

## The Suess effect in Fiji coral $\delta^{13}\text{C}$ and its potential as a tracer of anthropogenic $\text{CO}_2$ uptake

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### ARTICLE INFO

#### Article history:

Received 14 May 2012

Received in revised form 13 November 2012

Accepted 17 November 2012

Available online 27 November 2012

#### Keywords:

Coral

Carbon isotope

$\text{CO}_2$

Suess effect

Dissolved inorganic carbon

Fiji

Water depth

Skeletal extension rate

### ABSTRACT

In the context of increasing anthropogenic  $\text{CO}_2$  emissions, determining the rate of oceanic  $\text{CO}_2$  uptake is of high interest. Centennial-scale changes in  $\delta^{13}\text{C}$  of the surface water dissolved inorganic carbon (DIC) reservoir have been shown to be influenced by the carbon isotopic composition of atmospheric  $\text{CO}_2$ . However, the availability of direct oceanic  $\delta^{13}\text{C}$  measurements is limited and methods for reconstructing past  $\delta^{13}\text{C}$  variability of the oceanic DIC are needed. Geochemical reconstructions of DIC variability can help in understanding how the ocean has reacted to historical changes in the carbon cycle. This study explores the potential of using temporal variations in  $\delta^{13}\text{C}$  measured in five Fijian *Porites* corals for reconstructing oceanic  $\delta^{13}\text{C}$  variability. A centennial-scale decreasing  $\delta^{13}\text{C}$  trend is observed in these Fiji corals. Other studies have linked similar decreasing  $\delta^{13}\text{C}$  trends to anthropogenic changes in the atmospheric carbon reservoir (the “ $^{13}\text{C}$  Suess effect”). We conclude that solar irradiance is the factor influencing the  $\delta^{13}\text{C}$  cycle on a seasonal scale, however it is not responsible for the centennial-scale decreasing  $\delta^{13}\text{C}$  trend. In addition, variations in skeletal extension rate are not found to account for centennial-scale  $\delta^{13}\text{C}$  variability in these corals. Rather, we found that water depth at which a Fijian *Porites* colony calcifies influences both  $\delta^{13}\text{C}$  and extension rate mean values. The water depth– $\delta^{13}\text{C}$  relationship induces a dampening effect on the centennial-scale decreasing  $\delta^{13}\text{C}$  trend. We removed this “water depth effect” from the  $\delta^{13}\text{C}$  composite, resulting in a truer representation of  $\delta^{13}\text{C}$  variability of the Fiji surface water DIC ( $\delta^{13}\text{C}_{\text{Fiji-DIC}}$ ). The centennial-scale trend in this Fiji coral composite  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  time-series shares similarities with atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$ , implicating the  $^{13}\text{C}$  Suess effect as the source of the this coral  $\delta^{13}\text{C}$  trend. Additionally, our study finds that the  $\delta^{13}\text{C}$  variability between the atmosphere and the ocean in this region is not synchronous; the coral  $\delta^{13}\text{C}$  response is delayed by ~10 years. This agrees with the previously established model of isotopic disequilibrium between atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  and oceanic surface water DIC.

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### 1. Introduction

Many massive scleractinian corals are known to be high-quality archives for paleoclimate reconstruction due to their capacity to incorporate geochemical tracers into their aragonite skeletons (see Druffel, 1997; Corrège, 2006; Pandolfi, 2011 for reviews). Various proxies, such as skeletal Sr/Ca and  $\delta^{18}\text{O}$  have been shown to be accurate tracers of water temperature and  $\delta^{18}\text{O}_{\text{sw}}$  at many sites (Corrège, 2006 for a review). However, because of the influence of both kinetic and metabolic effects, the interpretation of the carbon isotopic composition ( $\delta^{13}\text{C}$ ) of coral skeleton is more complex (McConnaughey, 1989a, 1989b; Grottooli and Wellington, 1999; Grottooli, 2000). The kinetic effect is a physical process occurring during the incorporation of dissolved inorganic carbon (DIC) inside the coral's aragonite skeleton (McConnaughey, 1989b). Kinetic fractionation results from discrimination against heavy

isotopes of both C and O during the hydration and hydroxylation of  $\text{CO}_2$  (McConnaughey, 1989b). This effect produces a simultaneous depletion of  $^{13}\text{C}$  and  $^{18}\text{O}$  in coral skeleton relative to the composition of ambient seawater (McConnaughey, 1986, 1989b). Metabolic fractionation induces additional changes in the skeletal  $\delta^{13}\text{C}$  due to the processes of photosynthesis and respiration (coral/algae symbiotic system) (McConnaughey, 1989a; Grottooli and Wellington, 1999; Grottooli, 2000). These metabolic effects are controlled by external environmental factors, which indirectly affect coral  $\delta^{13}\text{C}$ .

In addition to the influence of metabolic and kinetic effects, decadal- and centennial-scale changes in  $\delta^{13}\text{C}$  of the surface water DIC are reflected in coral  $\delta^{13}\text{C}$  records (Swart et al., 1996a,b; Quinn et al., 1998; Linsley et al., 1999, 2000; Asami et al., 2005; Wei et al., 2009; Swart et al., 2010). These changes in the isotopic composition of the surface water DIC are influenced by the carbon isotopic composition of atmospheric  $\text{CO}_2$  (Druffel and Benavides, 1986). A solid understanding of the relationship between atmospheric  $\text{CO}_2$  and changes in the DIC is needed to refine oceanic  $\text{CO}_2$  uptake dynamics, which is vital in the context of ongoing global change. Monitoring of  $\delta^{13}\text{C}$  in surface water DIC

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has only recently begun (Quay et al., 1992), thus coral-derived  $\delta^{13}\text{C}$  data, could provide a means to reconstruct centennial-scale surface water DIC variability after a thorough assessment of all factors affecting mean annual  $\delta^{13}\text{C}$ .

Many studies have reported a secular-scale decreasing trend in coral  $\delta^{13}\text{C}$  over the 20th century that has been attributed to the addition of  $^{13}\text{C}$ -depleted  $\text{CO}_2$  into the atmosphere by the burning of fossil fuels and deforestation, and the subsequent influx of this  $^{13}\text{C}$ -depleted  $\text{CO}_2$  into the ocean (Keeling et al., 1980; Swart et al., 1996a,b; Quinn et al., 1998; Linsley et al., 1999; Asami et al., 2005; Wei et al., 2009; Swart et al., 2010). In this study we explore intracolony  $\delta^{13}\text{C}$  variability using a network of coral cores from Fiji ( $\sim 17^\circ\text{S}$ ,  $\sim 180^\circ\text{W}$ ). The generation of a coral composite  $\delta^{13}\text{C}$  reconstruction permits the extraction of a clearer signal of regional  $\delta^{13}\text{C}$  variability, which is used to evaluate the isotopic variability of Fiji surface water DIC over time. This will provide a better look at possible factors responsible for oceanic  $\delta^{13}\text{C}$  variability.

## 2. Materials and methods

### 2.1. Coral collection and sampling

Five coral cores from different regions of Fiji were utilized for this study (see Table 1). Linsley et al. (2004, 2006, 2008) and Dassié (2012) provide detailed descriptions of these cores and sampling locations. The coral cores were cut into  $\sim 7$ -mm-thick slabs at the University at Albany-SUNY with a modified tile saw. The slabs were X-rayed with an HP cabinet X-ray system at 35 kV for 90 s, then the X-ray negatives were scanned to generate X-ray positives. Density bands and growth axes were visible on the X-ray positives, which allowed us to determine the sampling tracks and to identify signs of diagenesis. Slabs were cleaned in an ultrasonic bath of deionized water for 30 min. Coral cores FVB1, 16F and FVB2 were also cleaned with a high-energy (500 W, 20 kHz) probe sonicator in a deionized water bath, for approximately 10 min on each slab face. The dried slabs were sampled using a low-speed micro-drill with a 1-mm-round diamond drill bit along the maximum growth axis in tracks (U-shaped groves) parallel to corallite traces, as identified in X-ray positives. A  $\sim 3$ -mm-deep by  $\sim 3$ -mm-wide groove was excavated at 1 mm increments.

