REPLY

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During the two decades since Chave et al. (1972) published their compilation of Holocene carbonate sedimentation rates, several other scientists have compiled similar data. Milliman (1974), Hay and Southam (1977), and Kinsey and Hopley (1991) all have compiled carbonate accumulation data and suggested that the probable calcium carbonate flux to shallow water is significantly higher than it would be if it were at equilibrium with known Holocene weathering rates. What is largely unavailable, a point that Mylroie addresses, are quantitative estimates of the differences between Holocene and late Pleistocene carbonate weathering rates.

Weathering-rate changes from Pleistocene to Holocene time are a topic of broad interest. The discovery of perturbations within oceanic Sr isotopic ratios recorded in planktonic foraminifera at these short time intervals will create additional interest in this issue (Dia et al., 1992). Certainly, any new data concerning changes in weathering rates from the last glacial into the Holocene, or new approaches for obtaining these data, will be welcomed by those of us engaged in carbon cycle modeling.

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Postcollisional extension of the Caledonide orogen in Scandinavia: Structural expressions and tectonic significance: Comment and Reply

COMMENT

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Fossen and Rykkelid (1992) extrapolated well-documented late Caledonian crustal extension in southwest Norway nearly 800 km northward into the Ofoten region, where such structures are only now beginning to be recognized. We strongly concur that the magnitude and style of postcontractional extension in the Caledonides is a significant question, and Fossen and Rykkelid have suggested a provocative, but speculative, interpretation of Ofoten. Fossen and Rykkelid (1992) presented few new data from Ofoten, however, and they scarcely acknowledged the sizable Ofoten literature base, leading us to question the validity of their interpretation. Herein, we comment on how aspects of their map, cross sections, and structural-metamorphic interpretations of Ofoten conflict with published data.

In Ofoten, detailed 1:20,000-scale structural studies along the western margin of the Rombak window (Hodges, 1985; Crowley, 1989) have documented a pervasive top-to-the-east thrust zone but have failed to recognize evidence of an "~1-km-thick west-dipping zone of mylonites with abundant down-to-the-west asymmetric structures" (Fossen and Rykkelid, 1992, p. 739). These and other workers (e.g., Steltenpohl and Bartley, 1988; Barker, 1989) developed a detailed, regionally consistent deformational and metamorphic sequence that does not readily conform to Fossen and Rykkelid's suggested regional D1 contraction followed by regional D2 extension.

Cashman (1990) first reported detailed observations on steep, west-dipping, top-to-the-west, extensional faults within the Rombak window, which were interpreted to record failure caused by flexural loading by eastward-advancing thrust sheets. Steltenpohl and Bartley (1988) documented late top-to-the-west transport in Ofoten as a regionally significant event, characterized by east-dipping, top-to-the-west ductile shear zones and meso-scale to regional-scale west-vergent back-folds. Our detailed structural and fabric analysis of the back-folds indicates that they are periodic structures with steep east-dipping axial surfaces that probably formed in layer-parallel shortening. The regional Ofoten synform itself is a major back-fold, and parasitic, smaller scale folds (2–3 km in amplitude) are prevalent along the western limb and die out eastward (Steltenpohl and Bartley, 1988), a pattern exactly opposite of Fossen and Rykkelid's (1992, Fig. 2, B and C) depiction. The observed back-folds thus do not resemble the ~5–10 km amplitude, recumbent, west-directed folds depicted in Fossen and Rykkelid's (1992) Figure 2, which was drawn through the same area we investigated. We also bracketed the absolute timing of late-phase folds in Ofoten between the Early Silurian and Late Devonian (Steltenpohl and Bartley, 1988). This age range led us to suggest that the late-phase folds in Ofoten were related to the event that formed the Devonian basins in southwest Norway, as Fossen and Rykkelid (1992) have done, although we considered both the basins and the folds to have formed in transpression rather than extension.

Metamorphism in this region is very complex, varying vertically through the nappe stack as well as laterally across the axial trace of the Ofoten synform (Barker, 1989; Steltenpohl et al., 1990). Late-stage, retrogressive, greenschist-facies assemblages are developed at almost all tectonostratigraphic levels and are concentrated along thrust zones (Barker, 1989). These assemblages, therefore, do not bear a simple relation to a hypothetical single, large, extensional fault along the western margin of the Rombak window.

