uninterrupted succession so it is now possible, for the first time, to estimate the probable duration of Karmutsen volcanism, albeit crudely. This is in the range of 2.5 to 3.5 m.y., the length of time encompassed by early Upper Ladinian to early Upper Karnian (Middle Dilleri Zone) less 1–2 m.y. for the siliceous sediments below the pillows assuming these to have accumulated at a rate similar to that for the lower Parson Bay Formation. The probable overall average rate of accumulation for the 20 000 to 22 000 ft (6100–6700 m) of basalt, therefore, would be in the range of 0.17 to 0.27 cm yr^{-1} (6100–6700 m ÷ 2.5–3.5 m.y.). This is roughly three to four times the apparent aggregate rate of accumulation of Quatsino platform carbonate, but, considering the corrections for postdepositional thinning of sedimentary rocks discussed above, something less than three or four times the rate of initial carbonate deposition. There is nothing in the volcanic rocks to indicate that significant corrections need to be made for postdepositional changes in their thickness. Volcanism was probably more spasmodic and more variable, but, on the average, not enormously more rapid than initial platform carbonate deposition, especially if intrastratal carbonate solution has the large effect that some attribute to it.

Parenthetically, the estimated rate for Karmutsen accumulation just derived is consistent with the assumption made earlier that the Quatsino Limestone represents the upper one half of the Dilleri Zone and that some 3500 ft (1070 m) of uppermost Karmutsen flows (and intravolcanic beds) represent the lower one half (1070 m at 0.17–0.27 cm yr^{-1} = 0.5 m.y. approx.).

Relative depths of deposition of flows, pillows, and pillow breccias–'emergent basalt sequences'

The flows of the upper Karmutsen Subgroup are from 2 to 50 ft (0.6 to 15 m) thick and though local variations in thickness and consequent variations in initial dip of up to 10° or so are common, they are generally continuous over hundreds to thousands of feet along the strike. Applying the Karmutsen average accumulation rate of 0.17 to 0.27 cm yr^{-1} (or perhaps a little less for the upper Karmutsen, as suggested below) to flows of this thickness, one finds that each flow 'represents,' as it were, hundreds to thousands of years of accumulation. Of course, each of these basalt flows, was emplaced in a very short interval of time as evidenced by the chilled contacts, amygdaloidal tops and intersertal to hyalopilitic textures. Moreover, apart from the lenticular intravolcanic sediment–pillow assemblages, which are rare and sporadic, the flows succeed each other conformably without recognizable vestiges of sediments or pillows or incipient pillows either between the flows or engulfed in them, either at the same stratigraphic level as the sediment–pillow lenses or at other levels. It is possible to traverse through the entire 9000 to 10 000 ft of upper Karmutsen without encountering any of the sediment–pillow assemblages, in the process crossing between 500 and 1000 flows each of which probably was emplaced and cooled in a few months. (Occasional, very thin selvages of cherty and chloritic material along some flow contacts are epigenetic alteration products). It is quite clear, therefore, that major amounts of time are not accounted for by the emplacement of basalt flows, or of any other material. This would still be true even if the duration of Karmutsen volcanism, or the fraction implicitly assigned to the upper Karmutsen in the earlier calculation of aggregate rate of accumulation, has been overestimated by a factor of ten, which seems unlikely. Ruling out abyssal nondeposition, the authors are led to conclude, therefore, that the basalt flows of the upper Karmutsen formed above water or in water

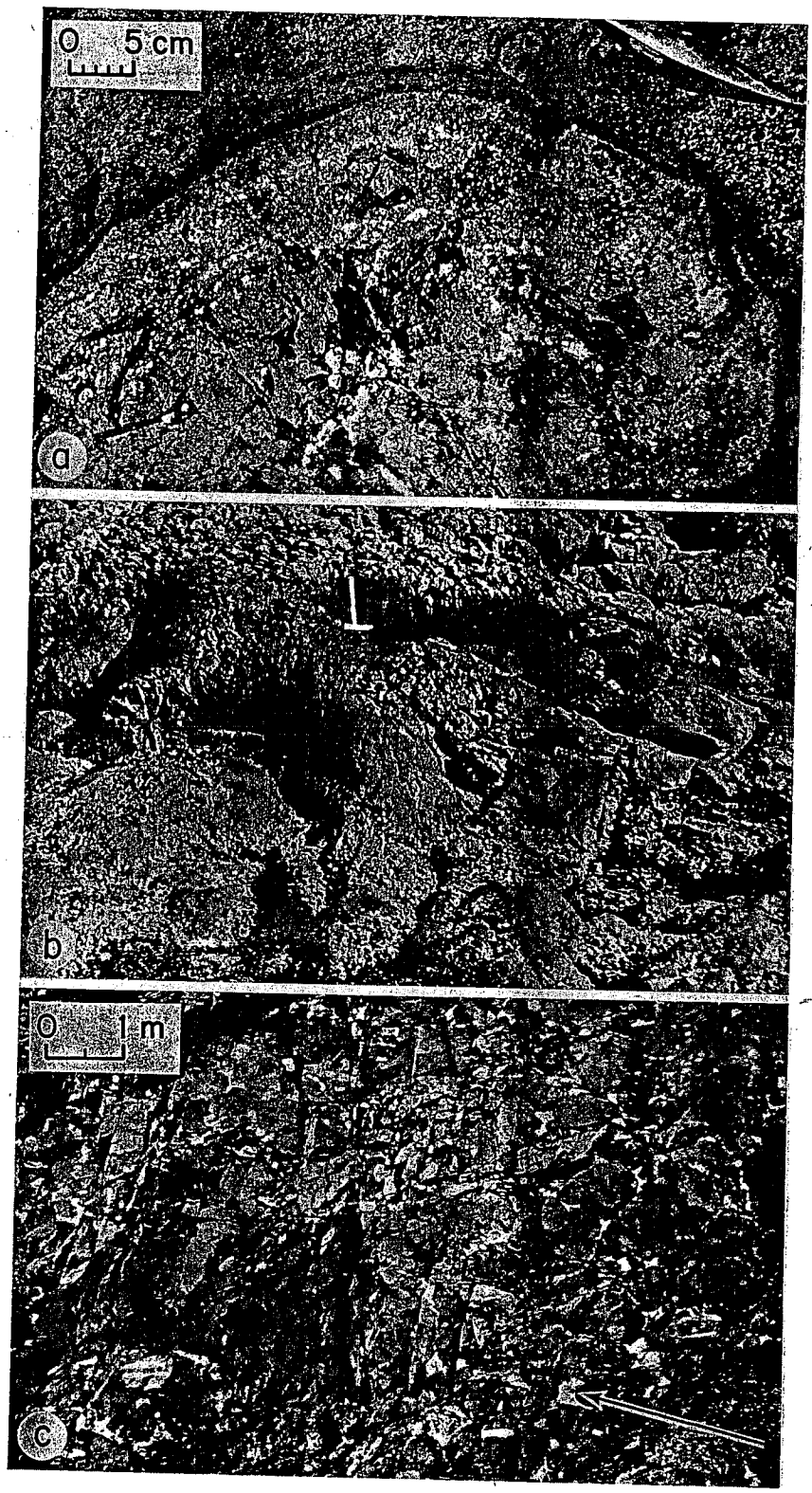
so shallow that no preservable sediment could deposit on their surfaces. The few intravolcanic basins that did develop are in small areas where accumulation of basalt flows was less rapid than subsidence and they were surrounded or nearly surrounded by shorelands of flow basalt. When lava once again entered these basins it brought about submarine accumulation of the pillow lava–pillow breccia sequence, directly on the soft sediment, in every case filling and obliterating the basin completely. The accumulation of pillowed basalt, therefore, was more rapid than subsidence.