### 2.2. Stable isotope analysis

Isotopic analyses followed the procedures summarized in Linsley et al. (2000, 2004, 2006, 2008). Approximately 100  $\mu\text{g}$  of coral powder per sample was dissolved in  $\sim 100\%$   $\text{H}_3\text{PO}_4$  at  $90^\circ\text{C}$  in a Multi-Prep sample preparation device, and the generated  $\text{CO}_2$  gas was analyzed on a Micromass Optima gas-source-triple-collector mass spectrometer at the University at Albany. For the first 14 years of growth, every 1 mm sample (200 samples) was analyzed, and

below this depth every other 1 mm sample was analyzed. Analysis of every other 1 mm sample resulted in a resolution of about 6 to 7 samples per year. Replicate samples were analyzed every 8 samples (every 16 samples after 200 mm). The average difference between duplicate  $\delta^{13}\text{C}$  analyses of 735 samples (for all cores) was 0.066‰. Samples of international standard NBS-19 were interspersed every  $\sim 10$  samples. The standard deviation of 1,039 NBS-19 standards analyzed for  $\delta^{13}\text{C}$  was  $\pm 0.016\%$ .

### 2.3. Chronology

The annual character of coral density banding (see Lough and Barnes, 1989; Barnes and Lough, 1993 for reviews) has proven useful at many sites for developing an accurate coral chronology (Knutson et al., 1972). Density bands for the corals in this study are not always distinct, therefore counting the number of growth bands is ineffective for making an accurate chronology. However, the clear annual cycle ( $\sim 8$  to  $15\text{ mm}\cdot\text{yr}^{-1}$ ) in the  $\delta^{18}\text{O}$  data from Fiji *Porites* allows for the construction of an accurate down-core chronology (see Linsley et al., 2004, 2006). The lightest (most negative)  $\delta^{18}\text{O}$  value in each seasonal cycle was attributed to the warmest month of the year and the heaviest (most positive) value to the coldest. Linsley et al. (2004, 2006) established chronologies for cores 1F and AB, whereas Dassié (2012) established chronologies for FVB1, 16F, and FVB2. This method has been verified by Sr/Ca measurements on coral core 1F (Linsley et al., 2004).

### 2.4. Skeletal extension rate

Skeletal extension rates were determined by counting the number of millimeters between two consecutive  $\delta^{18}\text{O}$  minima ( $\sim$ two consecutive Januarys in each  $\delta^{18}\text{O}$  record). The error associated with this method is  $\pm 1$  mm for the top-most (youngest) 200 samples and  $\pm 2$  mm below that depth when every other 1-mm sample were analyzed.

## 3. Results and discussion

The Fiji  $\delta^{13}\text{C}$  records (except for FVB2, which will be discussed in further detail in Section 3.3) have a secular long-term trend toward increasingly depleted  $\delta^{13}\text{C}$  values towards the present (Fig. 1). Identifying the cause of this  $\delta^{13}\text{C}$  trend is the main focus of this study.

### 3.1. Seasonal $\delta^{13}\text{C}$ variability

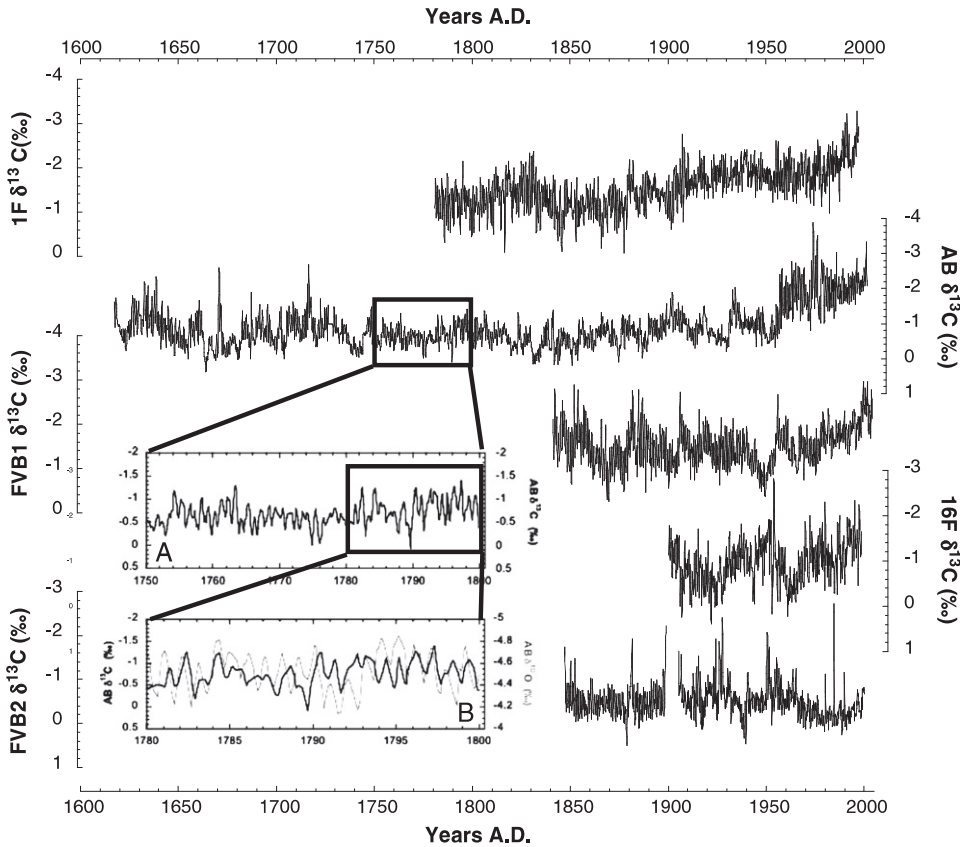
All  $\delta^{13}\text{C}$  coral records presented here have a discernible seasonal cycle (Fig. 1A), in which lower  $\delta^{13}\text{C}$  values are observed during austral winter months and higher  $\delta^{13}\text{C}$  values during austral summer months. When comparing monthly  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  variability, the two signals are not always in phase (e.g. Fig. 1B). The seasonal solar irradiance

**Table 1**  
Description of the coral cores used in this study, AB, 1F, FVB1, 16F and FVB2.

Core ID	Location	Latitude longitude	Water depth <sup>a</sup> (m)	Ext. rate <sup>b</sup> (mm/yr) and standard deviation	Mean $\delta^{13}\text{C}^b$ (‰) and standard deviation
1F	Savusavu Bay	16°49'S 179°14'W	10	12 (2.1)	-1.9 (0.24)
AB	Savusavu Bay	16°49'S 179°14'W	10	11 (2.6)	-1.4 (0.59)
FVB1	Vanua Balavu	17°20.1'S 178°56.7'W	6.0	13 (3.0)	-1.5 (0.3)
16F	Aiwa Island	18°19.21'S 178°43.01'W	3.5	15 (3.8)	-0.93 (0.36)
FVB2	Vanua Balavu	17°20.5'S 178°55.7'W	1.0	16 (4.6)	-0.50 (0.32)

<sup>a</sup> Water depth at top of coral colony.

<sup>b</sup> Values established from annual averaged data from the same common period (1905–1997).