Specific aspects of Fossen and Rykkelid's interpretation, therefore, are at odds with the published literature. We agree, however, that late top-to-the-west extension may be a significant factor in the structural evolution of the Ofoten region. We hope that future interpretations of this event in Ofoten will better consider what is already known about the geology of this area.

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REPLY

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Steltenpohl and Bartley may be right in saying that we should have acknowledged more of the existing literature from Ofoten. However, this is covered in more extensive papers and theses (Rykkelid, 1992; Rykkelid and Andresen, 1993, unpublished) that also give better documentation and discussion of new data. Our paper was a summary of several years of field work and research, utilizing kinematic analysis for areas where more "old-fashioned" methods—i.e., separation of deformation into a large number (4 to 6 in Ofoten) of distinct phases and correlation of folds and cleavages on a regional scale—had been applied. It is becoming gradually clearer to geologists that this traditional way of categorizing geologic structures is not conformable with the dynamic nature of orogenic processes, and the growing amount of literature that focuses on understanding deformation in terms of kinematics illustrates that a kinematic approach may in fact be more useful in understanding the structural dynamics of orogenic belts.

Kinematic indicators have generally been disregarded in previous studies, perhaps because of the traditional assumption that almost all Paleozoic ductile deformation in the Caledonides is related to thrusting. This may explain why previous authors overlooked most of the top-to-the-west postcontractional ductile shear along the basal thrust zone, as well as the 1-km-thick extensional Ofoten shear zone. Widespread back-folds (F6 in Hodges, 1985; F4 in Steltenpohl and Bartley, 1988; F5 in Crowley, 1989) were, however, related to west-directed shear by Steltenpohl and Bartley (1988), although this was never explained or discussed in much detail. Steltenpohl and Bartley (1988) appear to have related these folds to Caledonian deformation and to slightly older "but nearly coeval crossfolds" that interfere to give rise to the gneiss domes in the Ofoten area and possibly along the entire length of the Norwegian Caledonides. They further suggested that these gneiss domes are the result of Caledonian *compression* and that the late folds are the product of late Caledonian sinistral transpression. The latter interpretation is, of course, purely speculative, because no evidence for sinistral shear along northeast-trending shear zones has been reported.

They also suggested a correlation between these cross-folds and sinistral movement in the Devonian basins of southwestern Norway, completely disregarding the voluminous recent literature that documents the extensional nature of these basins. The cross-folds may equally well be explained by other mechanisms, such as southward transport of orogenic matter caused by isostatic-gravitational instabilities within the orogenic wedge or large-scale differential movements within the nappes during emplacement (e.g., Ridley, 1986; see also Romer and Bax, 1992). We argue that the back-folds are incompatible with Caledonian compression but fit very well with our observations of significant west-directed backsliding, which we relate to regional postcompressional extensional tectonics. We also believe that the largest of the basement windows in the Scandinavian Caledonides, i.e., the Western Gneiss region, is related to gently west dipping extensional shear zones and certainly did not form by fold interference. A similar explanation may in part explain the basement windows in the Ofoten area.

The back-folds in the Ofoten synform were enlarged on our profile to make the main geometry visible, and we agree that the actual folds are considerably smaller. However, the important point is the presence of abundant folds in the Ofoten synform, a situation that was reproduced in clay experiments where simultaneous back movement of Caledonian nappes above a developing synthetic extensional basement shear zone was modeled (cf. Fossen and Rykkelid, 1992). We thank Steltenpohl and Bartley for pointing out that back-folds are more rarely developed close to the Ofoten shear zone, and we emphasize that this is in full agreement with our experimental results where a "neutral surface" separates the area of folding from the basement shear zone by a zone of extension.

A first attempt to apply kinematic indicators was made by Cashman (1990), who recognized several extensional basement shear zones that she related to thrusting rather than to regional extensional tectonics. On the basis of later field work and experiments, we reached the conclusion that the extensional shear zones are more likely to postdate thrusting. The apparent truncation of the extensional shear zones by the overlying Treldal thrust can be explained by simultaneous backsliding of nappes and development of extensional shear zones in the basement, as modeled by Fossen and Rykkelid (1992).