More importantly, this cycle from submarine pillow lava to pillow breccia (commonly isolated-pillow breccia overlain by broken-pillow breccia) to subaerial or near subaerial flows is a customary phenomenon both at the small scale of the several upper Karmutsen occurrences and at the order-of-magnitude larger scale encompassing the entire Karmutsen Subgroup. It is prevalent also in other parts of the world in both older and younger rocks. Moore and Fisk (1969) concluded from the distribution of dredged and photographed pillows and pillow fragments around Hawaii that during eruption of pillows and pillow fragments the source vents themselves were in deep water; that during deposition of hyaloclastites they were in shallow water, water-soaked ground or adjacent to shorelines, and that during formation of lava flows of the shield volcanoes, they were and are above water. Moore and Fisk quote Kjartansson (1966, p. 34) for suggesting that on Iceland and Surtsey the change from predominantly pillow lava below to clastic volcanics above has occurred no deeper than 20 to 30 m. Saemundsson (1967) and Jones (1969) have also discussed sequences of subaqueous pillow lavas and pillow breccias capped by subaerial flows in Iceland. In one well-documented sequence, Jones estimates transition from close-packed pillow lava to breccia to have occurred at about 100 m below the presumed water surface. Higgins (1971) has suggested that a transition from basal close-packed pillow lavas through pillow breccias and aquagene tuffs capped by probable lava flows in metabasalts of Maryland may represent the surfaceward building of a lava pile on the flanks of a volcano. On the basis of size and abundance of vesicles and curves derived by Jones (1969), Higgins reasons that close-packed pillows were emplaced at depths between 100 and 400 m, but mostly at depths shallower than 300 m and, in some cases, possibly in water as shallow as a few tens of meters.

It remains to be shown whether in the Karmutsen pillowed basalts, which have undergone low-grade metamorphism to the prehnite–pumpellyite facies, the size and abundance of amygdules can be used as indices of depth of emplacement in the manner described by Moore (1965) and Jones (1969). Commonly, small regions outside amygdules have been partly or entirely converted to the same minerals as those inside the amygdules. For what it is worth, the diameters of subspherical amygdules in uppermost lower Karmutsen pillows and in upper Karmutsen pillows, are commonly as large as 0.5 cm (Fig. 5*a*), which is considerably larger than diameters of vesicles reported by Jones for depths of 100 m or less. Subspherical amygdules in flows are up to twice this size.

Rates of accumulation of pillow lava, pillow breccia, and flows, porosity of pillowed volcanic rocks

In the examples of emergent basalt sequences provided by the creation, filling and obliteration of intravolcanic basins in the upper Karmutsen, an explanation of the phenomenon based simply on local variations in subsidence rates alone is not convincing; for one thing, because in every case it requires first either nondeposition of basalt or accelerated subsidence of a very small area within a thick and relatively rigid basalt pile, followed by reduced subsidence in precisely that same small area and always within a relatively short time interval. In reality the conformable structure of the sediment–pillow assemblages and the lack of any evidence of folding or faulting, suggest that subsidence of the basins and subsidence of the neighboring basalt terrane was exactly coincident. Differences in the rate of supply of lava is not an attractive explanation either, since there is nothing to suggest that separate vents were located fortuitously within the basins, which would then supply an excess of lava to form pillows, and thus to fill them to the level of surrounding flows. On the con-


0 5 cm
a
b
0 1 m
c

trary, there are occurrences in which pillowed rocks located within the basinal assemblages can be observed to grade into flows outside, indicating a common supply of lava for pillows and for flows. Two alternative explanations are available. The first is that the extremely gentle slopes on the basin flanks would have promoted some downslope aggregation of pillows and especially of pillow breccias and aquagene tuff within the basins (viz. Carlisle 1963, p. 67–70). The branching tongues of budding pillows (Fig. 5*b*) and the slump and turbidite structures of pillow breccias give support to this postulate, but the gentleness of the basin slopes suggests that it may not be an adequate explanation in itself.

A second explanation, which is perhaps worthy of much wider attention, is that, given the area of accumulation and the rate of magma supply, pillows and also pillow breccias and aquagene tuffs, or hyaloclastites in general, will pile up more rapidly than flows simply because water-filled interstices are left between individual pillows (Fig. 5*c*) and between the shards and granules of aquagene tuffs. The 'porosity' of most close-packed pillow assemblages is appreciably less than that of rhombohedrally-packed spheres—26%—because the warm pillows conform somewhat to underlying surfaces, especially if the pillows are large, and also because some spalled pillow rim commonly settles in the interstices. Point counts across some lower Karmutsen and upper Karmutsen close-packed pillow lavas yielded 'porosites' of 6.7 and 9.0%, respectively. An additional internal porosity is developed in pillows as compared with flows because the quenched subspherical rims have sufficient strength to resist thermal contraction from cooling of the pillow core and at the same time they inhibit the escape of the volatiles. Some pillows crack and allow liquid lava to drain away. The initial porosity of aquagene tuff, prior to compaction or diagenesis is, at a guess, probably in the range of sand porosity —35 to 40%, but highly variable, rough, hackly surfaces on some fragments would tend to inhibit compaction and increase porosity, absence of sorting and abundance of extremely fine glass in some tuffs would reduce it. Hay and Iijima (1969, p. 353) report porosities of 33.1 to 39.4% for fresh tuff, from 21.9 to 30.5% for porous analcimitic, palagonitic, or moderately analcimitic tuff and from 8.2 to 17.1% for nonporous to dense palagonitic and highly zeolitic tuff in the tephras of Oahu. Allowing for penecontemporaneous compaction and diagenesis, a porosity estimate in the order of 15% for Karmutsen pillow breccias during basinal filling appears reasonable. On the basis of differential porosities alone, *i.e.* with a given rate of magma supply and neglecting possible downslope aggregation, then the relative rates of accumulation of close-packed pillows, pillow breccia, and flows would be roughly in the proportions 1.08 to 1.15 to 1 from which it follows that a basin 10 ft (3 m) deep would be filled to the same level as neighboring flows after the accumulation, for example, of 11.7 ft (3.6 m) of close-packed pillows and 70.4 ft (21.4 m) of pillow breccia. The greatest known thickness of pillowed rocks in the upper Karmutsen is in the range 600 to 700 ft (183–214 m), which would be sufficient, according to this line of reasoning, to fill a basin some 80 to 100 ft (24–30 m) deep, depending on the proportions of pillows and pillow breccia. Allowing for the fact that downslope aggregation would allow somewhat deeper basins to be filled the concept of differential rates of accumulation due to differential porosities, seems to accord rather well with observed dimensions and lithologies of upper Karmutsen intravolcanic basins.