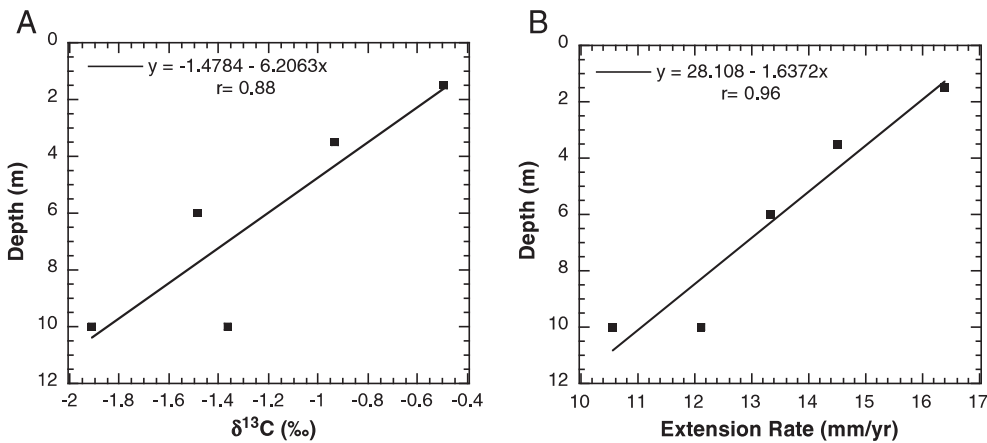
**Fig. 1.** Monthly resolution of Fiji coral  $\delta^{13}\text{C}$  data in ‰ from cores 1F, AB, FVB1, 16F and FVB2. Magnified box A. shows core AB from 1750 to 1800 for a clearer view of the seasonal variability. Magnified box B. shows core AB  $\delta^{13}\text{C}$  (black line) and  $\delta^{18}\text{O}$  (light grey line) from 1780 to 1800.

(SI) cycle precedes the sea surface temperature (SST) maximum by 1 to 2 months for most of Earth's surface. In our Fiji  $\delta^{13}\text{C}$  records, no significant relationship or consistent time lag is observed between  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  time series (e.g. Fig. 1B). Thus the seasonal SST cycle does not seem to be the factor causing seasonal  $\delta^{13}\text{C}$  variability. This finding is in agreement with Romanek et al. (1992), who concluded that  $\delta^{13}\text{C}$  in biogenic carbonate is independent of SST from 10 to 40 °C. Other studies conclude that increases in photosynthetic activity are also linked to increases in coral skeletal  $\delta^{13}\text{C}$  (Weber and Woodhead, 1970; Swart, 1983; McConnaughey, 1989a; Allison et al., 1996; Grottoli, 1999,

2000; Grottoli and Wellington, 1999; Omata et al., 2008). Therefore, the seasonal SI cycle could be indirectly responsible for the seasonal cycle observed in  $\delta^{13}\text{C}$  records by influencing photosynthetic activity (Swart et al., 1996b; Reynaud-Vaganay et al., 2001; Shimamura et al., 2008).

3.2. Water depth influences on coral  $\delta^{13}\text{C}$  and extension rate

Fiji coral cores were collected at different water depths, ranging from 1.5 to 10 m (Table 1), which were measured from the top of



**Fig. 2.** Regression equations and their correlation coefficients ( $r$ ) between annual average coral  $\delta^{13}\text{C}$  (over the common time period, 1905–1997) and coral collection depth (A), and between coral extension rate and coral collection depth (B).



each coral colony growth surface to the ocean surface. For the time period common to all coral records (1905–1997), an offset in the mean value (bulk value) is present between the different coral  $\delta^{13}\text{C}$  records. For the same common time period, the mean  $\delta^{13}\text{C}$  values are  $-1.9\%$  for 1F,  $-1.4\%$  for AB,  $-1.5\%$  for FVB1,  $-0.93\%$  for 16F, and  $-0.5\%$  for FVB2 (Table 1).

Individual coral skeletal  $\delta^{13}\text{C}$  bulk data (averaged over the common time period, 1905–1997) were plotted against their respective water depths (Fig. 2A). A negative linear regression ( $r^2 = -0.77$ ) between water depth and  $\delta^{13}\text{C}$  is observed ( $p < 0.01$ , using a bootstrap resampling method with 1000 resamples). Corals growing deeper in the water column have lower (more negative) mean  $\delta^{13}\text{C}$  values than corals growing at shallower water depths. The relationship between water depth and  $\delta^{13}\text{C}$  implies that water depth is the main factor controlling mean  $\delta^{13}\text{C}$  values in these corals, as previously reported by Weber and Woodhead (1970, 1971). The amount of photosynthetic activity by symbiotic zooxanthellae varies with the amount of light; light attenuation with increasing water depth results in reduced photosynthetic activity, as demonstrated by McCloskey and Muscatine (1984) and Grottoli (1999). It has been established that skeletal  $\delta^{13}\text{C}$  increases with increasing photosynthetic activity (Weber and Woodhead, 1970; Swart, 1983; McConnaughey, 1989a). This water depth- $\delta^{13}\text{C}$  relationship is consistent with other published studies (Weber et al., 1976; Fairbanks and Dodge, 1979; Swart and Coleman, 1980; McConnaughey, 1989a; Leder et al., 1991; Grottoli, 1999; Grottoli and Wellington, 1999; Grottoli, 2000).

An insignificant correlation is present between  $\delta^{18}\text{O}$  and water depth ( $r^2 = 0.013$  with  $p > 0.05$ , using a bootstrap resampling method with 1000 resamples). This suggests that the observed  $\delta^{13}\text{C}$ -water depth relationship in these corals is primarily related to metabolic, rather than kinetic factors, the latter of which would have simultaneously affected both  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values. Grottoli (1999) and Grottoli and Wellington (1999) published similar findings for corals from similar water depths.

Additionally, a distinct relationship is present between mean skeletal extension rate (SER) and water depth (Fig. 2B). For the common time period, the mean SER is  $12 \text{ mm}\cdot\text{yr}^{-1}$  for 1F,  $11 \text{ mm}\cdot\text{yr}^{-1}$  for AB,  $13 \text{ mm}\cdot\text{yr}^{-1}$  for FVB1,  $15 \text{ mm}\cdot\text{yr}^{-1}$  for 16F, and  $16 \text{ mm}\cdot\text{yr}^{-1}$  for FVB2 (Table 1). Corals residing at deeper water depths exhibit lower SERs (Fig. 2b). This linear relationship ( $r^2 = 0.92$ ;  $p < 0.01$ , using a bootstrap resampling method with 1000 resamples) has been previously described by Felis et al. (1998), Grottoli (1999), Rosenfeld et al. (2003), and Omata et al. (2005) for *Porites* at other sites. This is a light-dependent relationship; light attenuation increases with water depth, resulting in a decrease in SER. A decrease in photosynthetic activity reduces the amount of carbon available for the coral to grow (McCloskey and Muscatine, 1984; Grottoli, 1999). Another possible cause of this relationship could involve coral morphological changes with water depth (Grottoli, 1999), which is beyond the scope of this study.

Coral cores 1F and AB were collected from colonies residing at the same water depth, however their respective mean SER and mean  $\delta^{13}\text{C}$  values are slightly different. The difference between the mean  $\delta^{13}\text{C}$  and mean SER of AB and 1F could be used to set the regression error for both  $\delta^{13}\text{C}$ - and SER-water depth relationships. It is worth noting that this offset might be due to microenvironmental (Pandolfi, 2011 for a review) or genetic differences between the two colonies (Forsman et al., 2009), which are also beyond the scope of this study.