The retrogression to greenschist-facies assemblages may be common *within* the orogenic wedge, but information in the literature about the conditions near the basal thrust zone (uppermost part of basement and depositional cover sediments) is scarce and, in part, incorrect (compare undocumented information in Crowley [1989] with data in Rykkelid and Andresen [1993]). Further documentation and discussion of the M2 zoning shown in our Figure 2B can be found in Rykkelid and Andresen (1993, and unpublished).

We emphasize the importance of kinematic analysis being applied in conjunction with more classical structural techniques in future work in any part of the Caledonides.

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Deep Sea Drilling Project Site 612 bolide event: New evidence of a late Eocene impact-wave deposit and a possible impact site, U.S. east coast: Comment and Reply

COMMENT

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Poag et al. (1992) presented convincing evidence for a late Eocene impact-wave deposit on the U.S. east coastal plain. However, I disagree with their suggestion that their proposed impact crater and the impact-wave deposit resulted from the same impact event as the North American (micro)tektite strewn field. Instead, I suggest that their proposed impact crater and the impact-wave deposit most likely were caused by a slightly older impact event, which was not mentioned by Poag et al. (1992).

It has been well established that closely associated with the North American microtektite layer is a slightly older ejecta layer coincident with an iridium anomaly (Glass et al., 1985; Keller et al., 1987). The older layer contains clinopyroxene-bearing spherules, called microkrystites (Glass and Burns, 1988), and it has been found in the eastern equatorial Indian Ocean, equatorial Pacific, Gulf of Mexico, and Caribbean Sea (Glass et al., 1985; Keller et al., 1987). The impact ejecta at Site 612 consists of a microtektite layer on top and a microkrystite layer about 4 cm below (Glass, 1989). I agree with Poag et al. (1992) that the impact ejecta at DSDP Site 612 is the same age as the North American microtektite layer. More precisely, the microtektite layer at Site 612 correlates with the North American microtektite layer, and the microkrystite layer at Site 612 correlates with the widespread microkrystite layer associated with the iridium anomaly. Detailed examination of DSDP Site 612 by Miller et al. (1991) indicates that the microkrystite layer lies immediately above a major hiatus (>4 m.y.). This hiatus has been interpreted as the result of slumping triggered by a bolide impact (Glass, 1989), which apparently can account for the boulder bed (impact-wave deposit) identified by Poag et al. (1992). This means that the impact event that produced the boulder bed and the impact crater of Poag et al. (1992) most likely correlates with the microkrystite event rather than the younger North American microtektite event. Furthermore, the proposed impact crater of Poag et al. (1992) is unlikely to be the source of the North American microtektite layer because there is no detectable hiatus or reworked material associated with the microtektite layer at Site 612 (see Miller et al., 1991), which is about 30 km away from the proposed impact crater. Even if one considers the ejecta at Site 612 as the result of only one impact event, as Poag et al. (1992) apparently did, one should correlate the ejecta with the widely distributed microkrystite layer rather than the North American microtektite layer because microkrystites are present at Site 612, whereas they have never been found in the North American microtektite layer (Glass et al., 1985; Glass, 1989).

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REPLY

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Wei's comments presuppose that the stratigraphically successive abundance peaks of microkrystites and microtektites at DSDP Site 612 (Glass, 1989) resulted from two successive depositional events. We contend, however, that the stratigraphic and depositional relations of ejecta-rich sediments at Site 612 are more complicated than generally realized. We are currently reanalyzing the biostratigraphic record (planktonic foraminifera and calcareous nannofossils) and depositional history at Site 612, and we have reached the following preliminary conclusion: The two ejecta peaks noted by Glass (1989) and subsequent researchers do not record successive deposits; instead, they represent separate lithic constituents simultaneously deposited as part of a single debrite (debris-flow deposit). This debrite encompasses a 22 cm interval of core 612-21-5 (97–119 cm); it includes three different-colored chalk matrices; a variety of allochthonous clasts; mixed assemblages of Paleocene, early Eocene, middle Eocene, and late Eocene microfossils; and several types of ichnofossils (Fig. 1). The two largest, coarsest grained clasts (now clearly outlined by irregular oxidation rims at 111–113 cm and at 113–115.5 cm) yielded the peak concentrations of microtektites, glauconite grains, and other rocks and minerals documented by Glass (1989). Peak values of microkrystites came from matrix three, at 115.5–118 cm, where numerous smaller clasts are present. All