In an attempt to estimate actual accumulation rates, one might apply the concept of differential porosities to the whole Karmutsen Subgroup which, as shown earlier, represents some 2.5 to 3.5 m.y. of accumulation at an average rate of 0.17 to 0.27 cm yr^{-1}. In terms of relative stratigraphic thicknesses (Fig. 1) the lower (pillow lava) middle (pillow breccia), and upper (flow) units represent 41, 13, and 46%, respectively, but in terms of mass of

FIG. 5. Karmutsen volcanic rocks. (*a*) Amygdules in a pillow of an isolated-pillow breccia, upper Karmutsen Subgroup, Quadra Island. (*b*) Incipient downslope budding of pillows in a close-packed assemblage, upper Karmutsen Subgroup, Hyacinthe Bay, Quadra Island. (*c*) 'Porosity' of close-packed pillow lava; quartz-filled interstices (*e.g.*, arrow) between pillows in lower Karmutsen Subgroup, northeastern Vancouver Island.

magma emplaced, considering porosity, they represent 40, 12, and 48% respectively. Therefore, if, as a first approximation, one assumes an approximately uniform rate of magma supply throughout Karmutsen eruption, the durations of the three parts would be 1.00 to 1.40 m.y., 0.30 to 0.42 m.y. and 1.20 to 1.68 m.y. and the rates of accumulation would be 0.19 to 0.26, 0.20 to 0.28, and 0.17 to 0.24 cm yr^{-1}, respectively, or more simply 0.22 for pillow lava, 0.24 for pillow breccia, and 0.20 cm yr^{-1} for flows, each ±20%. The assumption of a uniform magma supply is unverifiable, of course, from present data; eruption may have waxed and waned in a variety of patterns. Or the slightly shorter magma column during submarine eruption of pillows as contrasted with subaerial eruption of flows may have enhanced magma supply leading to further increase in the stratigraphic rate of pillow accumulation. In essence, however, the assumption simply modifies the aggregate average accumulation rate of 0.17 to 0.27 cm yr^{-1} of the whole Karmutsen to accord with the differential porosities of its parts.[3]

Because shallow-water carbonates in the intravolcanic beds recur throughout much of the upper Karmutsen over northern and central Vancouver Island, regional subsidence during upper Karmutsen eruption can be assumed to have been more or less continuous and, in a broad way, in balance with the accumulations of flows, which is to say at an average rate of approximately 0.20 cm yr^{-1}.

Depths of deposition in intravolcanic basins and in the larger seaway

From the conformable geometry, we have already been led to conclude that each intravolcanic basin was an integral part of the upper Karmutsen, subsiding no more nor less rapidly than neighboring flows. From the fact that the occurrences extend only a few tens or hundreds of feet along strike, we are also led to believe that the basins were probably never very deep, even during deposition of the *Halobia* lime-shale facies for if the adjacent flows had subsided to appreciable depths, say several hundreds of feet, at any time, sedimentation of some sort would have occurred over a very much wider area than it has. Independent evidence for shallow deposition, even of the lime-shale facies, can be drawn from a comparison of the estimates for sedimentary accumulation rates and the upper Karmutsen subsidence rate. If the bioclastic limestone of microfacies 1 accumulated at a rate even grossly similar to that of the Quatsino Limestone, from much of which it is lithologically indistinguishable, the time required for its deposition prior to the beginning of *Halobia* lime-shale sedimentation is not very long and the increase in depth due to regional subsidence during its deposition, therefore, cannot have been very great. A bed of microfacies 1 limestone, 1 ft (30.5 cm) thick, for example, at the base of an intravolcanic occurrence would represent only some 500 y of sedimentation (30.5 cm at 0.06 cm yr^{-1}, uncorrected for postdepositional thinning) and taking regional subsidence at 0.20 cm yr^{-1}, which is the average rate estimated above for the upper Karmutsen Subgroup, the increase in depth of water above the surface of sedimentation at the end of that period would have been only about 2.3 ft (0.20×500 less $30.5 = 70.5$ cm). The subsidence rate undoubtedly varied over short periods, but even for unusually thick occurrences of microfacies 1 limestone, the increase in depth of water, under the stated assumptions, would have been only a very few tens of feet at the most. Equivalent calculations for the *Halobia* lime-shale beds indicate larger increases in depth during their deposition though it must be remembered that occurrences of this microfacies are typically only a fraction of an inch to a very few feet thick. Thus, assuming subsidence to have continued uniformly at the same rate and accumu-

[3]Estimating the length and width of the Karmutsen basalt assemblage conservatively as 400 km and 150 km (assuming Queen Charlotte Islands to have been adjacent to northern Vancouver Island during extrusion), the volume of basalt (3.7 to 4×10^5 km^3) and the rate of production of basalt (100 to 160 $\times$ 10^6 m^3 yr^{-1}) are of the same order of magnitude as those of the Columbia River Basalt and related volcanics (Baksi and Watkins 1973). The rate is about three to four times the estimated rate of production on volcanic islands, such as Iceland, Hawaii and the Azores, and six to twelve times the estimated rate in a typical spreading ridge system 400 km in length. High production rates would logically be associated with emergent basalts.

lation at the aggregate rate derived for the lithologically similar lower Parson Bay Formation, the time required for deposition of 1 ft of this microfacies (30.5 cm at 0.0003 cm yr^{-1}, uncorrected for postdepositional thinning) would be about 10 000 yr and the increase in depth of water during this period ($0.20 \times 10\,000$ less $30.5 = 1970.5$ cm) would be about 65 ft. As noted above, individual beds or thin sequences of this microfacies such as those in intravolcanic beds may have accumulated more rapidly than the lower Parson Bay as a whole, with consequent reduction in the estimated depth. Feldspathic wacke of microfacies 3, in those occurrences where it has been found, is at the top of some 3 to 20 ft (1 to 6 m) of microfacies 1 and 2 beds. Applying the same kind of reasoning, the depth of water when sedimentation of this microfacies began might have been as great as 100 to 675 ft (30–300 m). The evidence for turbid flow on a very small scale, mainly the presence of grain gradation and some load structures in microfacies 3 beds, does suggest that appreciable submarine relief had developed. On the other hand, microfacies 3 beds are commonly interlayered with lesser amounts of microfacies 2 apparently *in situ* and, in at least one instance with thin beds of microfacies 1. If the feldspathic wacke of microfacies 3 is indeed derived from disintegration of basalt then its abundance is a function of that process too.