### 3.3. $\delta^{13}\text{C}$ and coral extension rate

The different  $\delta^{13}\text{C}$  time series were centered by subtracting their mean values for the common time period (1997–1905) in order to facilitate a comparison between them. This process removed the influences of both water depth and colony-specific disparities on the mean values of the  $\delta^{13}\text{C}$  records (Section 3.2). Annual averages of the coral  $\delta^{13}\text{C}$  data (arithmetic mean of the monthly data) were compared to their respective SER values (Fig. 3). SER and  $\delta^{13}\text{C}$  are negatively correlated (Fig. 3);

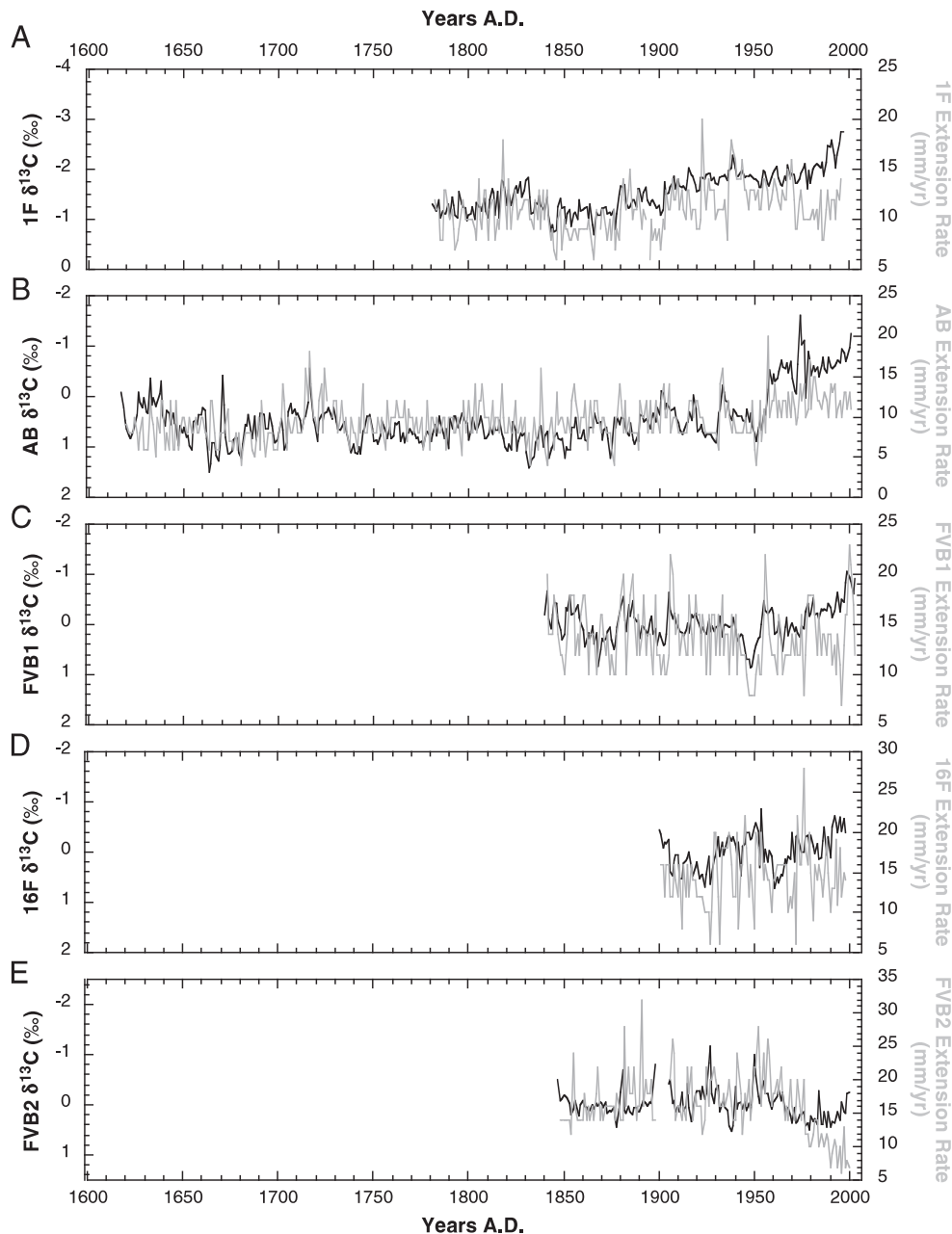
when SER increases,  $\delta^{13}\text{C}$  decreases. This suggests that SER and  $\delta^{13}\text{C}$  are influenced by a common factor. The correlation coefficients between SER and  $\delta^{13}\text{C}$  are  $-0.20$  ( $p < 0.05$ ) for 1F,  $-0.61$  ( $p < 0.01$ ) for AB,  $-0.41$  ( $p < 0.01$ ) for FVB1,  $-0.22$  ( $p < 0.05$ ) for 16F, and  $-0.44$  ( $p < 0.01$ ) for FVB2, with a degree of freedom (df) of 90 ( $df = n - 2$ ). The significance of the correlation coefficients was determined using the Pearson Product-moment correlation coefficient table, and will be used for the remainder of this report.

During the mid-20th century, the depletion of skeletal  $\delta^{13}\text{C}$  in all cores begin to accelerate, while their respective SERs remain generally constant (Fig. 3). The onset of the trend in the  $\delta^{13}\text{C}$  records might have reduced its correlation with SER. A slight increasing trend is present in the SER records; this trend is a result of the corals' cumulative growth. As the coral colonies grow toward the surface, an increase in light intensity leads to an increase in photosynthetic activity, which leads to increase in extension rate (see Section 3.2) (see Huston, 1985; Heiss et al., 1993; Hoogenboom et al., 2006 for reviews).

Both SER and  $\delta^{13}\text{C}$  values for coral core FVB2 present a peculiar pattern beginning around 1960 (Fig. 3). A decreasing trend is observed in all the other four Fiji coral records, whereas FVB2 mean  $\delta^{13}\text{C}$  undergoes a  $0.2\%$  increase between the pre-1960 and post-1960 periods. Additionally, there is a decreasing trend in SER values from 1960 to present ( $r = -0.85$ ;  $df = 38$ ;  $p < 0.01$ ). SER varies from  $32 \text{ mm}\cdot\text{yr}^{-1}$  to  $12 \text{ mm}\cdot\text{yr}^{-1}$  with a mean of  $17 \text{ mm}\cdot\text{yr}^{-1}$  for the older portion of the record, whereas the younger portion of the record presents a general decreasing trend going from an average of  $18 \text{ mm}\cdot\text{yr}^{-1}$  around 1960 to an average of  $7 \text{ mm}\cdot\text{yr}^{-1}$  for the most recent part of the record. Coral FVB2 grew in relatively shallow water (1 m water depth at low tide) and the top of the colony was partially dead. Coral reefs live near their upper thermal tolerance limits (Coles et al., 1976) as well as their ultraviolet (UV) radiation limits, and generally do not tolerate conditions outside of the normal (average) range (Glynn, 1996; Hoegh-Guldberg, 1999). Incidences of coral bleaching have been associated with warmer-than-normal sea surface temperature (SST) conditions (e.g. Glynn, 1984; Goreau et al., 1993; Hoegh-Guldberg and Salvat, 1995; Winter et al., 1998). Prolonged exposure to increased SSTs (even for a short period) can reduce coral growth rates (Jokiel and Coles, 1977, 1990; Coles and Jokiel, 1978; Glynn, 1993, 1996; Meesters and Bak, 1993) and leave the skeleton significantly enriched in  $\delta^{13}\text{C}$  (Allison et al., 1996). Light plays a major role in providing the energy that drives photosynthesis in zooxanthellae (Chalker et al., 1988). UV radiation has been reported to have destructive effects on zooxanthellae activity (Jokiel, 1980; Fisk and Done, 1985; Harriott, 1985; Oliver, 1985; Goenaga et al., 1988; Lesser et al., 1990; Gleason and Wellington, 1993) and may result in decreased growth rates (Lesser, 1996). Brown (1997) concluded that a combination of elevated temperature and high irradiance is a likely cause of coral bleaching in shallow waters. McGregor et al. (2011) pointed out that microatoll types of corals (reef-flat corals) might have a stronger resistance to changes in SST and solar irradiance. The coral colony from which core FVB2 was collected did not have a flat surface, and its extreme shallow-water environment of the core likely caused the colony to undergo bleaching and partial mortality, affecting both its SER and  $\delta^{13}\text{C}$  values for the top portion.

Previous studies have described an increase in SER following an increase in SST (Nie et al., 1997; Lough and Barnes, 2000; Bessat and Buigues, 2001). However, no specific relationship between SER and SST was observed in this study. One reason could be related to the error associated with the method used for determining SER, which is not as precise as the gamma densitometry method (Chalker and Barnes, 1990) used in the aforementioned studies. Another reason for not finding a relationship between SST and skeletal  $\delta^{13}\text{C}$  could be due to the small interannual SST cycle at Fiji. Lough and Barnes (2000) found that an SST increase of  $1^\circ\text{C}$  resulted in a  $3.1 \text{ mm}$  increase in *Porites* SER, and similarly, Nie et al. (1997) found a  $1.5 \text{ mm}$  increase per  $0.5^\circ\text{C}$  increase. The mean interannual SST amplitude at Fiji is





**Fig. 3.** Centered coral  $\delta^{13}\text{C}$  data (black lines) from the five Fiji coral cores 1F (A), AB (B), FVB1 (C), 16F (D), and FVB2 (E), and their respective extension rates in millimeters per year (gray lines).

$-0.26\text{ }^{\circ}\text{C}$  (Smith et al., 2008). Following Lough and Barnes' study, this would imply an interannual SER variability of  $\sim 0.8\text{ mm}$ . This calculated  $0.8\text{ mm}$  variability falls within the range of error associated with our SER measurements. The interannual SST range at Fiji is likely too small to cause any significant variability in SER.

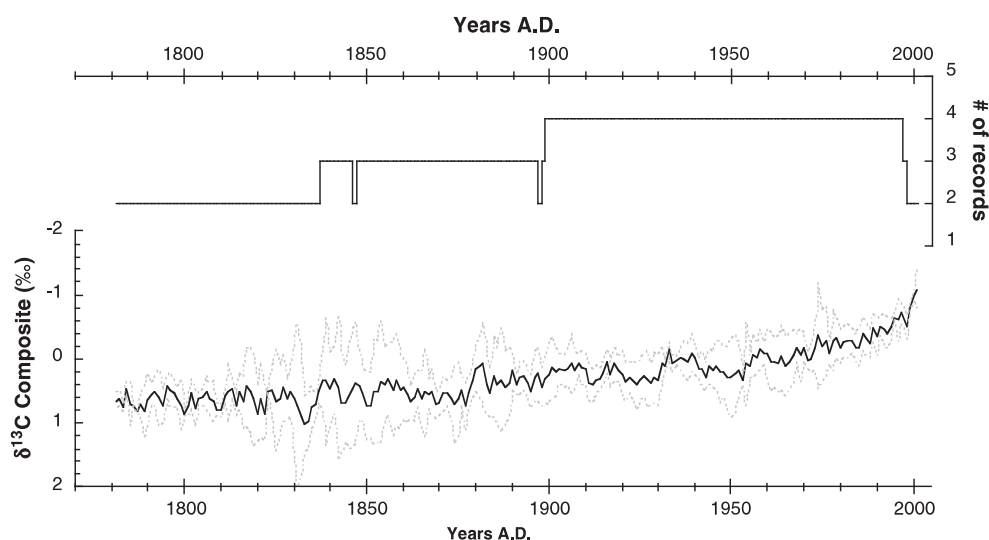
### 3.4. $\delta^{13}\text{C}$ composite and long term variability

#### 3.4.1. Composite

The creation of a composite record is a method traditionally utilized in tree ring studies. Previous coral-based studies have stressed the idea of using a network of coral records at multiple sites (Tudhope et al., 1995; Gagan et al., 1998; Linsley et al., 1999, 2006, 2008; Lough, 2004; Delong et al., 2007; Dassié, 2012). A coral core composite is likely to reflect regional-scale variability and filter out site-specific influences. In

this study, a  $\delta^{13}\text{C}$  composite was created by averaging the  $\delta^{13}\text{C}$  records (centered data) from the four coral cores: 1F, AB, FVB1, and 16F (Fig. 4). Centering the data provides the benefit of removing the water depth influence on the individual coral  $\delta^{13}\text{C}$  records. Section 3.3 discussed the peculiar pattern in FVB2  $\delta^{13}\text{C}$  starting around 1960 when both its skeletal  $\delta^{13}\text{C}$  and SER values begin decreasing, in disagreement with the other four Fiji coral records. An anomalous pattern was also observed in FVB2  $\delta^{18}\text{O}$  values (Dassié, 2012). The shallowness of the coral colony (1 m water depth) likely caused the colony to bleach and partially die. Therefore FVB2  $\delta^{13}\text{C}$  data were not included in the composite.

The  $\delta^{13}\text{C}$  composite only includes years where there are two or more records present; therefore it spans from 1781 to 2001 (Fig. 4). The confidence level of the composite (gray dotted lines in Fig. 4) takes into account the number of records used. For each year, the standard deviation between each core was determined, which was then divided by the



**Fig. 4.** Top panel: number of coral records included in the Fiji coral core composite. Bottom panel: centered annual averaged Fiji coral  $\delta^{13}\text{C}$  composite (solid black line) and its 95% confidence interval (dotted gray line).

square root of the sample size (number of coral cores used to create the composite). The confidence level was computed at 95%.

For the entire length of the record, the coral  $\delta^{13}\text{C}$  composite presents a general trend toward increasingly depleted  $\delta^{13}\text{C}$  values ( $r = -0.87$ ;  $df = 219$ ;  $p < 0.01$ ). This accelerating decrease exceeds the limits of the confidence interval for the composite  $\delta^{13}\text{C}$  (Fig. 4) and translates to a decreasing rate of  $\sim 0.0052\% \cdot \text{yr}^{-1}$  ( $\sim 1.2\%$  decrease over a 221-year period). Other studies on the Pacific Ocean have reported long-term decreasing trends in coral  $\delta^{13}\text{C}$  data (Quinn et al., 1998; Linsley et al., 1999; Stephans et al., 2004; Asami et al., 2005; Wei et al., 2009 and Swart et al., 2010 for a compilation). The magnitudes of the trends from these studies are summarized and compared with the values from this study in Table 2. The magnitude of decrease in  $\delta^{13}\text{C}$  observed in the Fiji corals is similar to that found by other studies that sampled *Porites* sp. from the Pacific Ocean (Table 2). These inter-site similarities suggest that the coral  $\delta^{13}\text{C}$  composite reflects regional, and not just local DIC variability. The rate of the  $\delta^{13}\text{C}$  decrease becomes higher during the 1960–1990 period ( $-0.014\% \cdot \text{yr}^{-1}$ ;  $r^2 = 0.65$ ,  $df = 28$ ,  $p < 0.01$ ), compared to a rate of  $-0.0062\% \cdot \text{yr}^{-1}$  ( $r^2 = 0.58$ ,  $df = 88$ ,  $p < 0.01$ ) for the 1900–1990 period. The rate of coral  $\delta^{13}\text{C}$  decrease is therefore accelerating with time, which aligns with the conclusion of Swart et al. (2010). However, rates of decrease for the Fiji composite are higher than the rate found by Swart et al. (2010) for their Pacific Ocean compilation ( $-0.0066\% \cdot \text{yr}^{-1}$  for the 1960–1999 period and  $-0.0027\% \cdot \text{yr}^{-1}$  for the 1900–1990 period). Fiji is located in the South Pacific Convergence Zone (SPCZ), which is an area of high wind that is favorable to a high rate of gas exchange between the atmosphere and the ocean (Wanninkhof and McGillis, 1999; Nightingale et al., 2000). Moreover, the Revelle factor, which is an indicator of the buffering capacity of seawater (Revelle and Suess, 1957; Zeebe and Wolf-Gladrow, 2001), is not constant through the Pacific Ocean (see Sabine et al., 2004 Fig. 3). These factors suggest that the rate of  $\delta^{13}\text{C}$  decrease is location-dependent.