Figure 6, reconstructing the development of one of the thicker upper Karmutsen sediment–pillow assemblages illustrates diagrammatically how the several estimates of rate and depth, deduced separately for volcanic and sedimentary rocks, form a consistent model. In some sediment–pillow assemblages, the sediments are much thinner than those shown, or not observed at all, perhaps because the present erosion surface does not intersect central parts of the basin; in others the sediments are thicker or the volcanic facies are in a different proportion. Nevertheless, the internal consistency of the model lends support to the notion that transitions from bioclastic limestone to *Halobia*-rich siliceous lime-shale and, in some cases, to feldspathic wacke can take place at depths of only a few feet to a few tens of feet below sea level. It also follows from the model that the depth at which the accumulation of pillows began in the intravolcanic basins is considerably less than the maximum thickness of the pillowed assemblage. For pillowed assemblages 600 to 700 ft (180–215 m) thick this depth is probably in the range of 100 to 200 ft (30–60 m). This is also the maximum depth of the sediment-water interface during deposition of *Halobia* lime-shale or feldspathic wacke. Although pillow breccias and aquagene tuffs accumulate down to essentially these depths also, because of downslope mass transfer, their origin is apparently related to shoaling, and, in the case of broken-pillow breccia, to the length of time that pillows and granulated basalt have been stacking up in unstable piles. On the other hand, since in many occurrences, close-packed pillows have come to rest on newly deposited coarse biogenetic limestone, they too can form at depths of only a few feet or tens of feet. Amygdule diameters, if reliable indicators of depth, reinforce these conclusions.

The sequence close-packed pillow lava to pillow breccia to flows represents emergent deposition and is to be widely expected.

The conclusion that intravolcanic *Halobia* lime-shale is a shallow-water epipelagic deposit seems equally applicable to the Parson Bay Formation especially in view of the remarkable textural, structural and compositional similarities. It may also apply to the *Daonella*-bearing pre-Karmutsen shales although the data on these are too sparse for certainty. The pre-Karmutsen Buttle Lake Formation is as clearly of shallow-water origin as is the Quatsino Limestone. The great thicknesses of these unquestionably shallow-water units imply, however, that subsidence rates must have been much lower during their deposition than during accumulation of the Karmutsen Subgroup. Large subsidence rates are uniquely associated with Karmutsen volcanism suggesting that they resulted from the outpouring of basalt magma itself. A rapid acceleration of subsidence during the early stages of volcanism would explain the exceptional thicknesses of lower Karmutsen pillow lavas and middle Karmutsen pillow breccias. A gradual deceleration of subsidence toward the end would explain the thinness of upper Karmutsen pillow assemblages and