### 3.4.2. Solar irradiance influence

With the understanding that SI and  $\delta^{13}\text{C}$  are positively correlated (relationship determined at seasonal time scale — Section 3.1) a long term decreasing trend in the SI data would be expected. The release of human-made aerosol particles into the atmosphere enhances light scattering and absorption, inducing a large reduction of SI reaching Earth's surface (Ramanathan et al., 2001). This global dimming effect has been recorded by instrumental data around the world during the 1950–1990 period (Ramanathan et al., 2001; Liepert, 2002; Liepert et al., 2004; and Wild et al., 2005), making this reduction in SI a possible cause for the centennial scale  $\delta^{13}\text{C}$  decrease observed in the coral skeleton. Peaks of aerosols are found near their source of production (Ramanathan et al., 2001), thus since the Fiji archipelago is not located near any major industrial sites, this dimming effect should not be a significant factor in Fiji. Moreover, no significant decrease in SI is observed during the late 20th century, and the global dimming effect fades after 1985 (Wild et al., 2005). Furthermore, coral extension rates, which are indirectly related to SI via photosynthesis (Sections 3.2 and 3.2) do not present any sign of change  $\sim 1950$ . Additionally, various global-scale SI reconstructions show a slight increasing trend over the last several centuries (Wang et al., 2005; Schmidt et al., 2011). Therefore, we conclude that the observed long-term decrease in coral  $\delta^{13}\text{C}$  cannot be attributed to SI variability.

### 3.4.3. Cumulative-growth influence

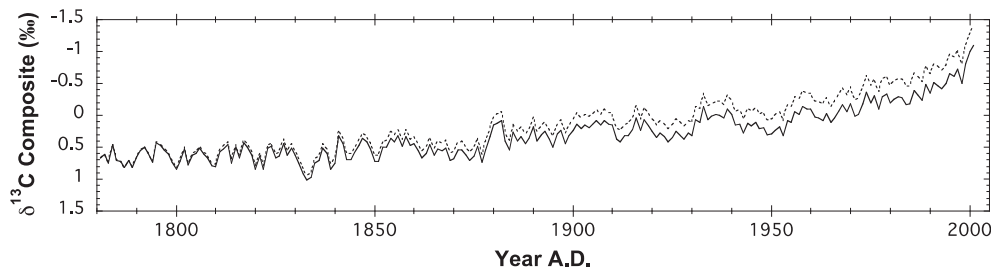
Section 3.2 discusses how skeletal  $\delta^{13}\text{C}$  decreases with increasing water depth. If water depth influences long-term skeletal  $\delta^{13}\text{C}$  variability,  $\delta^{13}\text{C}$  is expected to undergo an increase as the coral grows toward the surface (decrease of water depth). However, Fiji coral  $\delta^{13}\text{C}$  undergoes a long-term decrease (Fig. 4). Therefore the variation in water depth due to cumulative coral growth could not be responsible for the long-term decreasing  $\delta^{13}\text{C}$  trend. The water depth effect on  $\delta^{13}\text{C}$

**Table 2**

A comparison of the coral  $\delta^{13}\text{C}$  magnitude over common time periods, between the Fiji coral  $\delta^{13}\text{C}$  composite and other published studies.

Publication	Region	Period	Coral species	Change in $\delta^{13}\text{C}$	Change in Fiji composite $\delta^{13}\text{C}$ change <sup>a</sup>
Wei et al., 2009	Great Barrier Reef (SW Pacific)	1807–1997	<i>Porites</i> sp.	$\sim -1.1$	$\sim -1.3$
Asami et al., 2005	Guam (SW Pacific)	1950–2000	<i>Porites lobata</i>	$\sim -0.80$	$\sim -0.90$
Linsley et al., 1999	Clipperton Atoll (E Pacific)	1970–1994	<i>Porites lobata</i>	$\sim -0.40$	$\sim -0.50$
Stephans et al., 2004	New Caledonia (SW Pacific)	1967–1992	<i>Porites lutea</i>	$\sim -0.38$	$\sim -0.34$

<sup>a</sup> Change in Fiji composite  $\delta^{13}\text{C}$  for the coinciding time period (third column).



**Fig. 5.** Centered annual averaged coral  $\delta^{13}\text{C}$  composite (solid black line) and  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  (dotted black line) created from centered annual averaged coral  $\delta^{13}\text{C}$  composite with the influence of cumulative growth removed.

(Fig. 2) opposes the observed decreasing trend, since a decrease in water depth is linked to an increase in  $\delta^{13}\text{C}$ . In order to make the composite more reflective of the regional DIC, this cumulative growth influence on the composite  $\delta^{13}\text{C}$  record was calculated and removed. The average cumulative growth of all four coral cores from the entire length of the coral composite record (1781–2001) was determined to be  $\sim 2.5$  m. Using the equation of  $\delta^{13}\text{C}$  vs. water depth (Fig. 2), the total variation in  $\delta^{13}\text{C}$  resulting from a water depth change of 2.5 m was calculated. A fraction of this value was annually incorporated from 1781 to 2001 (Fig. 5 dotted line), which caused the slope of the  $\delta^{13}\text{C}$  vs. time relationship to increase, resulting in a higher rate of coral  $\delta^{13}\text{C}$  depletion. The rate of decrease of  $\delta^{13}\text{C}$  minus the cumulative growth influence is similar to the rate of  $\delta^{13}\text{C}$  decrease of the DIC as recorded by ship measurements in the Pacific Ocean (Kroopnick et al., 1977; Gruber et al., 1999; Quay et al., 2003) (Table 3). These results suggest that removing the water depth influence on the coral  $\delta^{13}\text{C}$  composite results in a truer representation of regional surface water DIC  $\delta^{13}\text{C}$  variability. The Fiji composite  $\delta^{13}\text{C}$  time series minus the cumulative growth influence will be referred as  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  for the remainder of this report. There is no known upwelling near our study sites in Fiji, therefore  $\delta^{13}\text{C}$  of the surface water DIC (and thus  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$ ) should principally vary as a function of atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$ .

#### 3.4.4. $^{13}\text{C}$ Suess effect influence

The continuous increase in atmospheric  $\text{CO}_2$  concentration since the early 19th century has been attributed to human activities (deforestation, agricultural practices, and the combustion of fossil fuels; see IPCC 4th Assessment Report for a review). This anthropogenic  $\text{CO}_2$  is depleted in  $^{13}\text{C}$ ; therefore the  $^{13}\text{C}/^{12}\text{C}$  ratio ( $\delta^{13}\text{C}$ ) of the atmospheric carbon reservoir is decreasing as the atmospheric  $\text{CO}_2$  concentration continues to increase. The ocean is one of the main global sinks of anthropogenic  $\text{CO}_2$  emissions (Quay et al., 1992), consequently, the last two-centuries of atmospheric  $\delta^{13}\text{C}$  depletion, termed “the  $^{13}\text{C}$  Suess effect”, is reflected in  $\delta^{13}\text{C}$  of the oceanic DIC reservoir.

The Fiji  $\delta^{13}\text{C}$  composite time series minus the cumulative growth influence ( $\delta^{13}\text{C}_{\text{Fiji-DIC}}$ ) has a long-term decreasing trend, with a rate of  $\sim 0.0066\% \cdot \text{yr}^{-1}$  ( $\sim 1.5\%$  decrease over a 221-year period). This decrease, present in many other published coral studies (Nozaki et al., 1978; Swart et al., 1996a,b; Quinn et al., 1998; Linsley et al., 1999, 2000; Asami et al., 2005; Wei et al., 2009; Swart et al., 2010) is attributable to the oceanic  $^{13}\text{C}$  Suess effect.

The low-frequency variability of the  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  record (frequencies inferior to 100 years) was extracted using singular spectrum analysis (SSA) software. A detailed description of this technique and its paleoclimate application is explained by Vautard and Ghil (1989), and Vautard et al. (1992). E. Cook from the Lamont–Doherty Earth Observatory wrote the SSA program used in this study. The extracted low frequency variability and its first derivative are plotted in Fig. 6. The depletion of  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  is not constant through time, but presents periods of acceleration and deceleration. The earliest part of the record undergoes a steady depletion until  $\sim 1825$ , when it begins to accelerate, reaching a maximum around 1880. For the next  $\sim 40$  years (1880 to 1920) the depletion decelerates before accelerating again, reaching an unprecedented rate (in comparison to the last 200 years) around 1995.

The atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  and atmospheric  $\text{CO}_2$  concentration data (obtained from Antarctic ice core and firn samples by Francey et al., 1999) are compared to the  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  long-term trend (Fig. 7). The  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  time series presents similarities with the atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  record and the reconstructed atmospheric  $\text{CO}_2$  concentration time series.

The atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  time series is discontinuous, therefore the raw atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  data had to be linearly interpolated into annual data (using the ARAND software package by Howell et al., 2006) prior to extracting the low frequency signal and calculating its first derivative (using the same methods as for  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$ ). The magnitudes of the coral  $\delta^{13}\text{C}$  and atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  decreasing trends (Fig. 8) are approximately the same ( $\sim 1.5\%$ ). The decelerating period (1880–1920) observed in the first derivative of  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  is not present in the 1st derivative of the atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  record (Fig. 8), however a halt in acceleration of the atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  depletion rate appears to precede the deceleration in  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  depletion. This slowing of the atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  depletion may have resulted in the deceleration of  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  depletion and/or some local factors may have affected the variability of the  $\delta^{13}\text{C}$  DIC (see Pandolfi, 2011 for a review).

The atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  depletion begins to accelerate in the early to mid-20th century, followed by an acceleration of  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  depletion (Fig. 8). The depletion observed between 1960 and 1990 in atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  (ice core data; Francey et al., 1999), is  $-0.030\% \cdot \text{yr}^{-1}$  ( $r^2 = 0.98$ ,  $df = 28$ ,  $p < 0.01$ ). This rate is close to the  $-0.023$  to  $-0.029\% \cdot \text{yr}^{-1}$  range found by Keeling (2005) from direct measurements of atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  at Manua Loa. The  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  depletion for the same time period is  $-0.015\% \cdot \text{yr}^{-1}$  ( $r^2 = 0.69$ ,  $df = 28$ ,  $p < 0.01$ ), which is  $\sim 50\%$  of the atmospheric  $\delta^{13}\text{C}$  depletion found by Francey et al. (1999) and  $\sim 52\%$  to

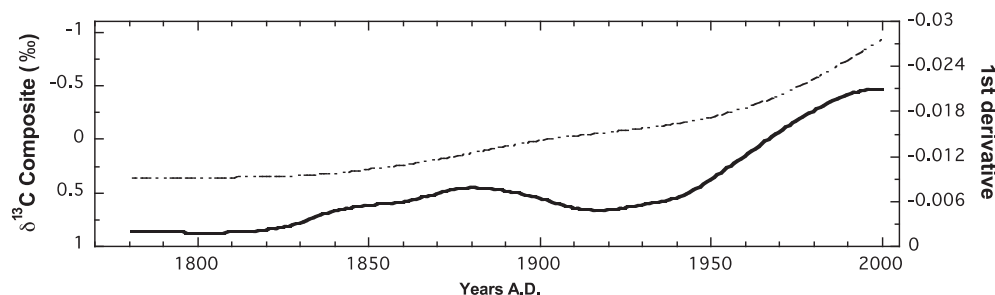
**Table 3**

A comparison of the coral  $\delta^{13}\text{C}$  depletion over common time periods, between the Fiji coral  $\delta^{13}\text{C}$  composite, minus the depth influence ( $\delta^{13}\text{C}_{\text{Fiji-DIC}}$ ), and published values of measured  $\delta^{13}\text{C}$  of the dissolved inorganic carbon (DIC).

Publication	Period	Region	Change in $\delta^{13}\text{C}$ of DIC	Change in $\delta^{13}\text{C}_{\text{Fiji-DIC}}$ <sup>a</sup>
Kroopnick et al., 1977	1970–1993	Pacific Ocean	$-0.39\%$	$\sim -0.30\%$
Gruber et al., 1999	$\sim 1980$ –1990	Tropical Pacific	$-0.15 \pm 0.06\%$	$\sim -0.19\%$
Quay et al., 2003	$\sim 1970$ s–1990s	Pacific ( $60^\circ\text{S}$ – $55^\circ\text{N}$ )	$-0.16 \pm 0.02\%$ /decade	$\sim -0.14\%$ /decade

<sup>a</sup> Change in  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  for the coinciding time period (third column).





**Fig. 6.** The extracted low frequency (frequencies less than 100 years) of the Fiji composite  $\delta^{13}\text{C}$  time series minus the cumulative growth influence ( $\delta^{13}\text{C}_{\text{Fiji-DIC}}$ ) (dashed black) and its first derivative (black).

65% of the atmospheric  $\delta^{13}\text{C}$  depletion determined by Keeling (2005). This difference in depletion between atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  and surface water DIC ( $\delta^{13}\text{C}_{\text{Fiji-DIC}}$ ) is similar to what Böhm et al. (1996) found using Sclerosponges from the Coral Sea. They determined that 50% to 65% of the atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  decrease is observed in the surface water DIC  $\delta^{13}\text{C}$  of the Coral Sea. Quay et al. (2003) found similar results; they determined that the decrease in  $\delta^{13}\text{C}$  of the DIC per decade represents about 70% of the measured atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  depletion over the 1970–1990 period.

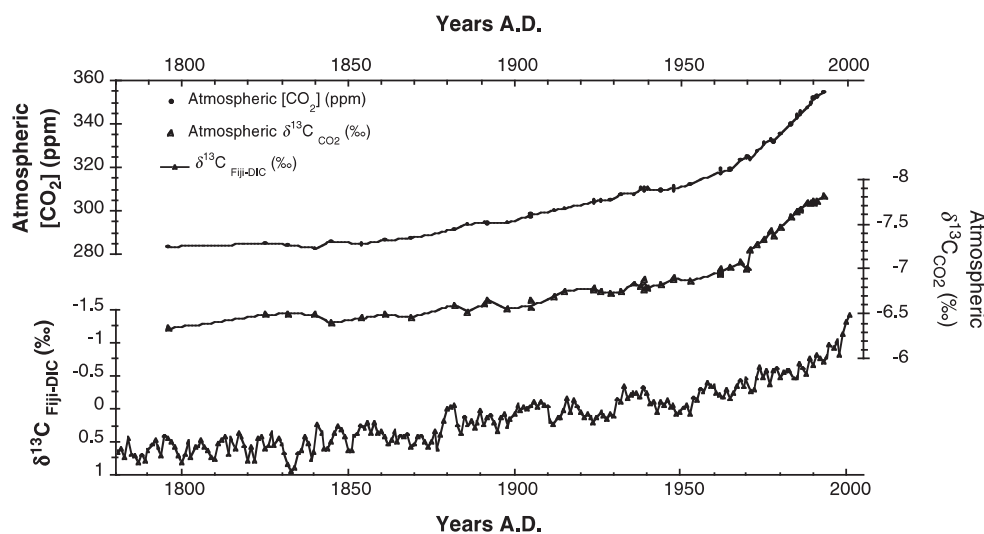
The increase in the rate of atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  depletion around the early-mid 20th century precedes the increase in coral  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  depletion (Figs. 7 and 8). The  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  depletion from 1970 to 2000 is  $0.023\% \cdot \text{yr}^{-1}$ , which is within the range of atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  depletion for the 1960–1990 period presented by Keeling (2005). This finding suggests that the  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  depletion lags that of atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  by  $\sim 10$  years. The correlation coefficient between atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  and  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  increases when the data are offset by 10 years (from  $r^2 = 0.64$  to  $r^2 = 0.72$  ( $df = 28$ ;  $p < 0.01$ )). This is in agreement with the established model of isotopic disequilibrium between atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  and oceanic surface water DIC being on the order of 10 years (Broecker et al., 1980; Walker and Kasting, 1992).