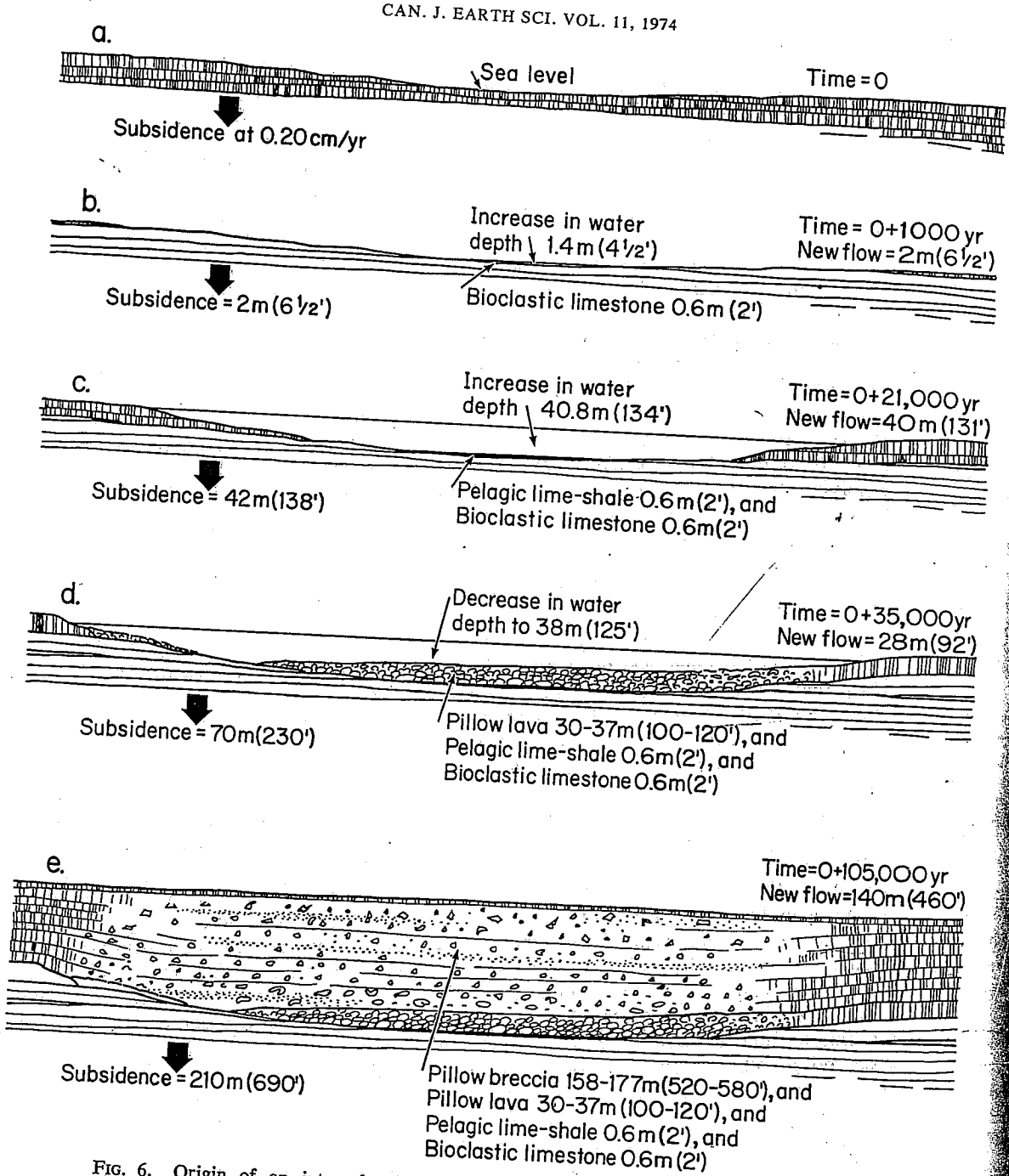


FIG. 6. Origin of an intravolcanic sediment–pillow assemblage. Horizontal scale foreshortened and unspecified. Rates of accumulation discussed in text. (*a*) Time 0. Incipient basin is formed through uneven distribution of lava. Regional subsidence is assumed to continue at 0.20 cm yr^{-1}. (*b*) One thousand years later, new flows roughly 2 m thick having failed to reach the basin, carbonate-forming benthos have induced the accumulation of approximately 0.6 m of bioclastic limestone (microfacies 1). Water depth has reached 1.4 m. (*c*) An additional 20 000 years later, benthonic life has been largely destroyed and roughly

would match the gradual waning of volcanism manifest by the increasing frequency of intravolcanic basins.

Conclusions

The Triassic shallow-water siliceous lime-shale facies of Vancouver Island has most of the characteristics commonly attributed to deposition in an anaerobic or near-anaerobic, but not turbid, environment including diagenetic pyrite, an excess of carbon, near absence of benthos, fine nongraded lamination and textural evidence for very slow deposition in a low energy environment. Explanations of why this passive accumulation should have taken the place of vigorous biogenic carbonate deposition in a shallow-water environment based solely on geomorphological changes such as partial closure of a marine embayment, creation of a wind break or current barrier or of a submarine bench extensive enough to dissipate the energy of waves and thus to reduce oxygenation are not compelling because the intravolcanic basins were even more constricted during deposition of the coarse bioclastic facies than during accumulation of the *Halobia* shale. Moreover, the shelf upon which the Parson Bay *Halobia* shale was deposited was no more extensive than that formed by the flat-lying basalt flows upon which the Quatsino Limestone formed. In each case, topography was subdued making it likely that wave action and circulation were not particularly strong over the whole area even during deposition of the very shallow-water bioclastic limestone. At very shallow depths oxygenation can be adequate for a wide range of biota with relatively little agitation of the water. Under these conditions, mere increase in depth of a few feet or tens of feet might be sufficient to explain the change from bioclastic limestone to epipelagic lime-shale.

Equally as characteristic as increasing depth, however, is the appearance and increasing abundance of basaltic plagioclase. The black impalpable siliceous and aluminous matrix, which cements the feldspathic wacke and which is also a major component of the underlying *Halobia* or *Monotis* shale, is without much doubt derived in large part from the glassy and other nonfeldspar components of basalt either through subaerial weathering or through granulation and subsequent decomposition of the lava as it flows into water. Subaerial ash falls are a less probable source in a basaltic terrane. Regional distribution of lithofacies and thicknesses, reported earlier in the paper, suggest that Karmutsen volcanism continued longer in the northwest, probably evolving into Bonanza volcanism on more remote islands. A variety of dissolved ions, similarly derived from the volcanics must have been dispersed throughout each intravolcanic basin and throughout the epicontinental seaways. Among these would be abundant silicon as well as phosphorus and several nutrient trace elements, but also some, which in minor or large amounts, would be toxic. The concentrations of all of these would build up rapidly in intravolcanic basins, less rapidly in the epicontinental seaways. Tappan (1971) has given a comprehensive discussion of the complexities of succession and rejuvenation in large and small ecosystems and of the effects of changes in the physical, chemical, and biological environment with particular attention to the supply of nutrients. She describes how an influx of nutrients, as in the eutrophication of a freshwater lake, may bring about the sudden dominance of one or a few species, a particu-

0.6 m of *Halobia*-rich epipelagic siliceous lime-shale (microfacies 2) has accumulated. Depending upon the extent and times of arrival of new basalt flows, some pillows or pillow breccia may have formed during this interval or, alternatively, prolonged weathering of exposed basalt may have yielded occasional floods of feldspathic wacke (microfacies 3). Depth of water in the basin has reached its maximum of 40.8 m. (*d*) Influx of basalt lava at the end of stage (*c*) resulted in submarine budding of pillows and their transport down slope directly on to pristine sediment. During 14 000 years roughly 30 to 37 m of close-packed pillows have accumulated and water depth in the basin has been reduced to 38 m. (*e*) Lava flowing into progressively shallower water during the final 70 000 years formed granules, globules, and whole pillows ranging widely in size, many of which were subjected to further disaggregation during slump or turbid flow. The basin is filled to sea level by 158 to 177 m of pillow breccia and aquagene tuff and the assemblage is succeeded by massive subaerial flows.

lar alga for example, at the expense of all others. "Some marine plankton may be toxic, particularly certain blue-green and dinoflagellate species, or so abundant as to deplete oxygen supplies and leave the bottom anaerobic." (Tappan 1971, p. 1065). A possible application of similar ideas to the present case is attractive.

Interesting as it may be, two observations seriously weaken such a chemical explanation. At the very base of intravolcanic beds and also of the Quatsino Limestone one commonly finds minor amounts of basaltic impurities; yet coarse bioclastic carbonate has accumulated here. In several places tongues of pillow lava and pillow breccia up to 100 ft (30 m) thick are intercalated within Quatsino Limestone without apparent destruction of the carbonate ecosystem.

The widely varying periodicity of Triassic volcanic-sediment environments on Vancouver Island seems to have resulted primarily from physical relationships involving mainly the rates of subsidence, volcanism, and organic and inorganic accumulation.

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BAKSI, A. K. and WATKINS, N. D. 1973. Volcanic production rates: comparison of oceanic ridges, islands, and the Columbia Plateau Basalts. Science, **180**, pp. 493–496.

CARLISLE, D. 1963. Pillow breccias and their aquagene tuffs, Quadra Island, British Columbia. J. Geol., **71**, pp. 48–71.

——— 1972. Bute Inlet map-area, Vancouver Island, British Columbia (92K). *In*: Geol. Surv. Can., Pap. 72–1, Pt A, pp. 21–23.

CARLISLE, D. and SUSUKI, T. 1965. Structure, stratigraphy and paleontology of an Upper Triassic section on the west coast of British Columbia. Can. J. Earth Sci., **2**, pp. 442–486.