The recent development of an oceanic  $\delta^{13}\text{C}$  database, including measurements made during the World Ocean Circulation Experiment and the Ocean Atmosphere Carbon Exchange Study programs allowed Quay et al. (2003) to estimate a global oceanic  $\delta^{13}\text{C}$  surface change of  $-0.16 \pm 0.2\% \cdot \text{yr}^{-1}$  per decade for the 1970–1990 period ( $-0.18 \pm 0.5\% \cdot \text{yr}^{-1}$  per decade for the Pacific Ocean). Prior to the work of Quay et al. (2003), studies using the change in  $\delta^{13}\text{C}$  of the DIC to estimate the

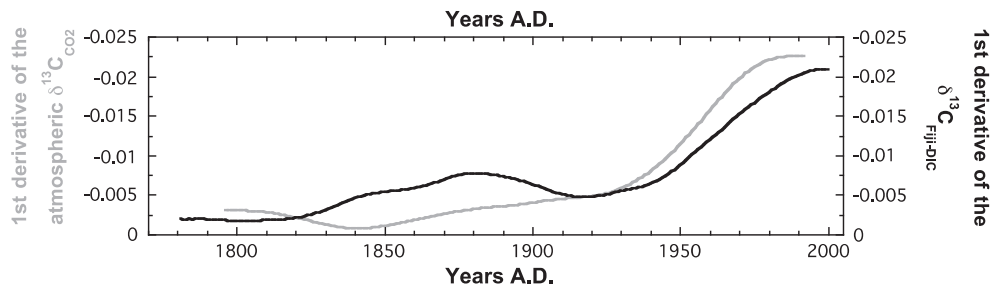
rate of anthropogenic  $\text{CO}_2$  uptake by the oceans (Quay et al., 1992, 2003; Bacastow et al., 1996; Heimann and Maier-Reimer, 1996) faced uncertainties in their calculations due to a lack of oceanic  $\delta^{13}\text{C}$  measurements. Using the ocean-wide  $\text{CO}_2$  uptake rate of  $-0.16\% \cdot \text{yr}^{-1}$  per decade, Quay et al. (2003) estimated that  $\sim 1.5 \pm 0.6 \text{ Gt C yr}^{-1}$  entered the global ocean over the period 1970–1990. The rate of  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  depletion we observed over the same period ( $\sim -0.19\% \cdot \text{yr}^{-1}$  per decade) is close to the Pacific Ocean  $\delta^{13}\text{C}_{\text{DIC}}$  rate as determined by Quay et al. (2003), however the rate of  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  depletion increases with time as discussed above. This increase leads us to conclude that the rate of oceanic  $\text{CO}_2$  uptake might have correspondingly increased. Similar reconstructions of past  $\delta^{13}\text{C}$  variability of the DIC using corals at other sites could help in determining the spatial and temporal patterns of past oceanic  $\text{CO}_2$  uptake, and could potentially be used as a baseline for ocean model simulations.

#### 4. Conclusion

This study highlights the inter-colony reproducibility of coral skeletal  $\delta^{13}\text{C}$ , as well as the benefits of establishing a coral  $\delta^{13}\text{C}$  composite record. Whereas solar irradiance seems to be the factor influencing the seasonal cycle in coral skeletal  $\delta^{13}\text{C}$ , we conclude that it cannot be responsible for the long term decreasing  $\delta^{13}\text{C}$  trend. Additionally, coral skeletal extension rate variability does not account for the long-term change in coral  $\delta^{13}\text{C}$ , since extension rate is relatively constant with time and does not change significantly after  $\sim 1950$  A.D. However, the water depth at which a coral colony is living influences both  $\delta^{13}\text{C}$  and extension rate mean values. Corals growing deeper in



**Fig. 7.** Fiji composite  $\delta^{13}\text{C}$  time series minus the cumulative growth influence ( $\delta^{13}\text{C}_{\text{Fiji-DIC}}$ ) (bottom), reconstructed atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  (middle) and reconstructed atmospheric  $\text{CO}_2$  concentration (top) time series. Atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  and atmospheric  $\text{CO}_2$  concentration data are obtained from Antarctic ice core and firn samples (Francy et al., 1999).



**Fig. 8.** The first derivative of the Fiji composite  $\delta^{13}\text{C}$  time series minus the cumulative growth influence ( $\delta^{13}\text{C}_{\text{Fiji-DIC}}$ ) (black) and the first derivative of the atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  (Francey et al., 1999) (gray). The first derivatives were made from the extracted low frequency (inferior to 100 years) variability of both records.

the water column present lower mean  $\delta^{13}\text{C}$  values and lower mean extension rates than corals growing at shallower water depths.

The Fiji  $\delta^{13}\text{C}$  coral composite was created by averaging the four centered coral records: AB, 1F, FVB1, and 16F. Centering the data provides the benefit of removing the water depth influence on the individual coral records residing at different water depths. A separate type of water depth influence results from a decrease in water depth as the top of the colony grows toward the surface, which induces a cumulative dampening effect on the long-term decreasing  $\delta^{13}\text{C}$  trend. This influence was removed from the  $\delta^{13}\text{C}$  composite, creating a refined  $\delta^{13}\text{C}$  Fiji composite ( $\delta^{13}\text{C}_{\text{Fiji-DIC}}$ ), which resulted in a truer representation of  $\delta^{13}\text{C}$  variability of the Fiji surface water DIC. The oceanic  $^{13}\text{C}$  Suess effect has high spatial variability (Quay et al., 1992), but the coral record composite utilized for this study provides a regional view of  $\delta^{13}\text{C}$  variability of the DIC around Fiji, as proven by its similarities with measured  $\delta^{13}\text{C}_{\text{DIC}}$  (Kroopnick et al., 1977; Gruber et al., 1999; Quay et al., 2003).

The long-term  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  signal becomes increasingly depleted toward the present. Similarities between atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  and  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  records suggest that the increase in isotopically-light atmospheric  $\text{CO}_2$  is reflected in surface water DIC, which is in turn reflected in Fiji coral  $\delta^{13}\text{C}$ .  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  variability is not synchronous with atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  variability. For a common time period, only ~50% of the atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  variability is represented in the  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$ ;  $\delta^{13}\text{C}_{\text{Fiji-DIC}}$  response to atmospheric  $\delta^{13}\text{C}_{\text{CO}_2}$  depletion is delayed by ~10 years. These relationships provide a means to study the dynamics of  $\text{CO}_2$  uptake at a given site (Quay et al., 1992; Bacastow et al., 1996; Heimann and Maier-Reimer, 1996).

In the context of increasing anthropogenic  $\text{CO}_2$  emissions, the rate of oceanic  $\text{CO}_2$  uptake might be strongly enhanced, especially in certain regions of the oceans. The lack of direct oceanic  $\delta^{13}\text{C}_{\text{DIC}}$  measurements raises the need for further understanding of past surface water  $\delta^{13}\text{C}_{\text{DIC}}$  variability. Reconstructions such as the coral skeletal  $\delta^{13}\text{C}$  data presented in this report may serve as invaluable resources for determining more accurate budget estimates of oceanic  $\text{CO}_2$  uptake.

## Acknowledgments

For coral core FVB1 and FVB2 collection, we thank the following persons and organizations: Saimone Tuilaulala (Director of Fisheries) and Aisake Batibasaga (Principal Research Officer) of the Government of Fiji, Ministry of Fisheries and Forests, for supporting this research program; Jennifer Caselle, David Mucciarone, Tom Potts, Stephan Bagnato, Ove Hoegh-Guldberg, and the J. M. Cousteau Resort in Savusavu (Fiji) for the assistance in the field sampling. For coral core 16F collection, we thank the following persons and organizations: Guy Cabioch, leader of the Paleofiji cruise. The officers and crew of IRD R/V Alis: Guy Cabioch, Yvan Joël Orepüller and Stéphanie Reynaud-Vaganay for their assistance in coral drilling. We would like to acknowledge the work of Stephen S. Howe for the acquisition of the  $\delta^{13}\text{C}$  data. The National Oceanic and Atmospheric Administration's CCDD Paleoclimate Program provided major funding to Braddock K. Linsley for this research.

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