CAROZZI, A. V. and FROST, S. H. 1966. Turbidites in dolomitized flank beds of Niagaran (Silurian) reef, Lapel, Indiana. J. Sed. Petrol., **36**, pp. 563–575.

CLOUD, P. E., JR. 1962. Environment of calcium carbonate deposition west of Andros Island, Bahamas. U.S. Geol. Surv., Prof. Pap. 350, 138 p.

CRICKMAY, C. H. 1928. The stratigraphy of Parson Bay, British Columbia. Univ. Calif. Publ., Bull. Dept. Geol. Sci., **18**, pp. 51–70.

DAWSON, G. M. 1887. Report on a geological examination of the northern part of Vancouver Island and adjacent coasts. *In*: Geol. Surv. Can., Annual Report 2 (1886), Pt B, pp. 1–129.

DOLMAGE, V. 1919. Quatsino Sound and certain mineral deposits of the west coast of Vancouver Island. Geol. Surv. Can., Summ. Report (1918), Pt B, pp. 30–38.

DZULYNSKI, S. and WALTON, E. K. 1965. Sedimentary features of flysch and greywackes. Elsevier, Amsterdam. 274 p.

FISCHER, A. G. 1964. The Lofer Cyclothems of the Alpine Triassic. *In*: Symposium on Cyclic Sedimentation, D. F. Merriam (*Ed.*), Kansas Geol. Surv., Bull. **169**, pp. 107–149.

GIVENS, C. R. and SUSUKI, T. 1964. Late Triassic fauna from interlava sediments of east-central Vancouver Island. (Abstr.). Geol. Soc. Am., Spec. Pap. 76, p. 203.

GUNNING, H. C. 1933. Zeballos River Area, Vancouver Island, British Columbia. Geol. Surv. Can., Summ. Report (1932), Pt A II, pp. 29–50.

HARLAND, W. B., SMITH, A. G., and WILCOCK, B. (*Eds.*). 1964. The Phanerozoic Time-Scale. Quart. J. Geol. Soc. London, 120S.

HAY, R. L. and IIJIMA, A. 1969. Nature and origin of palagonite tuffs of the Honolulu Group on Oahu, Hawaii. *In*: Studies in volcanology, R. R. Coats et al., (*Eds.*), Geol. Soc. Am. Mem. 116, pp. 331–375.

HIGGINS, M. W. 1971. Depth of emplacement of James Run Formation pillow basalts, and the depth of deposition of part of the Wissahickon Formation, Appalachian Piedmont, Maryland. Am. J. Sci., **271**, pp. 321–332.

HOADLEY, J. W. 1953. Geology and mineral deposits of the Zebellos–Nimpkish area, Vancouver Island, British Columbia. Geol. Surv. Can., Mem. 272.

JELETZKY, J. A. 1950. Stratigraphy of the west coast of Vancouver Island between Kyuquot Sound and Esperanza Inlet, British Columbia. Geol. Surv. Can., Pap. 50–37.

——— 1970. Some salient features of Early Mesozoic history of Insular Tectonic Belt, western British Columbia. Geol. Surv. Can., Pap. 69–14.

JONES, J. G. 1969. Pillow lavas as depth indicators. Am. J. Sci., **267**, pp. 181–195.

KJARTANSSON, G. 1966. Stapakenningin og Surtsey (A

comparison of table mountains in Iceland and the volcanic island of Surtsey off the south coast of Iceland). Náttúrufraedingurinn **36**, pp. 1–34 (not examined).

KUNIYOSHI, S. 1971. Chemical composition of the Karmutsen volcanics, Vancouver Island, British Columbia. (Abstr.). Geol. Soc. Am., Abstr. 84th Ann. Meet., Washington, D.C., Nov. 1971, p. 628.

MATHEWS, W. H. and MCCAMMON, J. W. 1957. Calcareous deposits of southwestern British Columbia. British Columbia Dep. Mines, Bull. No. **40**.

MOORE, J. G. 1965. Petrology of deep-sea basalt near Hawaii. Am. J. Sci. **263**, pp. 40–52.

MOORE, J. G. and FISKE, R. S. 1969. Volcanic substructure inferred from dredge samples and ocean-bottom photographs, Hawaii. Bull. Geol. Soc. Am., **80**, pp. 1191–1202.

MULLER, J. E. 1972. The Insular Belt. *In*: The Geology of the Southern Canadian Cordillera, 24th Int. Geol. Congr., Guidebook for Excursions A03-C03, p. 74.

MULLER, J. E. and CARSON, D. J. T. 1969. Geology and mineral deposits of Alberni map-area, British Columbia (92F). Geol. Surv. Can., Pap. 68–50.

MULLER, J. E. and RAHMANI, R. A. 1970. Upper Triassic sediments of northern Vancouver Island. *In*: Geol. Surv. Can., Pap. 70–1, Pt B, pp. 11–18.

SAEMUNDSSON, K. 1967. Vulkanismus und tektonik des Hengill-Gebietes in Südwest-Island. Acta Naturalia Islandica **2**, No. 7, pp. 1–105.

SILBERLING, N. J. 1961. Upper Triassic marine mollusks from the Natchez Pass Formation in northwestern Nevada. J. Paleontol. **35**, pp. 535–542.

SILBERLING, N. J. and TOZER, E. T. 1968. Biostratigraphic classification of the marine Triassic in North America. Geol. Soc. Am., Spec. Pap. 110.

SURDAM, R. C., SUSUKI, T., and CARLISLE, D. 1964. Upper Triassic section on Iron River, Vancouver Island, British Columbia. (Abstr.). Geol. Soc. Am., Spec. Pap. 76, p. 226.

SURDAM, R. C. 1968. The stratigraphy and volcanic history of the Karmutsen Group, Vancouver Island, British Columbia. Contributions to Geol. Univ. Wyoming **7**, No. 1, pp. 15–27.

SUTHERLAND BROWN, A. 1968. Geology of the Queen Charlotte Islands, British Columbia. British Columbia Dep. Mines, Bull. **54**.

TAPPAN, H. 1971. Microplankton, ecological succession and evolution. Proceed. N. Am. Paleontol. Convention, Sept. 1969, Pt H, pp. 1058–1103.

TOZER, E. T. 1967. A standard for Triassic time. Geol. Surv. Can., Bull. 156.

——— 1971. Triassic time and ammonoides: problems and proposals. Can. J. Earth Sci., **8**, pp. 989–1031.

Appendix A

Dilleri and Welleri Zone Faunal Assemblages on Vancouver Island and Vicinity

(See Fig. 4 for locations. Some changes in nomenclature will result from a revision of the Tropitidae and Juvavitinae proposed by Tozer 1971.)

(1) Open Bay, Quadra Island (Lat. 50°8′ 25″ to 35″, Long. 125°11′50″ to 12′30″) Dilleri Zone: Intensely folded and marbleized, light and dark layered upper Quatsino Limestone. The estimated unfolded exposed thickness is 100 to a few hundred ft (30 m +), but the total Quatsino thickness is unknown (Ref. Carlisle and Susuki 1964). Fauna: *Spirogmoceras* sp., *Californites* cf. *C. careyi* (Smith), *Traskites* (*Shastites*) cf. *T.* (*S.*) *compressus* (Hyatt & Smith), *Traskites* (*Stantonites*) cf. *T.* (*S.*) *rugosus* (Hyatt & Smith), *Hannaoceras* sp., *Tropites dilleri* Smith, *Tropites* cf. *T. dilleri* Smith, *Tropites* cf. *T. welleri* Smith, *Tropites* sp., *Discotropites* cf. *D. sandlingensis* (Hauer), *Discotropites* sp., *Paratropites sellai* (Mojsisovics), *Gymnotropites* cf. *G. americanus* Hyatt & Smith, *Tornquistites* (?), sp., *Bacchites* cf. *B. bacchus* (Mojsisovics), *Leconteiceras* sp., *Sagenites* (*Trachysagenites*) *herbichi* Mojsisovics, *Arcestes* sp.

(2) Iron River Dilleri Zone (Lat. 49°50′15″, Long. 125°29′40″): Light and dark layered upper Quatsino Limestone. The minimum measured thickness, in faulted blocks, to underlying Karmutsen is 325 to 375 ft (99–114 m). Fauna: *Torpites dilleri* Smith, *Arcestes* sp., *Halobia* sp.

Welleri Zone inferred (Lat. 49°50′50″, Long. 125°29′20″): Black laminated lenticular limestone layers within black lime-shale of the lower Parson Bay Formation 90 ft (27 m) above Dilleri Zone fauna occurrence. Fauna: *Discotropites* sp., *Halobia* sp.

(3) Patterson Creek (Lat. 50°0′45″, Long. 125°38′57″): Dilleri Zone: In 10 to 30 ft (3–9 m) of faulted upper Quatsino Limestone not less than 550 ft (168 m) above Karmutsen flows. Fauna: *Tropites dilleri* Smith, *Paratropites sellai* Mojsisovics, *Gymnotropites* (?) sp. indet., *Hannaoceras* sp., *Halobia* cf. *H. superba* Mojsisovics.

(4) Christensen Point – 1.25 mi (2 km) east. (Lat. 50°50′10″, Long. 128°10′50″) Dilleri Zone inferred: In light and dark layered upper Quatsino Limestone with minor sills above light gray thick-bedded, faulted Quatsino Limestone and not less than 600 ft (180 m) above Karmutsen flows. Fauna: *Tropites* cf. *T. dilleri* Smith, *Halobia* sp.

Welleri Zone inferred: In laminated dark limestone and lime-shales of uppermost Quatsino Limestone with minor sills approximately 30 ft (9 m) stratigraphically above Dilleri

Zone fauna occurrence and immediately below Parson Bay beds. Fauna. *Juvavites* cf. *J. strongi* Smith, *Hannaoceras* sp., Arcestids, *Halobia* sp.

(5) Southeast of Kains Lake (Lat. 50°41′36″, Long. 127°40′20″) Dilleri Zone inferred: Middle Quatsino Limestone with not less than 300 ft (92 m) of limestone exposed, but not mapped, above Karmutsen flows. Faunas *Paratropites* cf. *P. sulcatus* (Calcara) of Silberling, 1961, *Septocardia* (?) sp.

(6) Dick Booth Creek (Lat. 50°40′50″, Long. 127°37′35″) Welleri Zone: Medium gray limestone beds 2 to 10 in. (5–25 cm) thick interbedded with black limeshale in a section 25 ft (7.5 m) thick transitional between upper Quatsino Limestone and lower Parson Bay Formation. The thickness of Quatsino is as at locality 5 above. Fauna in upper bed: *Tropites* cf. *T. welleri* Smith, *Tropites* sp., *Discotropites* sp., *Anatomites* sp., *Proarcestes* (?) sp., *Halobia* cf. *H. suberba* Mojsisovics. Fauna in lower bed: *Tropites* sp., *Gymnotropites* sp., *Sirenites* (?) sp., *Halobia* sp.

(7) Holberg Quarry (Lat. 50°40′40″, Long. 127°59′15″) Dilleri Zone inferred: Upper Quatsino Limestone; coarse bioclastic light gray beds with sparse black partings above thick-bedded limestone. Fauna: *Tropites* cf. *T. dilleri* Smith. (J. E. Muller has collected *Tropites* cf. *T. welleri* Smith from beds apparently higher in the section exposed in this same quarry which he also places in Quatsino Limestone. Personal communication 1970.)

(8) Goodspeed River (Lat. 50°41′45″, Long. 128°0′40″ west side of bridge on branch road west from logging road NE51) Welleri Zone inferred: In dark gray nonfissile lenticular limestone beds 1 to 14 in. (2–36 cm) thick with interbedded black lime-shale in a section 12 ft (3.7 m) thick transitional between Quatsino Limestone and Parson Bay Formation and immediately overlain by undoubted Parson Bay beds. Quatsino Limestone several hundred feet thick is exposed up stream. Fauna: *Tropites* sp., *Discotropites* sp., *Gymnotropites* sp., *Hannaoceras* cf. *H. nodifer* Hyatt & Smith, *Arcestes* sp., *Halobia* sp.

(9) Tsulten River tributary, Beaver Cove area (Lat. 50°29′, Long. 126°53′) Welleri Zone: Dark gray nonfissile limestone bed 10 to 12 in. (25–30 cm) thick and adjacent lens within black lime-shale of the lower Parson Bay Formation. The thickness of underlying Quatsino Limestone is uncertain because of faulting but 7 mi (11 km) to the southeast at Bonanza Lake not less than 850 ft (260 m) and possibly as much as 1750 ft (530 m) of Quatsino Limestone resting on Karmutsen flows has been mapped. Fauna: *Tropites* cf. *T. welleri* Smith, *Juvavites* cf. *J. knowltoni* Smith, *Tropites* sp., *Halobia* sp.

(10) Sproat Lake (MacMillan Bloedel logging road Branch 302) (Lat. 49°13′40″, Long. 124°58′10″) Dilleri Zone inferred: Sandy weathering calcarenite and skeletal calciruddite in lower part of upper Quatsino Limestone with sparse dark interbeds. The measured thickness of Quatsino to underlying Karmutsen is 650 ft (198 m) and to overlying Parson Bay Formation is 200 ft (61 m). Fauna: *Tropites* cf. *T. dilleri* Smith, *Arcestes* sp., *Halobia* sp.

Welleri Zone (Lat. 49°13′30″, Long. 124°58′10″): Medium to dark gray limestone lenses within black lime-shale of the lower Parson Bay Formation approximately 400 ft (120 m) above inferred Dilleri Zone beds. Fauna: *Tropites welleri* Smith, *Discotropites* cf. *D. smithi* A. Kutassy, *Anatomites* cf. *A. knowltoni* (Smith), *Anatomites* cf. *A. strongi* Smith, *Pamphogosirenites* cf. *P. pamphagus* (Dittmar), *Pinacoceras* sp., *Arcestes* sp., *Halobia* sp. (Muller and Carson 1969, p. 15, report on inferred Welleri Zone fauna from 'Quatsino Formation' south of Sproat Lake. The lithology is not given.)

(11) Klaskino Inlet (Lat. 50°10′, Long. 127°51′35″ south side small unnamed bay) Dilleri Zone: Uppermost Quatsino Limestone (see text). The Quatsino is faulted and intruded by dikes immediately down section from the fossilierous beds and is observable over a proven stratigraphic thickness of only some 50 ft (15 m) above Karmutsen flows.

Fauna in upper bed: *Tropites dilleri* Smith, *Paratropites sellai* Mojsisovics, *Discotropites* cf., *D. sandlingensis* Hauer, *Bacchites* cf. *B. bacchus* Mojsisovics, *Bacchites* cf. *B. sphaericus* Smith, *Arcestes* sp.

Fauna in middle bed: *Tropites dilleri* Smith,

Tropites sp., *Paratropites sellai* Mojsisovics, *Gymnotropites* sp., *Bacchites* cf. *B. bacchus* Mojsisovics, *Proclydonautilus* (?) sp.

Fauna in lower (main) bed: ***Tropites dilleri*** Smith, *Paratropites* cf. *P. sellai* Mojsisovics, *Paratropites* sp. *Paratropites* (?) sp. *Sagenites* (*Trachysagenites*) *herbichi* Mojsisovics, *Bacchites* cf. *B. bacchus* Mojsisovics.

Welleri Zone: The collection is from one of five main lenticular dark limestone beds with black *Halobia*-rich lime-shale referred by the present authors to the lowermost Parson Bay Formation (see text). Fauna: *Tropites welleri* Smith, *Tropites* sp., *Gymnotropites* cf. *G. californicus* Smith, *Discotropites* cf. *D. lineatus* Smith, *Discotropites* cf. *D. mojsvarensis* Smith, *Discotropites* sp., *Pamphagosirenites* cf. *P. pamphagus* (Dittmar), *Sagenites* (*Trachysagenites*) cf. *S.* (*T.*) *herbichi* Mojsisovics, *Juvavites* cf. *J. strongi* Smith, *Anamotites* cf. *A. knowltoni* (Smith), *Arcestes* sp., *Dieneria* (?), sp., *Halobia* sp.

PILLOW BRECCIAS AND THEIR AQUAGENE TUFFS, QUADRA ISLAND, BRITISH COLUMBIA[1]

DONALD CARLISLE[2]

ABSTRACT

Pillow breccias consist of whole or disaggregated volcanic pillows in an abundant matrix of cogenetic basic tuff. They form in water and are thought to be common rocks, confused at times with other volcanic breccias.

On Quadra Island, British Columbia, the breccias usually overlie ordinary pillow lava which rests in turn upon basalt flows or upon pelagic limestone. In the transition to isolated-pillow breccia the matrix increases gradationally upward to as much as 80 per cent; the pillows become isolated, diversiform, and ill-sorted. Some pillows were broken after settling but before compaction. The matrix, consisting of cogenetic globules, granules, shards, and basaltic fragments produced beneath water, is an aquagene tuff. Subaqueous shard formation is discussed. Broken-pillow breccia forming the top of the successions consists of unsorted pillow fragments set in similar aquagene tuff which is irregularly flow-laminated and somewhat resembles an ignimbrite. Laminated aquagene tuff beds with load casts in their upper surfaces occur within broken-pillow breccia.

Ordinary pillow lava accumulated in relatively clear water, isolated pillows in increasingly turbid, vapor-charged water. Increasing thicknesses of incoherent, steam-laden pillows and tuff led to instability, subaqueous slumping, and fragmentation.

INTRODUCTION

The literature on pillow structures in volcanic rocks goes back more than 130 years, and although there are many references to associations of pillows and tuffaceous rocks, almost all the discussions have emphasized the pillows themselves and close-packed accumulations of unbroken pillows. Textbook illustrations of pillow lavas invariably show the close-packed variety. Undoubtedly this is the way pillows typically occur. But pillowed volcanics in which a very large part or even the bulk of the rock is made up by a tuffaceous matrix or in which the pillows are not whole are rather common also. Possibly they are much more common than has been generally recognized. The Palagonite formation of Iceland is an example though an unusual one. These pillow-tuff rocks are not merely inclusions of pillows in a water-lain aerial pyroclastic; the tuff is intimately related to the pillows and is composed largely of basic glass usually identical with the pillow rims but altered in varying degrees. In some places these pillow-tuff rocks grade into rocks that are entirely of basic tuff.

Such rocks—in part pillow or pillow-derived and in part cogenetic basic tuff or lapilli tuff—are *pillow breccias*.[3] A study of them yields some interesting information on the genesis of pillow structure and the environment it represents. It throws some light also on the formation of volcanic tuff entirely beneath water—a rock type for which the term "aquagene tuff" is herein proposed.

A first impression of pillow breccias as one sees them in the field can be had from plate 1. These are fragmental rocks with apparent clasts of andesite-basalt of a variety of shapes, from a few inches to a few feet in diameter, set in a fine- to medium-grained matrix which may be of a different color than the large clasts but apparently also clastic itself. At least the matrix resembles a compact tuff or lapilli tuff in some expo-

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[3] The term "pillow breccia" apparently was used first in the present sense by Henderson (1953) for an occurrence in the Precambrian of the Yellowknife District with isolated unbroken pillows. It was again proposed in 1955 by Chapman for pillow-shaped fragments of diabase or gabbro in granitic material. The term had already been used by Henderson at the time of Chapman's proposal, however, and it would seem that some other equally descriptive term for rocks entirely unrelated to classical pillow lavas would be more suitable. Henderson's "pillow breccia" is equivalent to "isolated-pillow breccia" herein.